A multiproxy palaeolimnological reconstruction of the nature and timing of climatic changes in the Northern Isles from the end of the last glaciation through the early Holocene

Melanie Vanessa Kingsbury

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School of Natural Sciences
University of Stirling

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STATEMENT OF ORIGINALITY

I hereby confirm that the research contained in this thesis is original, has been completed by the author and all work contained herein has not been submitted for any other degree.

All published material has been duly acknowledged and cited.

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Melanie V. Kingsbury

Date: ………………..…….

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ABSTRACT

The Northern Isles are strongly influenced by changes in the North Atlantic Ocean-atmosphere system and, as they project northwards from the British Isles, provide an ideal geographical opportunity to study changing climatic gradients during the last glacial/ interglacial transition along with the detection of regime shifts. Three proxies, diatoms, pollen, and micro-XRF sediment chemistry, have been employed to explore the nature and timing of environmental changes within the water columns and the wider catchments of Loch of Sabiston, Orkney, and Loch of Clumlie and Loch of Grimsetter, Shetland to better understand the nature and timing of environmental change within and among the island groups. The records are constrained by radiocarbon dating, supported by tephrochronology, and the Greenland ice core chronology to enable the comparison of the records produced by this study with previous research in the North Atlantic region.

The diatom and lithological results from Loch of Sabiston suggest early deglaciation at c. 23,000 cal BP followed by gradual warming (GI-1e) punctuated by the cooling events coeval with GS-1 and GI-1b. However, the pollen record reflects a lagged response in the development from colonising cold tolerant vegetation to more temperate shrub and woodland communities. The Oracadian signal is dominated by the switching on and off of the accumulation of marl which serves as a supporting indicator of warmer conditions. The Shetland landscape appears to have been deglaciated later at c. 16,400 cal BP, but also has clear representation of GI-1e and the cooling events of GI-1b and GS-1. Both the Shetland and Orkney records record the dramatic cooling of the Younger Dryas but also suggest a two stage change from colder and drier to colder and wetter conditions before the onset of the Holocene. Shetland appears to have experienced less extreme climatic changes in comparison to Orkney despite being in the same present phytogeographical region. This is likely due to the former persisting in the arctic domain and the latter being closer to the latitudinal shifts in the warmer ocean circulation of the North Atlantic during the LGIT.

Comparison of the three proxies demonstrates that they may differ by several hundred years in their response to dramatic climatic changes and, therefore, highlights the strength of multi-proxy approaches to reconstructing Quaternary environments. Combining proxies such as diatom and μ-XRF scanning techniques will provide a greater understanding of the processes occurring during environmental change in this region.

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GLOSSARY OF ACRONYMS AND TERMS

Term	**Acronym**
$\mathbf{CO_3^{2-}}$	Carbonate ion
$\mathbf{HCO_3^-}$	anion of carbonic acid (bicarbonate)
μ-XRF	micro- X-ray fluorescence
a.s.l.	Above sea level
AMOC	Atlantic meridional overturning circulation
AMS	Accelerator Mass Spectrometry
BIIS	British and Irish Ice Sheet
BP	before present (present = 1950)
Ca+	Calcium ion
CaCO3	Calcium carbonate
CO2	Carbon dioxide
CONISS	Constrained Incremental Sums of Squares cluster analysis
EDDI	European Diatom Database
FIS	Fennoscandian Ice Sheet
FRE	Freshwater reservoir effect
GRIP	Greenland Ice Core Project
H+	Hydrogen ion
H2CO3	Carbonic acid
H2O	Water
HTM	Holocene Thermal Maximum
IACP	Intra-Allerød cold period
LGIT	last glacial-interglacial transition
LGM	Last Glacial Maximum
MS	Magnetic Susceptibility
NAC	North Atlantic Circulation
OH-	Hydroxide ion
PBO	Preboreal Oscillation
PCA	Principal Components Analysis
RMSEP	Root Mean Square Error of Prediction
SST	Sea surface temperature
TLP	Total land pollen

Chapter 1

1. Introduction

1.1. Introduction

Reconstructing past terrestrial and aquatic changes in the Scottish Northern Isles in response to ocean-atmosphere changes in the North Atlantic during the Late Glacial into the early Holocene will be the focus of this research. The Last Glacial-Interglacial Transition (LGIT) is well recorded throughout Northern Europe (Birks, 1994; Bradshaw et al., 2000; Coxon, 2016; Eldevik et al., 2014). Abrupt events such as the Younger Dryas cold period and the Bølling - Allerød Oscillation, along with less observed intra-Allerød oscillations have been recorded in lake sediments in the region (e.g. Birks and Birks 2008; Palmer et al. 2012; Brooks et al. 2012; Turner et al. 2015; Muschitiello and Wohlfarth 2015). However, the timing of these events across the region has been found to be diachronous (Lane et al., 2013; Muschitiello and Wohlfarth, 2015; Väliranta et al., 2015). Future climate model scenarios predict that the North Atlantic region may potentially see major climatic changes (Jansen et al., 2007; Moritz et al., 2002) such as an increase in temperature (~ 2.1 °C), the prediction of warmer and drier summers, rising sea levels and winter storminess which would bring an increase in precipitation (Arnell et al., 2015). It is important to determine how the terrestrial environment could respond to these changes in atmosphere and ocean currents and the Northern Isles are ideally located to undertake this type of study due to the significant maritime influence.

The islands are strongly influenced by periods of cyclic activity, such as the North Atlantic Oscillation (NAO), and this climatic variability influences both terrestrial and aquatic ecology (Ottersen et al., 2001), although the magnitude of the influence of these events is varied.

For example, the LGIT was characterised by abrupt high-magnitude climatic changes while in contrast, the Holocene is characterised by more subtle environmental changes, based on palaeoclimatic observations (Williams et al., 2011). There is also a discrepancy in the timing of the occurrence of climatic events throughout Northern Europe (Lane et al., 2013; Muschitiello and Wohlfarth, 2015). General temperature trends for northern Europe have been developed through pollen, macrofossil, chironomid and beetle records and have been compared to the high-resolution Greenland ice core records (e.g. Birks and Birks 2008; Birks et al. 2012; Palmer et al. 2012; Brooks et al. 2012; Turner et al. 2015; Muschitiello and Wohlfarth 2015). However, the trends observed are not necessarily synchronous across regions and can show temporal differences. These differences may in part be due to the low resolution of the available data (Brooks et al., 2012; Muschitiello and Wohlfarth, 2015). The timing of key features of the LGIT, such as the Bølling-Allerød interstadial, which appears to be locally determined and diachronous across the study region as well (Lane et al., 2013; Muschitiello and Wohlfarth, 2015). A better understanding of the nature and timing of ecosystem responses to landscape change may provide insight into the climatic drivers.

Palaeoecology is based on the principal of uniformitarianism, and that past environmental changes may provide information on how a region will respond in the future. The justification for studying the LGIT and the early Holocene in the North Atlantic is that present forecasts predict that the temperature fluctuations that occurred at the end of the last glaciation, which was warmer than the last two centuries, *c.* 9,000-10,000 years ago, will be similar to those that will be reached this century (Williams et al., 2013). The North Atlantic region also experienced a different rate of temperature change than the rest of Europe during this time. For example, Greenland warmed by 10 ºC in under 50 years while other parts of Europe experienced less warming (2-5 ºC) just as abrupt, over the same period (Willis and MacDonald, 2011). This illustrates the need to study the response of multiple regions since climatic changes have been observed to affect regions differently in both magnitude and duration.

The focus of this research is on the archipelagos of Shetland and Orkney, collectively known as the Northern Isles (Figure 1.1), which lie at the convergence of the North Atlantic and the North Sea. These two island groups are part of the same phytogeographical climatic region along with the Faeroe Islands to the northwest (Bunting, 1994). The archipelago of Orkney is located 16 km off the north coast of mainland Scotland and consists of approximately 90 islands, 20 of which are inhabited. The Shetland Islands are located 165 km to the northeast of the Scottish mainland and comprise over 100 islands, 13 of which are inhabited and includes the islands of Foula and Fair Isle. Any changes in the North Atlantic system will be established in the ecological record of these islands both in the terrestrial and freshwater record.

Previous reconstructions of environmental change in the Northern Isles have predominantly focused on vegetation change (palynology) (e.g. Keatinge and Dickson 1979; Hulme and Durno 1980; Birnie 1981; Bunting 1994) and stratigraphic analysis (geomorphology and lithostratigraphy) (e.g. Leinert and Keen 2000; de la Vega Leinert 2007) with some studies using macrofossils, beetles, molluscs, and organic content and even fewer studies using diatoms (e.g. Hulme and Shirriffs 1994; Bunting 1996; Birnie 2000; Leinert and Keen 2000; Whittington et al. 2003; Whittington et al. 2015). However, most of these methods overlook the sensitivity of lacustrine ecosystems, which are present in abundance on these islands, to environmental changes. By studying the changing lake ecology, more detailed information on the surrounding environment will be obtained to enable the detection of thresholds or instability in the aquatic communities and together with pollen records will allow for more robust inferences concerning the drivers of environmental change.

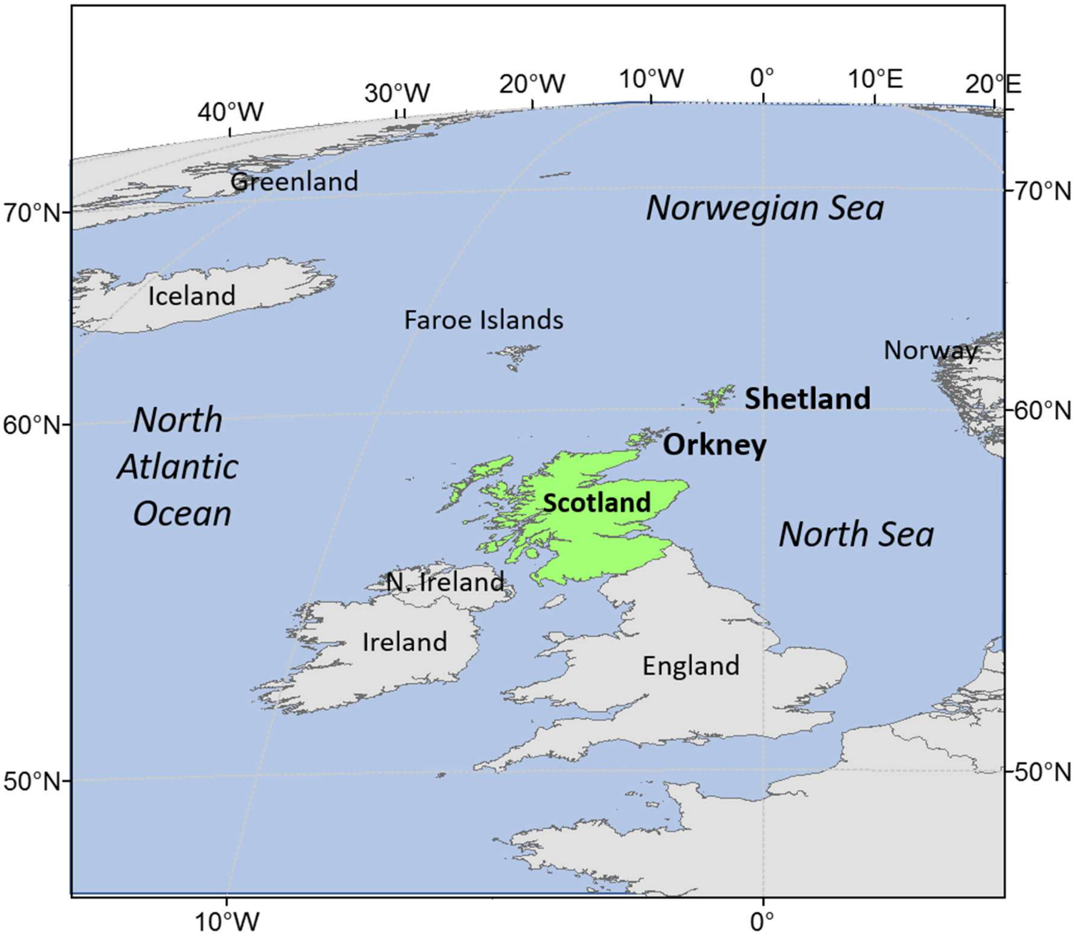

Figure 1.1 North Atlantic region showing the location of Shetland and Orkney in relation to the wider region.

1.2. Aims and Objectives

This project will apply a multi-proxy approach using biotic and abiotic proxies to reconstruct the ecological responses of lochs on Shetland and Orkney to interpret the changes in LGIT and Holocene climate. The principal technique used for this study is diatom analysis. Their short life-cycle and sensitivity to ecological changes make them ideal indicators of rapid environmental change (Last and Smol, 2001). This allows for the detection of rapid responses to variations in the record that will be in phase with the underlying causes of these changes. With other proxies such as pollen, there can be a lag in the ecological response to environmental changes due to the life cycle of terrestrial vegetation (Last and Smol, 2001). The addition of more diatom-based studies will provide high-resolution information of changes that have occurred in the freshwater environment along with detecting ecological change that the terrestrial record might not necessarily detect. The analysis of pollen along with the analysis of the physical and chemical characteristics of the lake sediments will support the diatom records produced for this study to explore the terrestrial and sedimentary history of the catchment alongside the aquatic responses. A single proxy on its own can only provide part of the narrative of the changes occurring. Therefore, a more detailed understanding of the nature of environmental change will be obtained by using a multi-proxy approach since changes occurring at different spatial and temporal resolutions can be revealed due to the individual nature of the proxies. High-resolution diatom and µ-XRF (micro- X-ray fluorescence) records will be analysed to determine significant changes in the diatom assemblage along with the geochemistry of the sites.

Recently there has been an increase in the development of methods to detect early changes due to the presently changing climate (Dakos et al., 2015; Scheffer et al., 2012; Streeter and Dugmore, 2013). Therefore, this research addresses the following research questions:

1) What is the synchroneity of multiple proxies in recording environmental change within the lake catchment and how do the results compare to the rest of Shetland and Orkney?

2) How synchronous and sensitive are the results from the study lochs with other proxies when identifying changes in the North Atlantic and other known climate/ environmental data (i.e. Greenland ice core records)?

3) Is there evidence of a ‘warning signal’ in the diatom, pollen, and geochemical records that could be a precursor to a regime shift in the environment?

The Lateglacial and early Holocene palaeoenvironmental changes will primarily be the focus of this research. However, results of the later Holocene will be introduced to put the nature and timing of climatic changes during the Lateglacial and early Holocene into context with those of the later Holocene and illustrate the probable role of early human influence in the catchment.

Despite the amount of research that has been undertaken on Shetland and Orkney regarding the terrestrial response to ecological changes, there is more to be discovered. It is recognised that even though these islands occupy a similar climatic zone, there are differences to the response both on the island groups as well as between the island groups. Determining the timing of events by using aquatic proxies like diatoms, which indirectly provides information on the environment, will aid in providing a high-resolution record of changes and will fill in some of the gaps of the ecological history of the Northern Scottish Islands.

1.3. Thesis Structure

This thesis is organized into six chapters. Chapter 2 provides a review of current understanding of environmental change in the Northern Isles and the wider North Atlantic from published literature. Chapter 3 outlines the theory and background of the methods used in this research and discusses any caveats that exist. Chapter 4 presents the results from Loch of Sabiston, Orkney along with Loch of Clumlie and Loch of Grimsetter (both from Shetland) respectively. The synthesis and discussion in Chapter 5 considers the implications of the results of this study in the wider regional context and Chapter 6 concludes the thesis.

1.4. Conventions

With palaeoenvironmental reconstructions there can be confusion when discussing different time periods and many acronyms are used, especially across different disciplines (Fig.1.2). In this thesis, the following conventions and abbreviations will be adhered to:

- Any dates that are derived from radiocarbon analysis will be presented as ^{14}C yrs. BP (Millard, 2014);
- Any dates presented that have been calibrated and/or from an age-depth model will be presented as the best fit date cal BP and rounded to the nearest 10;
- Following suggestions from Rasmussen et al. (2014) the Blytt–Sernander system of names will be used;
    - Bølling-Allerød will be used for the generally mild interstadial period from *c.* 14,600 – 12,900 cal BP;
    - Younger Dryas will be used for the stadial period between Bølling-Allerød and the Holocene *c.* 12,900 – 11,700 cal BP.

  - The Holocene will be used for the warm period beginning at the end of the Younger Dryas *c.* 11,700 cal BP instead of Preboreal/ Boreal (Fig. 1.2).

- Any dates that are presented from other research have been calibrated using Calib 7.1 and the IntCal13 calibration curve. These updated calibrated dates are presented in Appendix B.

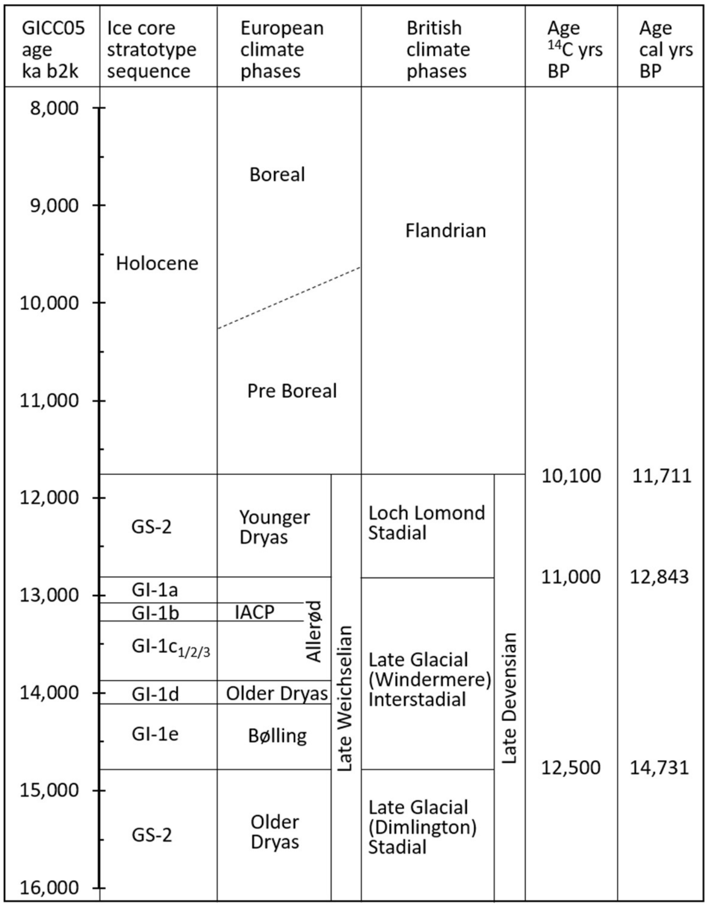

Figure 1.2 Climato-stratigraphic sub-divisions covering the Late Pleistocene-Early Holocene transition showing the various naming conventions commonly found in literature in this region. This thesis will use a combination of European and Ice-core terminology as out outlined above. NGRIP, GRIP, and GISP2 integrated chronology based on Rasmussen et al. (2014). Figure adapted from Mithen et al. (2015).

Chapter 2

2. Literature Review

2.1. Introduction

The focus of this research is on the nature and timing of events in the Northern Isles from deglaciation through to the early Holocene. This chapter will cover the mechanisms of North Atlantic Ocean-atmosphere circulation and its climate to provide background information on the major climatic influences for this region as the marine environment directly influences the terrestrial environment. A summary of findings from palaeoenvironmental sites situated in the North Atlantic will then provide a regional context to climatic changes that have occurred. The island archipelagos of Orkney and Shetland, which make up the Northern Isles, will then be discussed in more detail, including geology, soils, present land use, and population, along with their palaeoenvironmental history. This research is also focused on the timing of regional responses to changing climate and therefore the information presented will outline key findings, and highlight questions that have arisen that this research will address in this context. Finally the concept of tipping points and regime shift will be introduced along with Fisher Information to explore the nature and timing of landscape and climate changes.

2.2. The North Atlantic Region

The North Atlantic is a transitional region between the Arctic and the temperate climates of North America and Europe. This area consists of the North Atlantic Ocean and the North Sea, and includes Greenland, Iceland, Faroe Islands, Norway, the Northern Isles of Shetland and

Orkney, along with the rest of the British Isles (Fig. 2.1). The marine environment heavily influences the landmasses in this region and, therefore, terrestrial changes reflect the climatic changes that have occurred. The North Atlantic Current (NAC) is one of the main drivers of climate in this region as it is the North Atlantic arm of the Atlantic meridional overturning circulation (AMOC) which transports heat around the globe (Thirumalai and Richey, 2016). The North Atlantic Current splits into two in the North Atlantic off the west coast of Ireland (Fig. 2.2), with one half, the Norwegian Atlantic Current, passing through the Faroe-Shetland channel and the other half towards Greenland. This transport of heat from the southern hemisphere into northern latitudes stops sea ice from extending and a temperate climate to be observed in the region compared to other regions at similar latitudes (Rahmstorf, 2007).

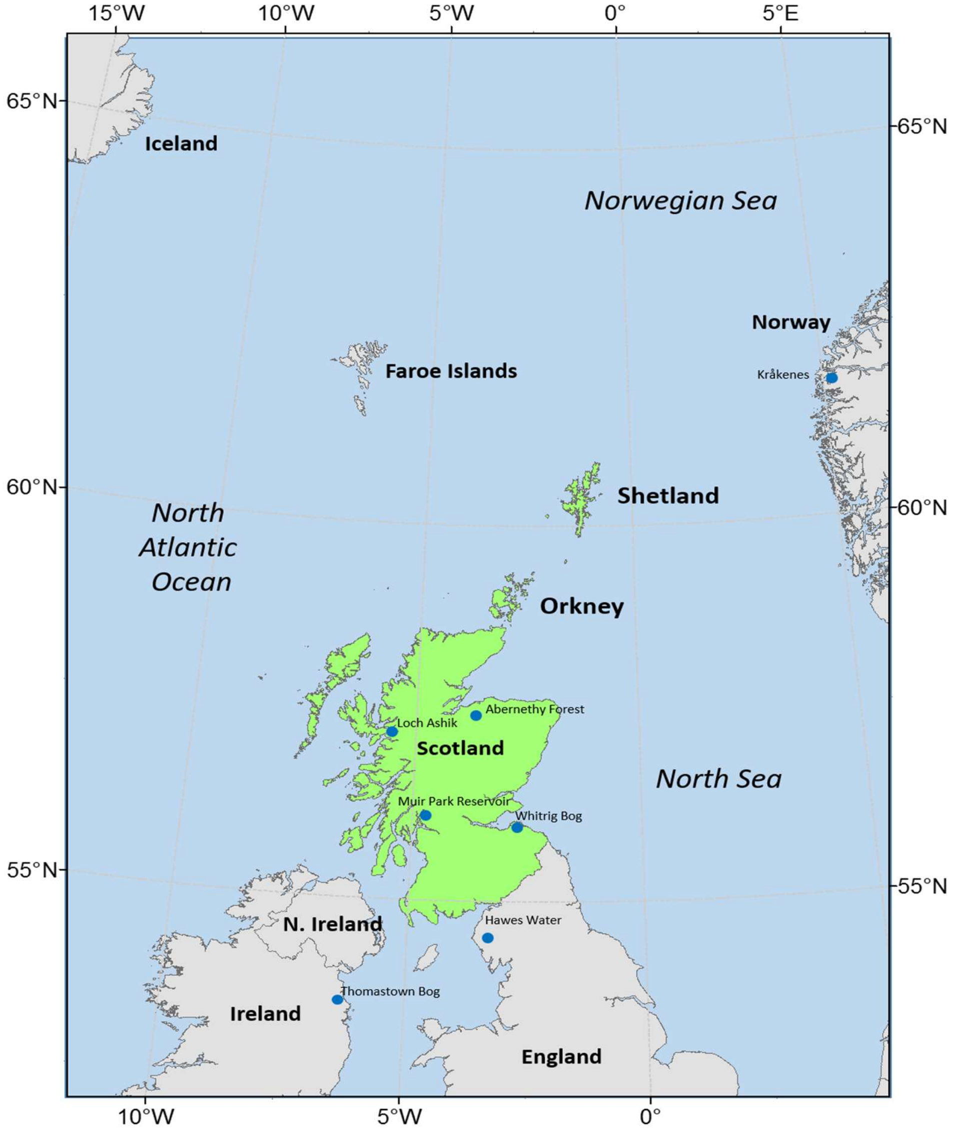

Figure 2.1 Regional map of the North Atlantic showing the main study areas of Shetland and Orkney along with sites mentioned in the text (blue circles).

The NAC is influenced by the hydrological cycle and the heating of sea surface waters (Clark et al., 2002). In the past the NAC appears not to have been as stable as presently and has undergone major changes in the Quaternary, with three different modes observed in the palaeoenvironmental record (Clark et al., 2002). These modes were defined as: 1) the modern mode which sees deep water forming in the Nordic Seas; 2) the glacial mode where deep water forms in the subpolar North Atlantic; and 3) the Heinrich mode where the NAC is switched off due to freshwater inputs from meltwater floods or glacial disintegration (Clark et al., 2002; Rahmstorf, 2007). This third mode was the probable trigger for the Younger Dryas and the 8.2 ka cold events (Clark et al., 2002; Rahmstorf, 2007). These changes in ocean circulation directly affect surface climate and, therefore, plays a vital role in modulating the climate of North Atlantic landmasses.

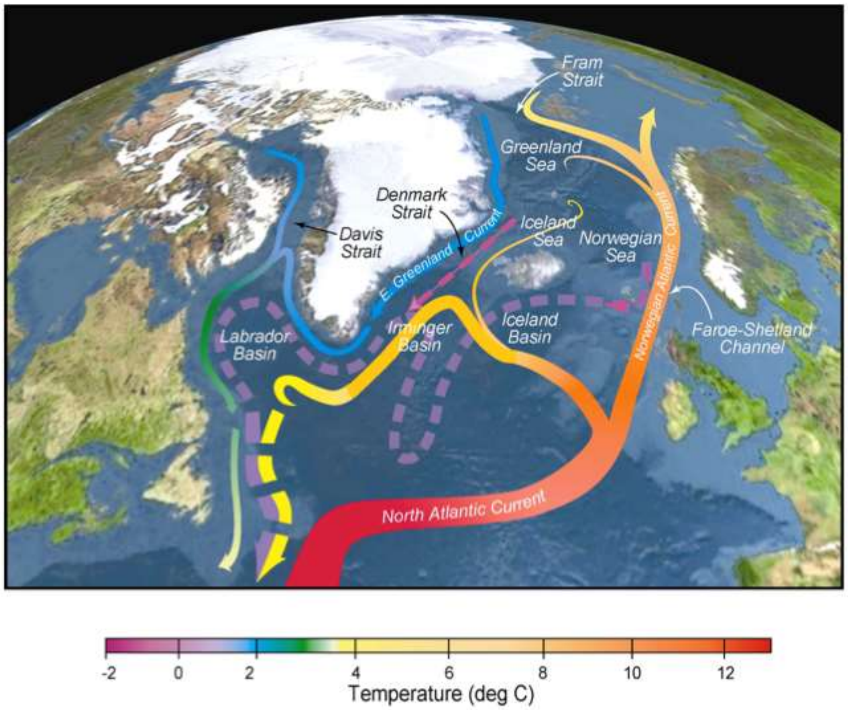

Figure 2.2 The Nordic Seas showing the circulation of surface currents (solid curves) and deep currents (dashed curves) that form the AMOC (from Curry and Mauritzen (2005)).

Solar irradiance, which works on longer-time scales, is another important climatic influence during the end of the last glaciation and the early Holocene. Solar insolation is the amount of solar radiation on the surface of the earth and directly influences sea surface temperatures (SST). Changes in solar irradiance through time have caused periods of instability and influenced warmer and/ or colder events evidenced in the palaeoecological record (Rahmstorf, 2007). There is a complex relationship between solar insolation, the NAC which impacts sea ice, and SST. The interaction of these, therefore, influences the climate of the North Atlantic and its effects on the terrestrial environment. The Holocene thermal optimum from *c.* 9400 BP to *c.* 8000 BP is thought to have been caused by high summer insolation (Koç et al., 1993). Around 6500 cal BP, summer insolation began to decrease and continued to do so through the Holocene (Birks and Koç, 2002; Solignac et al., 2004).

2.3. Regional Palaeoenvironmental Synthesis

The palaeoenvironmental history of the North Atlantic has been well studied using marine sediment cores and the glacial history explored through glacial geomorphology (Bickerdike et al., 2016; Hall, 2013; Lambeck, 1995; Walker et al., 1994). More recently offshore multibeam echo sounding studies have complemented the onshore geomorphic record (Bradwell et al., 2008; Bradwell and Stoker, 2015; Hall, 2013). Lake and peat bog records have been used to explore in detail the terrestrial palaeoclimate, often employing multiple proxies such as chironomids, pollen, ^{18}O isotopes, organic content of sediment, diatoms, and biogenic silica which extend from the end of the last glaciation through the Bølling-Allerød interstadial, Younger Dryas stadial and the early Holocene (e.g. Jones et al. 1989; Bradshaw et al. 2000; Brooks and Birks 2000; Marshall et al. 2002; Andersen et al. 2004; Bedford et al. 2004; Kaplan and Wolfe 2006; Andresen et al. 2007; Birks and Birks 2008; Palmer et al. 2012; Massa et al. 2012; Brooks et al. 2012; Turner et al. 2015; Muschitiello and Wohlfarth 2015; Brooks et al. 2016) to inform the Greenland ice core records. The following sections provide an overview of the general trends of North Atlantic palaeoenvironmental change from the end of the last glaciation to the early Holocene. The most detailed results come from chironomid records throughout the North Atlantic and will be used to illustrate these changes along with other proxies. The dates in the section headings are from the Greenland ice core records which are made up of cores GRIP, NGRIP, and GISP2 hereafter called Greenland ice cores (Rasmussen et al., 2014) are used for reference.

2.3.1. Deglaciation & the Bølling – Allerød (>c. 14,700 – c. 12,900 cal BP)

The last glacial interglacial transition (LGIT) is divided into the following events: deglaciation (LGM), the interstadial period of the Bølling - Allerød, the Younger Dryas Stadial, and the early Holocene. The onset of the termination of the last glacial maximum (LGM) was characterized by the melting of the ice sheets and the subsequent release of fresh water into the northern oceans shutting down the AMOC causing low surface salinity and cold SSTs (Eldevik et al., 2014). The British-Irish ice sheet (BIIS) had retreated and most of Britain was uncovered by *c.* 16,000 years ago (Bradwell and Stoker, 2015; Kelly et al., 2017). On land, this change in the oceanic environment translated into arctic conditions characterised by permafrost, terrestrial instability/ erosion, and barren mineral soils. As vegetation became established a Lateglacial flora developed with heliophilous plant communities that indicated an open areas, disturbed ground, and cold environment which sometimes included *Artemisia* (Tipping, 1991).

At *c.* 14,700 cal BP, warming increased along with the return of the AMOC and increasing SST in the North Atlantic region which signalled the commencement of the Bølling–Allerød interstadial (Bromley et al., 2014; Ebbesen and Hald, 2004; Koç et al., 1993). The progression of climatic changes which occurred during the Bølling–Allerød were complex, with several cold periods occurring during the interstadial as evidenced in the Greenland ice cores (Rasmussen et al., 2014) and are outlined in Figure 2.3. The Abernethy Forest record detected three centennial scale cold periods that relate to Greenland ice core events GI-1d and GI-1c (Rasmussen et al., 2014) and the Intra Allerød Cold Period (IACP) GI-1b. At Muir Park Reservoir (Brooks et al., 2016), located at the southern limit of the Loch Lomond basin, GI-1d and the GI-1b along with the warming of the Allerød was detected in the chironomid record, while Whitrig Bog (Brooks and Birks, 2000) and Loch Ashik only detected GI-1d (Brooks et al., 2016, 2012) due to their various locations across Scotland.

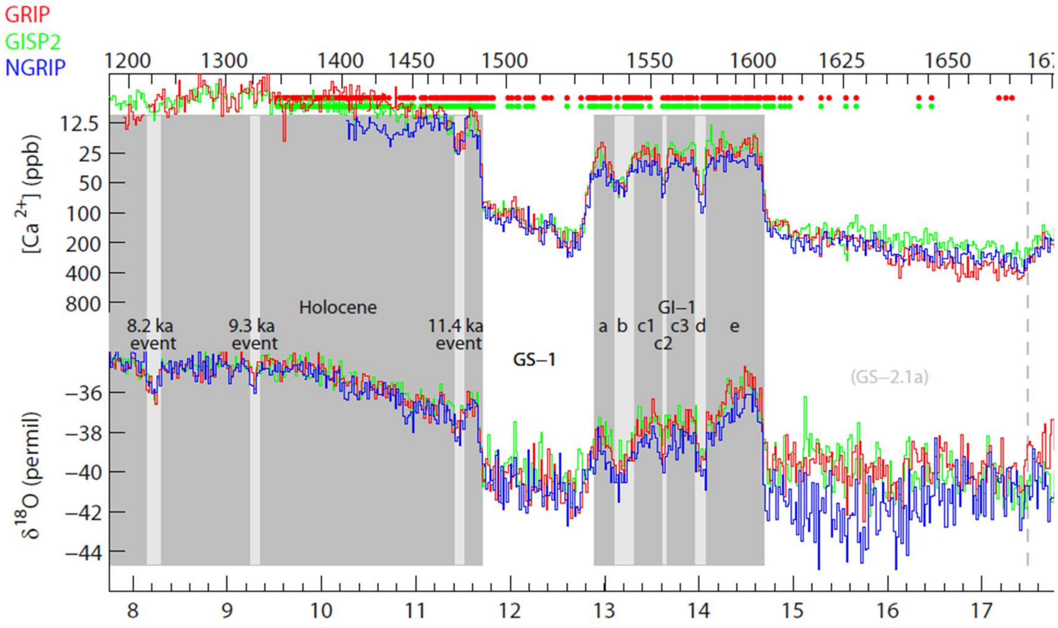

Figure 2.3 Greenland ice core terminology as it relates to the Bølling-Allerød interstadial illustrating multiple cold periods. From right to left the equivalent old terminology along with the ice core designations: GS-2.1a = Older Dryas; GI-1e = Bølling; GI-1d = Older Dryas; GI-1c3 = Allerød; GI-1c2 = cold period; GI-1c1 = Allerød; GI 1b = Intra Allerød Cold Period; GI-1a = Allerød; GS-1 = Younger Dryas (modified from Rasmussen et al. (2014)).

Chironomids were used along with micro-XRF (μ-XRF), pollen, sediment and isotopes to reconstruct a high-resolution multi-proxy record from Thomastown Bog located on the east coast of Ireland that distinguished the Bølling-Allerød interstadial along with an Intra Allerød Cold Period and the transition from the Younger Dryas into the Holocene (Turner et al., 2015). A comparison of several of these records with the $\delta^{18}O$ NGRIP Greenland Ice Core record is illustrated in Fig. 2.4. This evidence illustrates the individual regional responses to the wide spread warming event of the Bølling–Allerød with the colder periods more evident in some areas than others. For example, Abernathy Forest demonstrates greater variation in temperature during GI-1a and GI-1b compared to Loch Asik (Fig. 2.4).

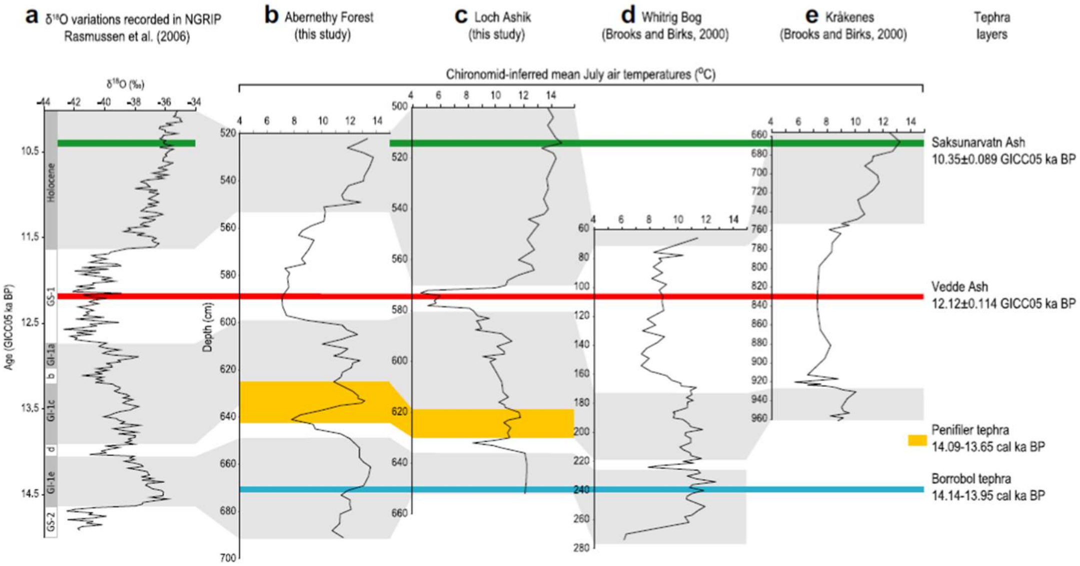

Figure 2.4 Chironomid-inferred mean July air temperature from sites located in Scotland and Norway (from Brooks et al. (2012)) compared with NGRIP oxygen isotope data from Greenland (Rasmussen et al., 2006).

2.3.2. Younger Dryas Stadial (c. 12,900 – 11,700 cal BP)

In the North Atlantic, the abrupt return to colder, glacial conditions with extensive sea-ice cover and stable cold wet winters in Europe characterised the first part of the Younger Dryas with the second half exhibiting a less cold and dry, unstable climate (Bakke et al., 2009; Baldini et al., 2015; Bradley and England, 2008; Coope et al., 1998; Ebbesen and Hald, 2004; Eldevik et al., 2014; Isarin et al., 1998). This return to glacial conditions triggered a re-advance of the Scottish ice sheet (Benn et al., 1992; Bickerdike et al., 2016; Golledge, 2010; Hall, 2013) and is thought to be caused by an influx of fresh water from the collapse of North American ice dammed pro-glacial lakes (Broecker et al., 1989). Chironomids were used to show variations in the way each location responded to changes in the Younger Dryas indicating regional responses to this climate change based on sites in Scotland and Northern England (Fig. 2.1) that have a Lateglacial record (Loch Ashik, Abernethy Forest, Hawes Water and Muir Park Reservoir (Bedford et al., 2004; Brooks et al., 2016, 2012)). The results demonstrate that there is a definite division of the Younger Dryas into two distinct periods (Fig. 2.5). At Loch Ashik, the second half of the Younger Dryas becomes colder, whereas at Abernethy Forest there is evidence of warming which is similar to the results at Glen Roy which also showed warming at the end of the Younger Dryas (Palmer et al., 2012). However, the later cooling recorded at Loch Ashik may have been due to its proximity to the Skye Ice Field (Brooks et al., 2012) (Fig. 2.5). The differences observed between the first half and second half of the Younger Dryas is thought to have been due to changes in seasonality, sea-ice conditions, and atmospheric forcing (Brooks et al., 2016; Eldevik et al., 2014).

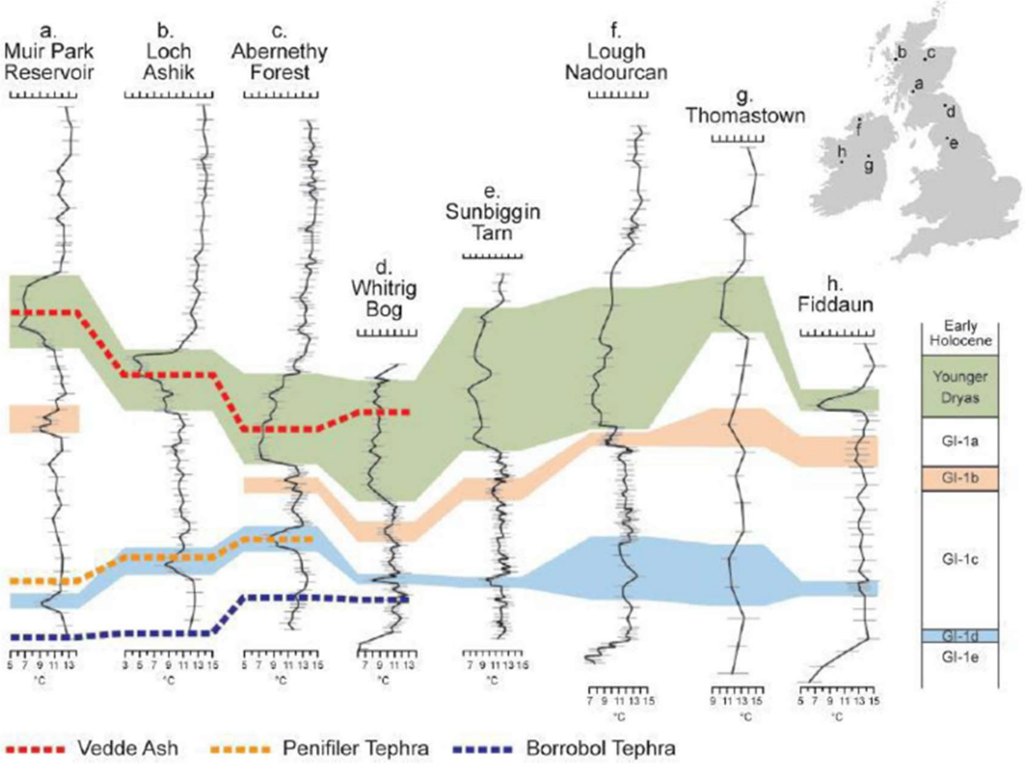

Figure 2.5 Chironomid-inferred mean July air temperature from sites from Britain and Ireland demonstrating the 2 phases of the Younger Dryas (from Brooks et al. (2016)).

2.3.3. Early to mid-Holocene (c. 11,700 cal BP – c. 6000 cal BP)

Increased warming defined the end of the Younger Dryas and the beginning Holocene *c.* 11,700 cal BP (Birks and Birks, 2008; Brooks et al., 2012; Coope et al., 1998; Jennings et al., 2014; Nesje and Dahl, 1993; Rasmussen et al., 2014). The transition to the Holocene was a dynamic period and was influenced by causes such as melting ice sheets, the establishment of atmospheric and marine currents, along with local environmental thresholds (Andresen et al., 2006). It is now understood that this warming was probably more complicated and was not spatially synchronous across regions (Andresen et al., 2006). Warming tended to be step-wise and punctuated by cold(er) events and these events have been recorded in sea surface temperature (SST) reconstructions, pollen, diatom and chironomid records, along with the Greenland ice cores $\delta^{18}O$ archives (Wanner et al., 2011). Records of reconstructed SST temperatures from the North Atlantic indicate a rapid increase of temperature during the beginning of the Holocene which most likely reflects changes in ocean circulation bringing warmer waters northwards, coupled with higher solar insolation (Eldevik et al., 2014). The Holocene Thermal Maximum (HTM) in the North Atlantic began at *c.* 9000 to *c.* 7000 cal BP and lasted to *c.* 5000 cal BP. However, macrofossils from aquatic plants in Finland (Väliranta et al., 2015) suggest that the onset of the HTM may have occurred between *c.* 11,500 and 8,500 cal BP. This conclusion is also supported by SSTs reconstructed in the North Atlantic (Eldevik et al., 2014). This discrepancy is due to the fact that aquatic plants are more sensitive to environmental change owing to their quicker dispersal rates compared to that of trees (Väliranta et al., 2015).

The beginning of the Holocene in the North Atlantic is punctuated by several warm and cold oscillations which are evidenced in the Greenland ice core records (Rasmussen et al., 2014). Two of these events, the preboreal oscillation (PBO) and the 8.2 ka event, stand out in various records across the North Atlantic. The PBO was a gradual cooling that began *c.* 11,400-11,520 cal BP (Rasmussen et al., 2014) and lasted for 100-150 years. By *c.* 11,320 cal BP, warming had resumed. The changes brought on by this cooling were variable across the North Atlantic. It was surmised that regions where warm-climate vegetation (such as *Betula* and *Filipendula* (Björck et al., 1997)) established itself were affected more than regions where cooler-climate vegetation was still present (such as grasses and dwarf shrubs) and therefore more able to tolerate the sudden reversion back to a colder climate (Björck et al., 1997). The PBO in Europe was determined to be cooler and more wet while in the northern regions like Norway and Iceland, it caused colder and drier conditions (Alley et al., 1993; Björck et al., 1997; Kelly et al., 2017; Rasmussen et al., 2007; Whittington et al., 1996).

The 8.2 ka event which again saw a return of colder and drier conditions across the North Atlantic and was a result of fresh water input into the North Atlantic (Alley and Ágústsdóttir, 2005). During the 8.2 ka event in Greenland, the climate cooled and became drier evidenced by a decrease in snow accumulation and became stormier indicated by an increase in dust and sea salt spray (Alley et al., 1997). In Sweden and Norway there was a 1 °C decrease in July inferred temperatures and in the United Kingdom, there was an increase in seasonal differences with more severe winters (Alley and Ágústsdóttir, 2005).

2.3.4. Complications with dating the Late-glacial

Defining the chronology of a site is a crucial step in determining the timing of events in the palaeoecological record along with producing age-depth models for the record. However, developing age-depth models is mired with uncertainties during the process. These include uncertainties in the material that is to be dated and the lack of dateable material, the nature of sediments (i.e. carbonate content) and the fact that not every level can be dated as well as analytical uncertainties that arise with the dating process itself (Parnell et al., 2011). Budget constraints can also limit the strength of the model based on the number of dates that are able to be obtained (Blaauw et al., 2007). The dates themselves have their own uncertainties due to the calculations and the resulting calibrated date can be multimodal with no true best point estimate while statistical applications are used to estimate these uncertainties in the calibrated dates in order to develop a working age-depth model (Blaauw et al., 2007).

The global carbon cycle and the rate of ^{14}C production has not been uniform and this will also affect radiocarbon dating by the presence of plateaus in the radiocarbon dating curve (Hajdas and Ivy, 2006). At several points in the past, ^{14}C in the atmosphere decreased at the same rate as the decay of ^{14}C, which would cause any dates during that time to be indistinguishable from one another (Trumbore et al., 2016) (e.g. during the Lateglacial and the 10-kyr ^{14}C age plateau (Hajdas and Ivy, 2006)). This issue has been mitigated by U/Th dating of corals and accelerated mass spectrometry dating (AMS) ^{14}C dating of terrestrial macrofossils in annually laminated sediment (Hajdas and Ivy, 2006) providing ‘fill in dates’ to make the calibration more robust.

When comparing palaeoenvironmental records from across a large region there is the tendency to align these records and ignore the dating uncertainty that is inherent in each record. This can result in a ‘sucking in’ of separate events or ‘smearing’ out in time of single events (Baillie, 1991; Blaauw et al., 2006), with the possibly of concluding synchroneity when there is none. More studies with multiple proxies, at a higher resolution, and with more robust dating are being undertaken and are starting to untangle asynchronies across records and it is emerging that there is a time transgressive occurrence of climate change in this region.

In the North Atlantic, the tendency is to compare records to the Greenland ice core records, however it has now been demonstrated that the response to the abrupt climatic change occurring at the beginning of the Younger Dryas was time transgressive across Northern Europe (Lane et al., 2013; Muschitiello and Wohlfarth, 2015). The time transgressive nature of environmental changes during the LGIT across Northern Europe has been demonstrated using the Vedde ash layer which highlights unambiguously the temporal differences between sites (Lane et al., 2013) by providing a benchmark to correlate temporal differences in response to climate. Four sites across Northern Europe that differed in latitude were analysed by standardizing the published dates for the Allerød – Younger Dryas transition using IntCal13 (Lane et al., 2013; Muschitiello and Wohlfarth, 2015). At each site, the transition to the Younger Dryas occurred a few hundred years before and after the established GRIP date. Vegetation changes reflecting the conditions of the Younger Dryas at the two lower latitude sites occurred earlier than the vegetation changes at the more northern sites. This suggests that there is a need for more independently dated records and tuning records to the ice cores is likely to be erroneous. The development of tephrachronologies for the North Atlantic region is mitigating the issues outlined above by providing the ability to correlate palaeoenvironmental records by using known and well dated tephra layers as anchor points. This then allows for the detection of synchroneity of records to be established such as the Vedde ash layer (Lowe, 2015; Lowe and Walker, 2015; Timms et al., 2016).

The alignment of changes of different proxies across regions also needs to be considered. The use of proxies from different environments (aquatic, terrestrial, ice core) with different life histories and environmental sensitivities can also lead to conflicting results, especially when trying to determine synchronous and asynchronous events. At Kråkenes, Norway (Fig 2.2) a well-dated record of pollen, diatoms and macrofossils indicates that the vegetation changed at a different rate than the diatoms though the end of the Lateglacial and during the first 600 years of the Holocene (Birks and Birks, 2008; Bradshaw et al., 2000). The changes observed in the diatoms were more gradual and asynchronous in the Holocene than the pollen record. The diatoms did not change until stabilization of the catchment and an increase in organics occurred which is synchronous with the changes observed in the lithology of the core and not the pollen (Bradshaw et al., 2000). Bayesian analysis of many age-depth models can assist in determining the timing of events across regions by providing a probability of the events occurring at a given time (Blaauw et al., 2007) to help avoid the problems discussed by Baillie (1991).

2.3.5.Summary

The nature and timing of palaeoenvironmental changes that have occurred in the North Atlantic Region since the end of the LGM through the early Holocene have been well documented using a variety of proxies. At millennial or centennial-scale there is a higher probability that climatic changes were synchronous, but at sub-centennial or sub-decadal the changes may be time-transgressive (Blaauw et al., 2007). We need high-resolution and well dated records to be more confident in the timing of events that are occurring. Due to the complex spatial nature of climate change, proxies can be affected by other factors than just a changing climate, and therefore there can be instances where synchroneity could be caused by age model errors, mistakes in interpretation or wishful matching of regional records (Blaauw et al., 2007).

2.4. Orkney

2.4.1. Present Orkney

The Orkney Islands are located 16 km north of mainland Scotland off the coast of Caithness (Fig 2.6). There are around 90 islands, most of which are very small and uninhabited. According to the census in 2011, the total population of Orkney was 21,349 spread over 20 islands (National Records of Scotland, 2012) which is just a fraction of what it had been in the past (Davidson and Jones, 1985). The present climate is mediated by the sea which provides intense winds and mild temperatures that fluctuate from a low of 1.9 °C in the winter to average highs of 16.0 °C in the summer. Average yearly rainfall is 1039 mm and average yearly wind speed is 13.5 knots (1981-2010 records from Kirkwall http://www.metoffice.gov.uk, accessed October 12, 2016).

The topography of Orkney is made up of low rounded hills smoothed down by glaciers (except for the island of Hoy, Fig. 2.7) and is composed mostly of Devonian Middle Old Red sandstones, mudstone and siltstone with glacial till (EDINA, 2016) (Fig. 2.8). The predominant soil cover on Orkney is peat (33.8%) which includes blanket peat, peaty gleys and podzols, followed by noncalcaerous soils (30.8%) and brown forest soils (24.2%) (Fig. 2.9). Open fresh water only makes up 3-4% of the surface area of Orkney (EDINA, 2007) (Fig 2.9). Improved grassland dominates the land cover (35.7%), along with low productivity grassland (27.4%) and dwarf shrub heath (13.1%) (EDINA, 2007).

Agriculture and resource extraction have transformed the overall surface of Orkney through the application of fertilizers, drainage techniques, peat cutting for fuel, and ploughing since being settled. Agriculture accounts for over 60 % of present land use on Orkney which includes beef and dairy cattle, sheep farming and some crops such as barley, potatoes, oilseed rape and oats (Land Use Consultants, 1998). The remainder of the land includes three main vegetation categories: 1) upland communities above 100 m above sea level (a.s.l.) which is too poor for agriculture; 2) coastal plant communities and 3) grass heath with tall herb and wetland communities amongst agricultural land (Davidson and Jones, 1985). There is no native woodland on Orkney with the exception of Hoy (Keatinge and Dickson, 1979) though there are several plantations made up of deciduous trees (Davidson and Jones, 1985).

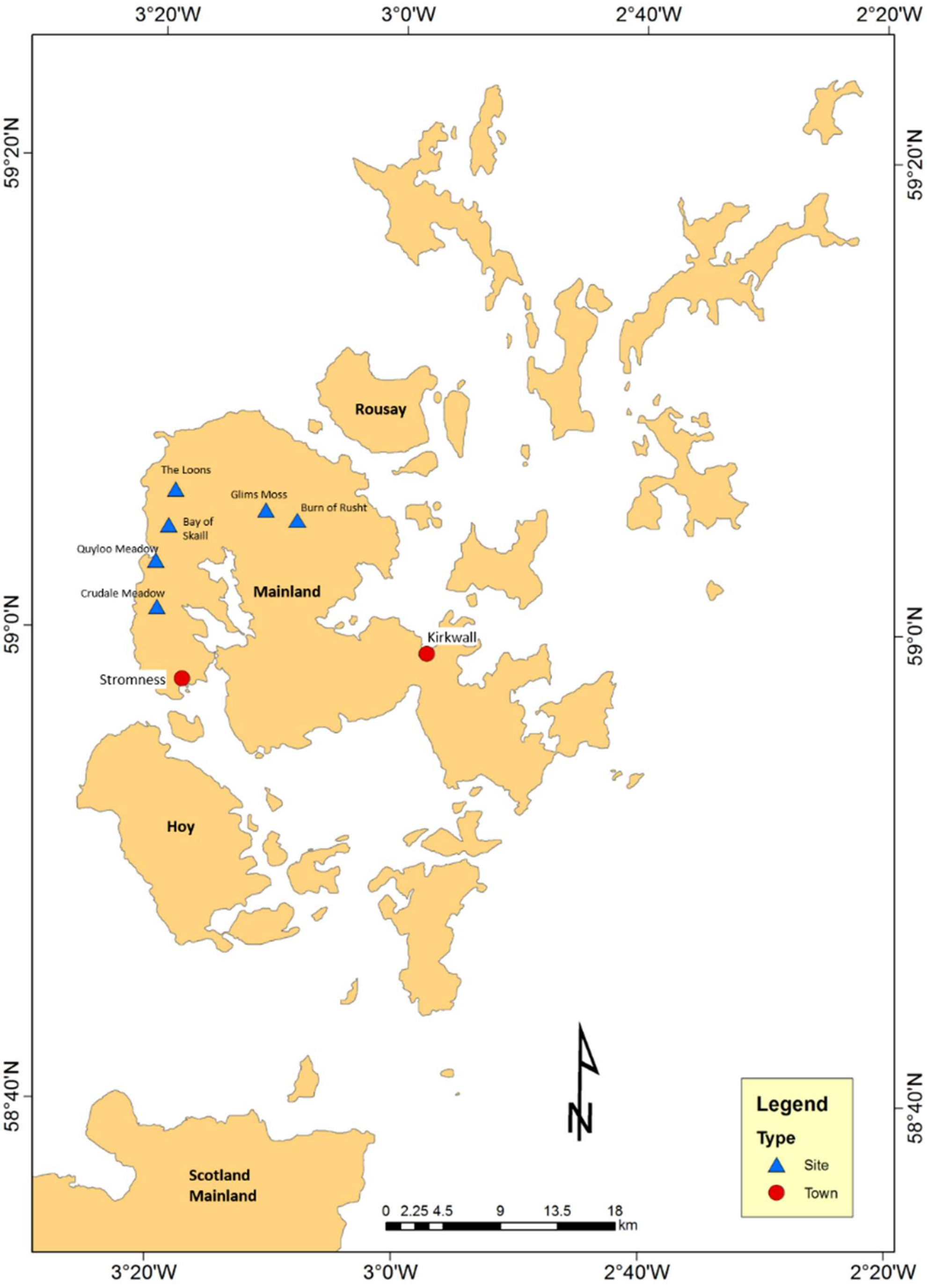

Figure 2.6 The island archipelago of Orkney showing towns (red dots) and locations that are mentioned in the text (blue triangles).

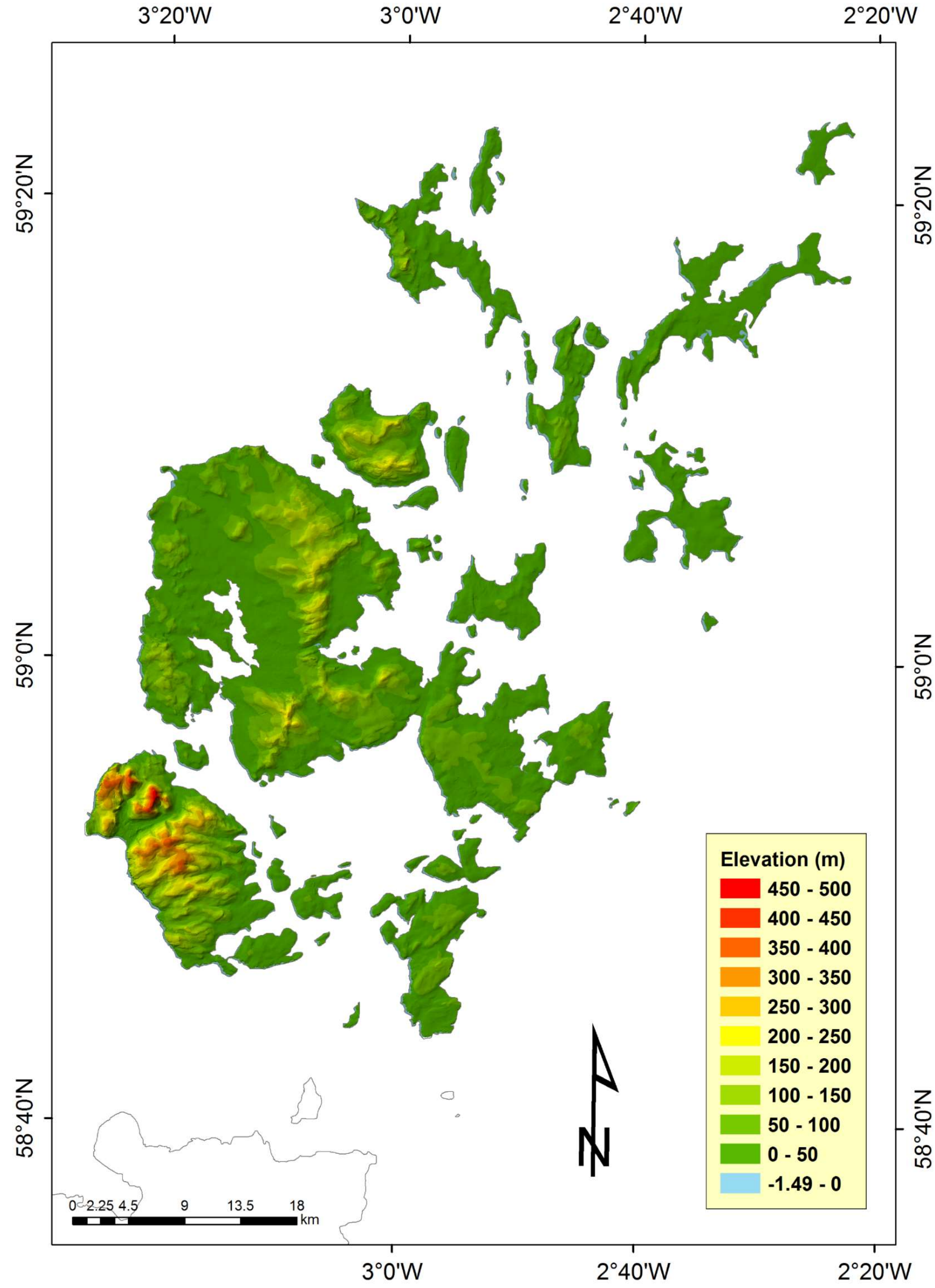

Figure 2.7 The topography of Orkney illustrating the gentle relief of the islands except for the island of Hoy (bottom left) (OS Terrain 5, Scale 1:10000).

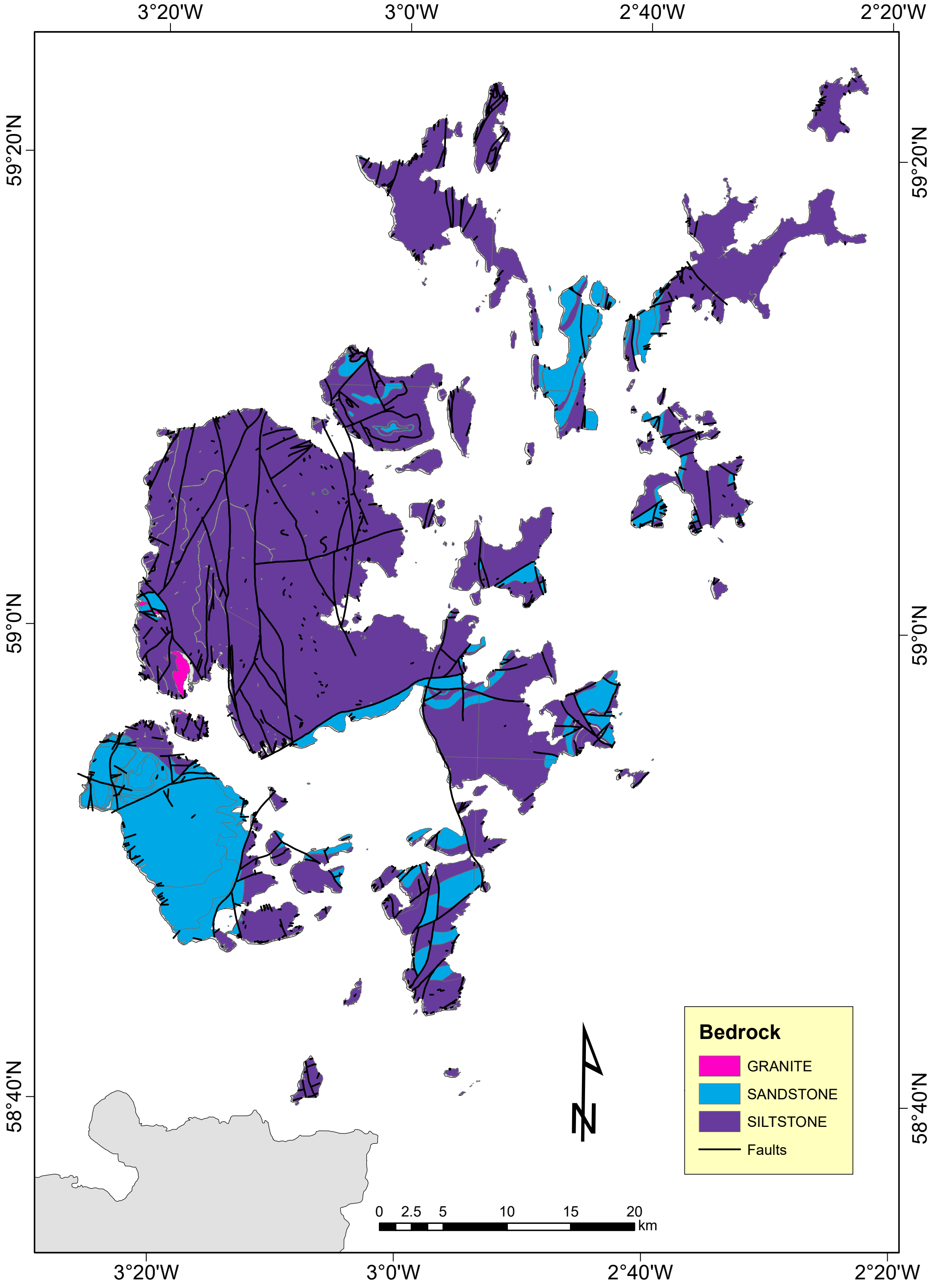

Figure 2.8 The bedrock configuration of the Orkney Islands (EDINA 2016). Most of the land consists of sandstone and siltstone (Old Red Sandstone). There are numerous faults also present on the islands.

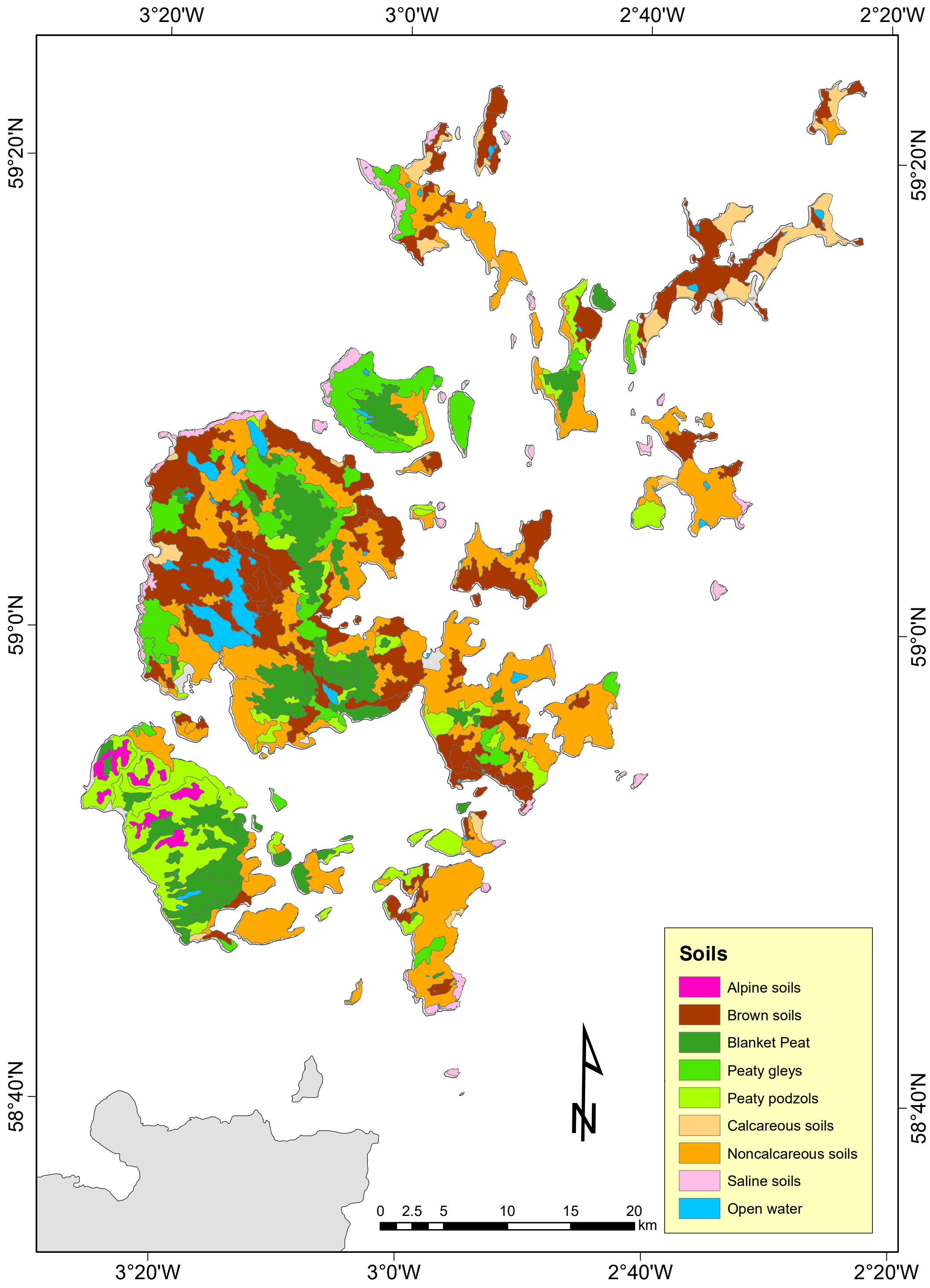

Figure 2.9 The soil profile for the Orkney Islands (EDINA 2007). The calcareous soils on Mainland tend to have marl lakes located in the region. The peaty areas tend to be restricted to the higher elevations.

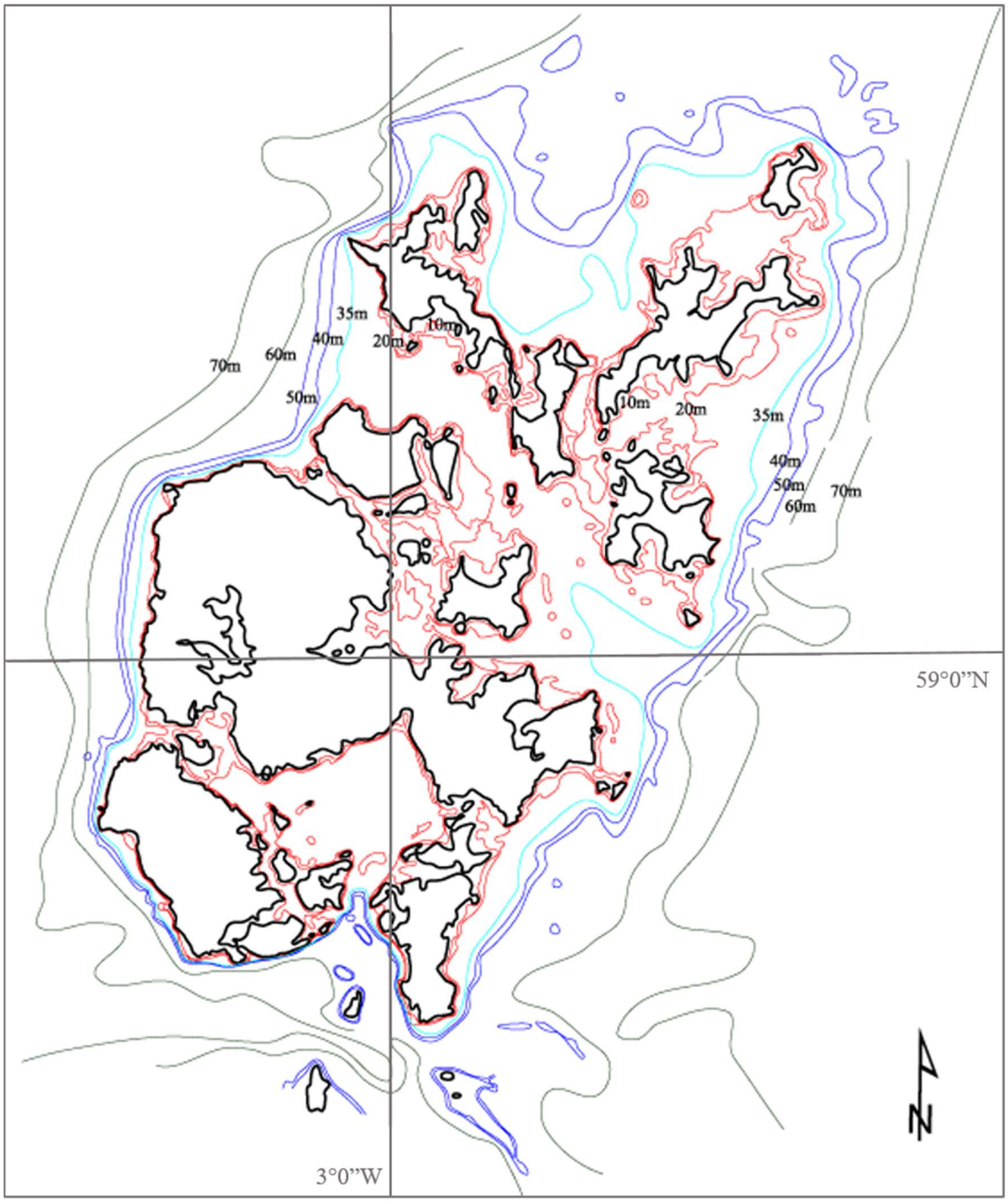

Figure 2.10 Bathymetric map of Orkney. Since the LGIT *c.* 10,000 BP the sea was around 20 m below present indicated by the red contour line. By *c.* 8,000 BP. Sea level was ~ 8 m below present day and by *c.* 7,000 BP ~ 4 m below. By *c.* 5,500 BP. sea levels reached present day levels (http://www.landforms.eu/orkney/bathymetry.htm).

The coast of Orkney has changed considerably since the end of the LGIT. Orkney was attached to mainland Scotland until it was separated by rising sea levels at *c.* 13,000 BP (Davidson and Jones, 1985) and present sea levels were reached at *c.* 4000 years ago (Fig. 2.10). Early deglaciation of the Northern Isles resulted in much of the postglacial isostatic uplift occurring during the Lateglacial and early Holocene and so relative sea level (RSL) change during the latter Holocene was limited to ~ 1 to 2 m (Smith et al., 2012).

2.4.2. The Freshwater Aquatic Environment

The low-lying relief of the Orkney islands resulted in many small and very shallow freshwater lochs except for Loch of Stenness and Loch of Harray which are the two largest lochs on the archipelago. The small lochs tend to have moderate to high natural nutrient levels and have a strong maritime influence. Due the unique position of Orkney in the North Atlantic off the coast of mainland Scotland, the lochs on Orkney exhibit a range of types that are considered to have a high natural heritage value (Scottish Natural Heritage, 2001). For example, on Hoy there are acidic-oligotrophic lochs, while a few of the northern islands of Orkney have lochs that are similar to the machair lochs of the Western Isles with high calcium content and moderate nutrients (Scottish Natural Heritage, 2001). Presently, eutrophication is the biggest pressure on the lochs of Orkney along with agriculture and water regulation (Scottish Natural Heritage, 2001).

One particular aspect of the freshwater lochs on Orkney is that many are calcareous or were marl lakes at some point in their development. Loch of Sabiston, which is part of this study, is one such lake. Marl lakes in the British Isles tend to form where the geology is carbonate based, such as Orkney's Old Red Sandstone (Pentecost, 2009). Marl lakes have a high pH (average pH of 8 with some lakes as high as 9 in shallow, highly photosynthetic water bodies (Pentecost, 2009)), and have a rich macrophyte and submergent flora, including *Chara,* which is associated with marl formation (Pentecost, 2009; Pentecost et al., 2006). These characteristics therefore present unique biological and chemical properties of the lochs. The vegetation found in marl lakes also provides suitable habitat for molluscs and crustaceans. Marl lakes can range from oligotrophic to eutrophic classifications; however, oligotrophic lakes exhibit phytoplankton that are more commonly associated with eutrophic conditions (Wiik et al., 2013). The macrophyte community in marl lakes consists of calcareous species such as *Potamogeton lucens, P. crispus*, *Elodea canadensis*, *Myriophyllum spicatum* and the macrophytic algae the Charophytes (Wiik et al., 2013). Most marl lakes are nutrient poor due to the co-precipitation of phosphorous (P) with calcium carbonate, which reduces bioavailable P and thus acts as a buffer to eutrophication (Pentecost, 2009) but only to a certain point (Wiik et al., 2013). The periphyton in marl lakes tends to be dominated by diatoms that are normally associated with soft water, nutrient poor lakes along with other species that could indicate nutrient enrichment such as *Asterionella formosa* although in low numbers (Pentecost, 2009).

Marl is precipitated out of the water column by several different processes, the most common by photosynthetic activity which occurs mostly in summer. The calcium ion originates from groundwater input, the bedrock of the catchment and sediment fluxing, while the carbonate ions come from the atmosphere, the respiration of plants, and bacterial processes (Kelts and Hsü, 1978). Photosynthetic activity removes CO_2 from the water along with increasing oxygen contect of the water. This then decreases H+ making the water more alkaline and causing CO_3^{2-}compounds (ie $CaCO_3$, $SrCO_3$) to precipitate. The following equations represent this process (Pentecost et al., 2006; Wetzel, 2001; Wiik et al., 2013):

$$Ca^{2+}{}_{(aq)} + CO_3^{2-}{}_{(aq)} \tag{1}$$

$$\updownarrow$$

$$CaCO_{3\,(s)} + H_2CO_{3(aq)} \leftrightarrow Ca^{2+}{}_{(aq)} + 2HCO_{3(aq)}^{-} \tag{2}$$

$$\updownarrow \qquad\qquad \updownarrow$$

$$CO_{2(aq)} + H_2O_{(l)} \qquad CO_3^{2-}{}_{(aq)} + H^{+}{}_{(aq)} \tag{3}$$

$$\updownarrow \qquad \updownarrow$$

$$CO_{2(g)} \quad H^{+}{}_{(aq)} + OH^{-}{}_{(aq)} \tag{4}$$

Marl precipitation in a lake is dependent on the pH, the temperature of the system, and photosynthetic activity. The warming of the loch reduces the solubility of calcite along with a decrease in the amount of CO_2 in the water which in turns can supersaturate the water. This is visualized in Fig. 2.11 which outlines the physical processes that occur to enable carbonate precipitation in lakes (2.12a) and how the relationship changes over the course of a year (2.12b).

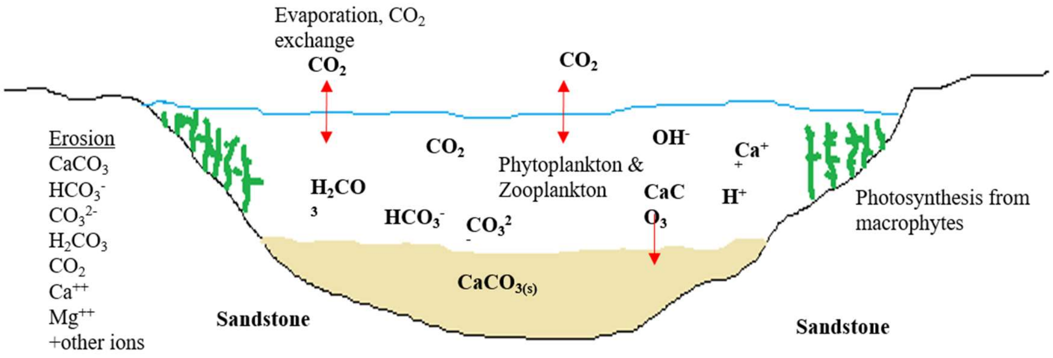

Figure 2.11 Illustration of the carbon sequence that forms the marl precipitate in lakes. For carbonate to precipitate out of the water, temperature must increase which drives out CO_2 from the water. H_2CO_3 breaks down to release the CO_2 out of the water. Now there is too little H_2CO_3 to keep the reaction in equilibrium (equations 1-4) therefore reaction (2) must more to the left to replace the lost H_2CO_3 and $CaCO_3$ precipitate out (illustration made from data from (Kelts and Hsü, 1978; Wetzel, 2001).

There is also a direct relationship between pH and temperature in the formation of marl. Based on the temperature profile in Figure 2.12a, the threshold appears to be around 10ºC for the onset of marl precipitation. However, it is unknown if this is universal as there is very little information found on this subject. When applying this information to palaeoenvironmental data, it might be a good indicator of threshold temperatures such as those that are observed at the end of the last glaciation and during the Younger Dryas where marl precipitation is switched “off”.

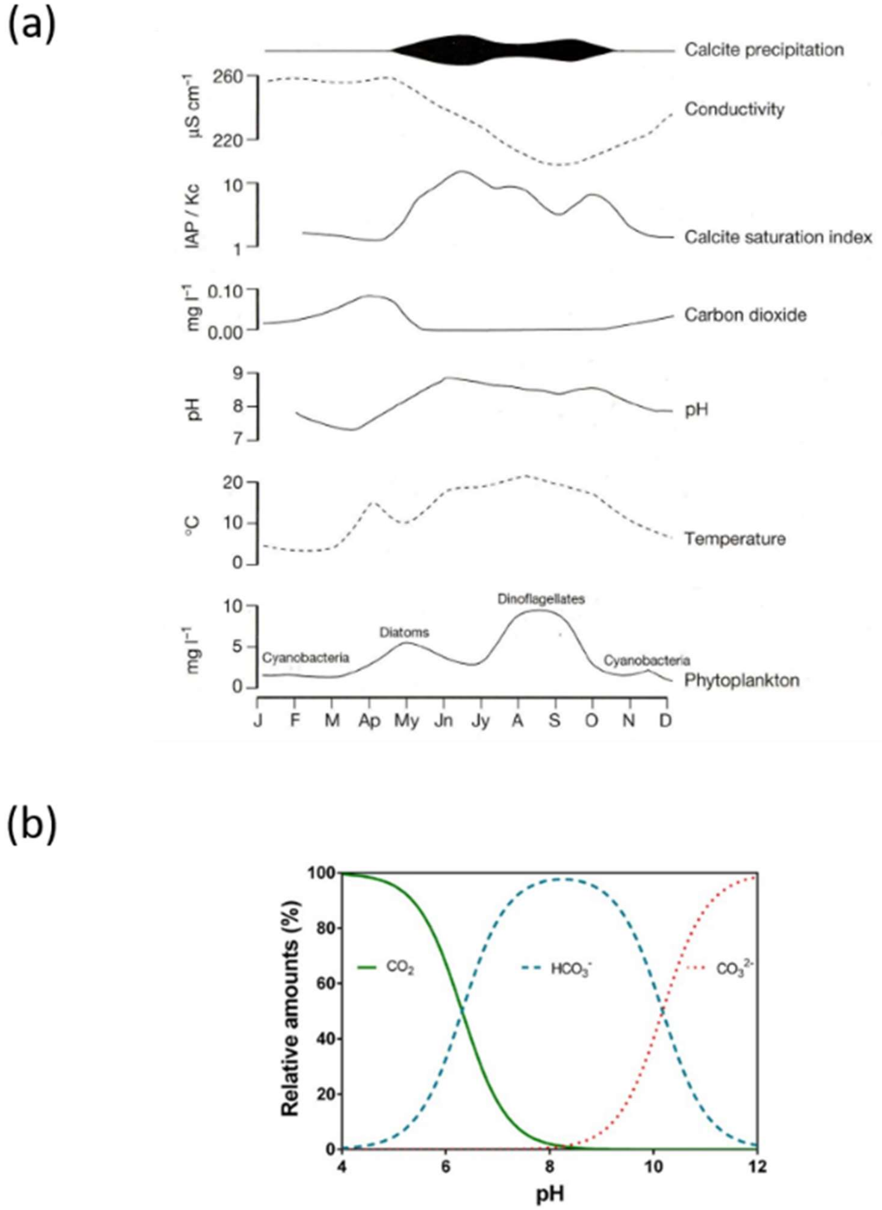

Figure 2.12 An example from Lake Zurich (1974 to 1975) illustrating the relationship between various parameters and marl precipitation (a) illustrating the state of a marl lake throughout the course of a year and the relationship between water variables from Kelts and Hsü (1978) along with (b) demonstrating the relationship between pH and the relative proportions of inorganic carbon species of CO_2 (Wetzel, 2001).

2.4.3.Previous Palaeoecological Research

Despite Orkney's rich archaeological history, environmental records of past vegetation and landscape characteristics in relation to archaeological sites tend to be highly localized and discontinuous (Farrell, 2009). Most records are limited to the Bronze Age (from *c.* 4000 years BP) (Bunting et al., 2001; Downes, 1994; Hedges, 1974, 1976; Keatinge and Dickson, 1979) and few in the Neolithic (from *c.* 6000 years BP) (Bates et al., 2016; Davidson et al., 1976; Keatinge and Dickson, 1979), and are single point reconstructions constrained to the archaeological point of interest. There are limited records that extend into the Mesolithic (Bunting, 1996; Edwards et al., 2005; Hulme and Durno, 1980; Keatinge and Dickson, 1979; Leinert and Keen, 2000; Moar, 1969; Whittington et al., 2003) and fewer that span as far as the Late Upper Palaeolithic (Birnie, 2000; Bunting, 1994; de la Vega Leinert, 2007; Hulme and Shirriffs, 1994; Whittington et al., 2003). There has been more recent published research on the palaeoenvironments of Orkney, however, most are concerned with sea levels (Teasdale et al., 2011), glaciation and isostatic uplift (Bradwell et al., 2008; Clark et al., 2012; Phillips et al., 2008), storminess (Tisdall and McCulloch, 2013), and archaeology (Bates et al., 2013; Farrell et al., 2012).

The most frequent method of reconstructing environmental changes on the Orkney islands is pollen analysis along with biostratigraphy and sediment analysis. However, these methods overlook the ecology of freshwater lake systems and how they respond to changes in climate. Proxies such as diatoms have been used rarely on Orkney and only in the context of tracking sea level changes and the occurrences of tsunamis (Bates et al., 2013; de la Vega Leinert, 2007; de la Vega Leinert et al., 2012). These studies employ the change in diatom assemblages from freshwater to marine species to track sea inundation and the timing of sea level rise in the record. Over the last ten years there is research ongoing on the Lochs of Stenness and Harray by the Rising Tides Project investigating sea level rise in these lochs along with reconstructing the palaeoenvironment in relation to the Heart of Neolithic Orkney archaeological sites (Wickham-Jones et al., 2009). However, these analyses focus specifically on marine influence and not necessarily direct climatic influences on the diatom assemblages.

2.4.4.Proxies for Reconstructing Environmental Change on Orkney

Palynology has been the primary proxy in the reconstruction of past landscapes and environments on Orkney. However, several other proxies have been used over time in conjunction with pollen. Whittington et al. (2015) used molluscs as part of a multiproxy environmental reconstruction. The mollusc data suggested that a change in mollusc species composition did not indicate a change in the environment at Crudale Meadow, but rather a change in total numbers at each sampling level (total numbers were used due to low numbers of species). This because the species present being first colonizers are ubiquitous, and have wide tolerances of water characteristics such as salinity, temperature, or substrates (O'Connor and Bunting, 2009; Whittington et al., 2015). At Quoyloo Meadow, O'Connor and Bunting (2009) also found early pioneer mollusc species such as *Lymnaea peregra*, that also have a wide range of tolerances and are quick to colonize new habitats dominate their records. Based on the assemblages, they determined that the area was covered with ephemeral, well oxygenated freshwater pools for most of the record with a change to stagnant conditions occurring prior to evidence of terrestrialization near the top of the core (O'Connor and Bunting, 2009).

Lesser used proxies on the islands include macrofossils and isotopes. The $\delta^{18}O$ isotope record from calcite at Crudale Meadow was sensitive enough to pick out major changes such as the Bølling – Allerød, the inter-Allerød cold period, the Younger Dryas, and the Pre-Boreal Oscillation at the start of the Holocene. Crudale Meadow represents the most northerly Lateglacial record from Orkney and is comparable with records from Greenland ice core along with those from the North Atlantic (Whittington et al., 2015) and appears to be the only record that has employed isotope analysis. However, the dating of the record was marred by old carbon influences and therefore the chronology used was based on Greenland records (Whittington et al., 2015).

Macrofossils compliment pollen analysis as they provide evidence for the physical presence of vegetation as the presence of pollen may be due to long distance wind transport. However, not all lake sediments contain macrofossils, so they tend not to be used as a single proxy. Macrofossils can also refine the taxonomic level since some pollen can only be identified to genus or family (Smol et al., 2001). On Orkney, terrestrial macrofossils have been used to confirm the presence of *Corylus avellana* and the absence of *Myrica* (Bunting, 1994, 1996) and as samples for radiocarbon analysis (de la Vega Leinert et al., 2012).

2.4.5. Deglaciation & the Bølling – Allerød (> c. 14,700 – c. 12,900 cal BP)

Orkney was most likely free of glaciers by ~ 16,000 years ago (Bradwell and Stoker, 2015). There are only a few records on Orkney that include the Lateglacial and they are located at Crudale Meadow and Quoyloo Meadow (Bunting, 1994; Moar, 1969; Timms et al., 2016; Whittington et al., 2015) and these Lateglacial records from exhibit the tripartite glacial – interstadial – glacial sequence in their sediment (Whittington et al., 2015). This sequence is characterized by minerogenic sediments changing to more organic material followed by a return to minerogenic sediment and then a final shift into postglacial organic sediment (Bunting, 1994; Whittington et al., 2015). Typically, the site starts with barren sediment after glacial retreat which then sees herbaceous taxa such as Poaceae, and Cyperaceae, and dwarf shrub heath being established (Fig. 2.13) along with the rise in organic content of the sediment and decreasing erosion in the catchment as vegetation stabilizes the soils (Bunting, 1994; Whittington et al., 2015). This would be consistent with the onset of the Bølling-Allerød interstadial (*c.* 14,690 cal BP). However changes in vegetation lagged behind changes in the sediment at these sites (Bunting, 1994).

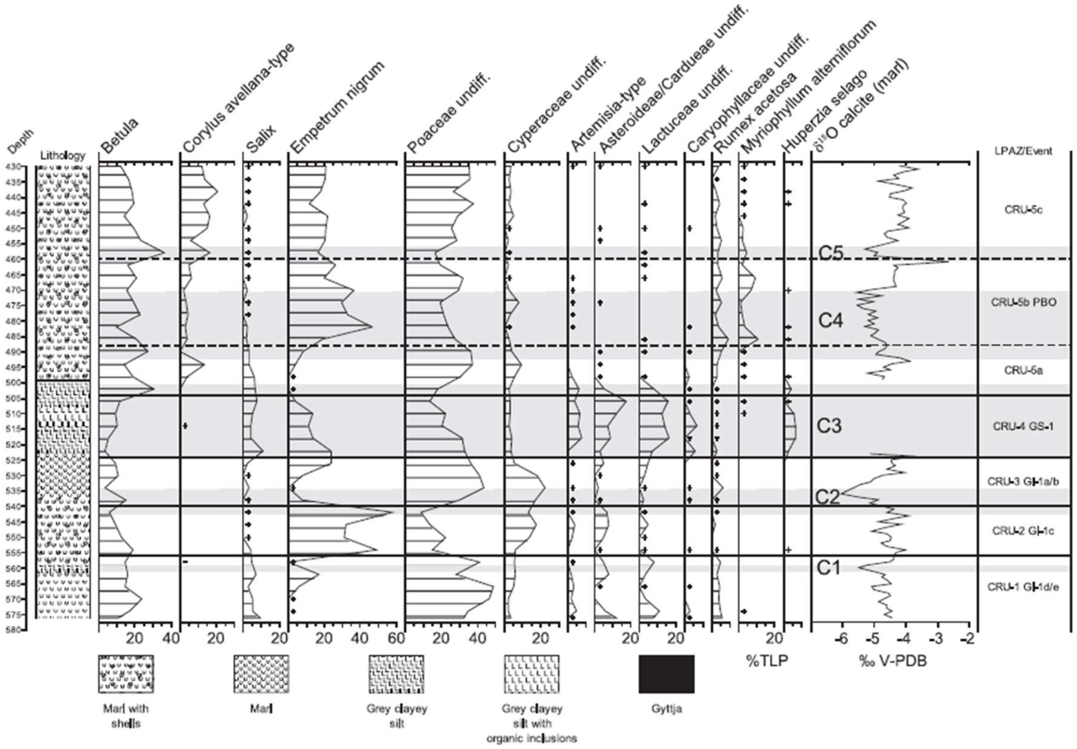

Figure 2.13 Lateglacial and early Holocene pollen and δ^{18}O sequence from Crudale Meadow, Orkney (from Whittington et al. (2015)).

2.4.6. Younger Dryas Stadial (c. 12,900 – 11,700 cal BP)

The Younger Dryas saw a return to a colder climate on Orkney with the possibility of glacial activity on Hoy (Bunting, 1994; Golledge, 2010). Minerogenic sediment returned with soil destabilization occurring in the catchment along with a reduction of organic sediment (*c.f.* Figure 2.12). Vegetation became sparse on Orkney, consisting of arctic-alpine dwarf species such as *Salix* (willow) and *Empetrum* (crowberry). *Artemisia,* which is associated with low temperatures and disturbance, (Birks and Heiri, 2010) is also found during the Younger Dryas on Orkney (Bunting, 1994). Not all pollen records indicate cold events during the Bølling and Allerød, however, GI-1d and GI-1b are present in the sediment and isotope record from Crudale Meadow (Bunting, 1994; Whittington et al., 2015).

2.4.7. *Early to mid-Holocene (c. 11,700 cal BP – c. 6000 cal BP)*

At the beginning of the Holocene, *c.* 11,700 cal BP, *Corylus avellana* (hazel) begins to dominate on Orkney along with tall herbs (Bunting, 1994). Progressing into the early Holocene until *c.* 6000 cal BP woodlands expanded throughout Orkney. However, on Orkney the timing of woodland decline ranged from *c.* 7250 cal BP on Hoy to as late as *c.* 3800 BP at Burn of Rusht on Mainland, with the decline occurring *c.* 6000 cal BP at most sites (Farrell, 2009). Following on from the decline in woodland, peat formation spread on the Orkney Islands with the earliest occurring on Hoy *c.* 8000 cal BP and the latest *c.* 2000 cal BP at Lesliedale Moss with the average occurring *c.* 3000 cal BP (Farrell, 2009). This shows the extent of regional variation in vegetation and landscape changes across this small island archipelago.

2.4.8. *Dating Complications*

Due to the calcareous nature of the sediment and the geology on Orkney, there has been difficulty in confidently dating these records using radiocarbon dating. Studies that have been undertaken on Orkney with continuous records generally cover short periods (i.e. ~ 2000 years (Leinert and Keen, 2000), or covering the last ~ 6000 years (Keatinge and Dickson, 1979)), and are dated solely by tephrochronology or biostratigraphy (Farrell, 2009). There are few studies that have radiocarbon dates (Bunting, 1996; de la Vega Leinert, 2007; Keatinge and Dickson, 1979) and they have been complicated due to the calcium carbonate in the sediment. The Freshwater Reservoir Effect (FRE) is caused by the uptake of dissolved infinite aged carbon from the water column by plants and other aquatic organisms which leads to older radiocarbon ages than the true age of the organic material being reported (Marty and Myrbo, 2014; Shotton, 1972; Yu et al., 2007). The AMS dating technique uses smaller samples of vegetable and other organic matter than conventional radiocarbon techniques, which reduces the need for bulk samples and the potential of contamination from old carbon from reworked sediment. However, the FRE may still affect AMS age estimates. The chronology building for Orcadian records has been assisted by the development of tephrochronology and the identification of distinct isochrones that have been accurately and precisely dated outside of Orkney (Timms et al., 2016). This will allow for more robust dating of Orkney records than was previously attained which will provide the foundation in determining synchroneity between and within proxies and study sites for this study.

2.4.9.Summary

Around 16,000 years ago, Orkney became ice free and was colonized by dwarf shrub heath and herb species resembling a sparse arctic-alpine environment. As the interstadial commenced thermophilous species began to appear. A return to glacial conditions occurred interrupted this advance in the middle of the Bølling-Allerød and again during the Younger Dryas. During this time, the presence of hazel was negligible until after the beginning of the Holocene when birch also began to increase and a tall herb grassland was dominant. In the early Holocene, a birch-hazel scrub woodland predominated with open areas consisting of grasses and sedges. Though it is not part of the scope of this research, the archaeological record put the arrival of humans on Orkney at *c.* 7500 years ago. Pollen evidence of woodland decline and a charcoal peak suggests human clearing (Bunting, 1994). The woodland decline on Orkney has been shown that it was not synchronous across the islands (Farrell, 2009) with two periods occurring, one *c.* 5500 cal BP and the other *c.* 3800 cal BP.

The palaeoenvironmental records from the LGM to *c.* 6000 years BP are scarce on Orkney. There are only three palaeoecological investigations that extend into the Lateglacial, *c.* 14,000 cal B; Crudale Meadow/Yesnaby and Quoyloo Meadow (Bunting, 1994; Moar, 1969; Whittington et al., 2015) located in north west Mainland and two that extend to *c.* 12,000 cal BP; Scapa Bay (de la Vega Leinert, 2007) and Stove (Morris and Emery, 1986)) in southeast Mainland. The lack of continuous, high resolution records from Orkney that cover the LGIT and early Holocene demonstrate the need for more comprehensive multi-proxy palaeoenvironmental records in order to better understand the nature and timing of climatic changes that have occurred across the North Atlantic region which this study will address.

2.5. Shetland

2.5.1. Present Shetland

Shetland is positioned in the North Sea 125 km north of Orkney and 210 km west of Norway at a latitude of 59° 30' to 60° 25'N, 0° 43' to 2° 7' W (Fig 2.13). It is an archipelago made up of over 100 islands, which includes Fair Isle to the south and Foula to the west, and presently it has a population of around 23,000 inhabitants (National Records of Scotland, 2012). Past glacial cover, which scoured the land leaving low hills and shallow lochs (Fig 2.14), has formed the topography of Shetland. The present climate of Shetland, like Orkney is influenced by the marine environment, which provides intense winds and mild temperatures that fluctuate from a low of 1.5ºC in the winter to average highs of 14.5ºC in the summer, with average yearly rainfall of 1257 mm and average yearly wind speeds of 14.5 knots (1981-2010 recorded at Lerwick (Met Office 2016).

The northern margins of the British and Irish Ice Sheet (BIIS) covered Shetland during the LGM and was one of the first areas to become ice free during the Lateglacial (Bradwell et al., 2008; Golledge et al., 2008; Hall, 2013). Thus, much of the isostatic rebound was achieved before the Holocene. By *c.* 10,000 years ago, sea levels were ~ - 65 m a.s.l., by *c.* 5000 BP ~ -10 m a.s.l. and reached present levels by *c.* 3000 BP (Lambeck, 1995). Due to early isostatic uplift during the Lateglacial, Shetland is the only place in Scotland not to have evidence of marine incursions during the Holocene (Sissons, 1967). Low lying ground on Shetland is mostly confined to the coasts on the West and in valleys running north-south through Mainland. Higher ground is found at Ronas Hill, the highest point of Shetland (Fig 2.15), in ridges running north-south through Mainland and a ridge along south Mainland island along with high regions on Unst (Gillespies, 1998).

Shetland has a very complex geology, and the Walls Fault is a main feature of Shetland geology which runs north-south through the middle of Mainland (Fig 2.16). Whalsay, Yell and the western portion of Unst along with Mainland is predominantly Caledonian Metamorphic rock. The southeast of Mainland along to the coast and the western lobe is made up of sandstone while the north-west portion of Mainland is mostly granite with some volcanic rocks at Esha Ness and Papa Stour and Metamorphic basement at Fetlar (Hall, 2013). The east portion of Unst consists of Oceanic Crust and Mantle. Shetland also has several major faults running mostly north-south through it (Fig 2.16). At the end of the last glaciation sea levels around Shetland were ~ -100 m a.s.l. (Bondevik et al., 2005) (Fig. 2.17).

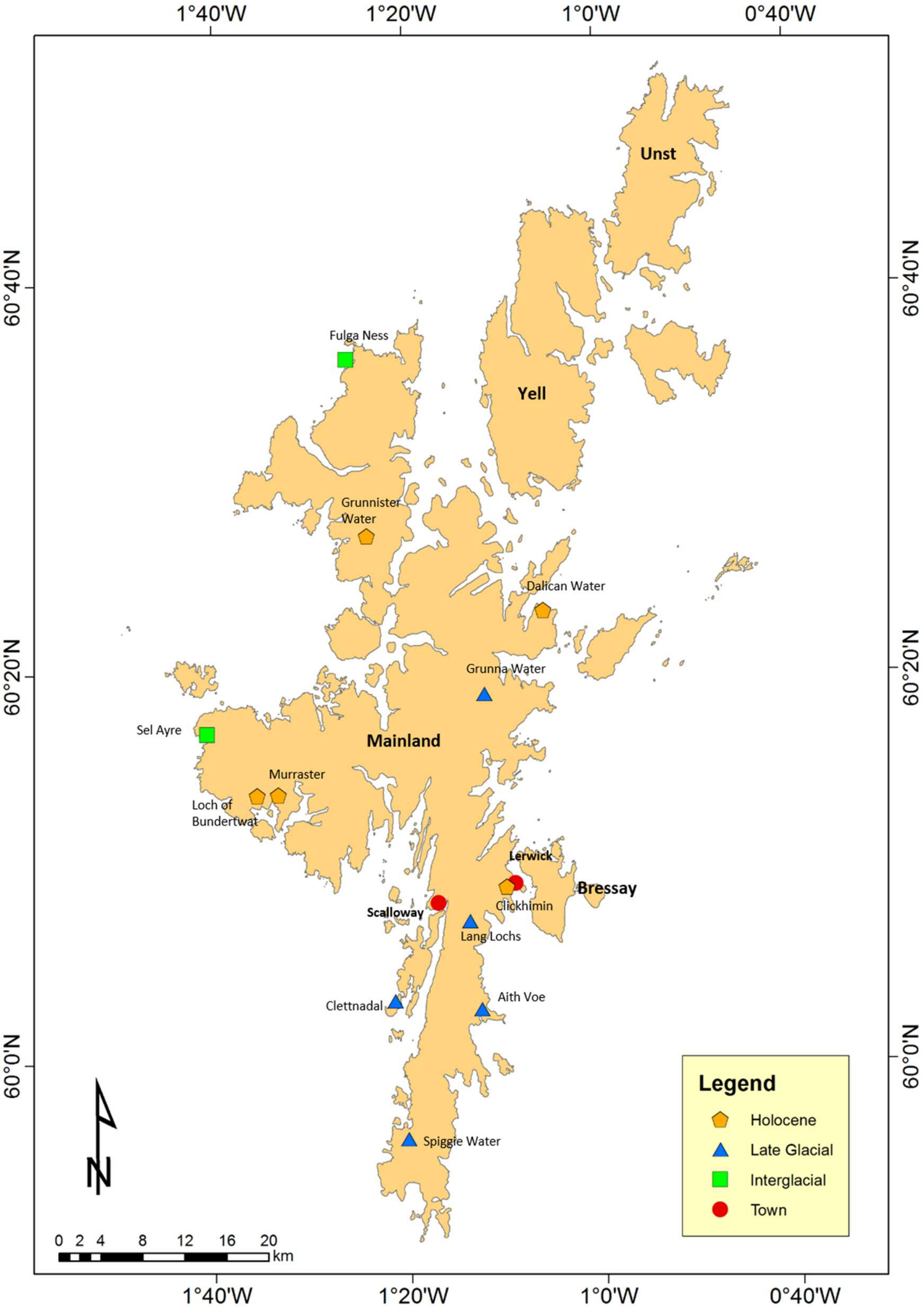

Figure 2.14 The island archipelago of Shetland showing present day towns (red dots) and sites that are mentioned in the text coded based on the time-period each covers.

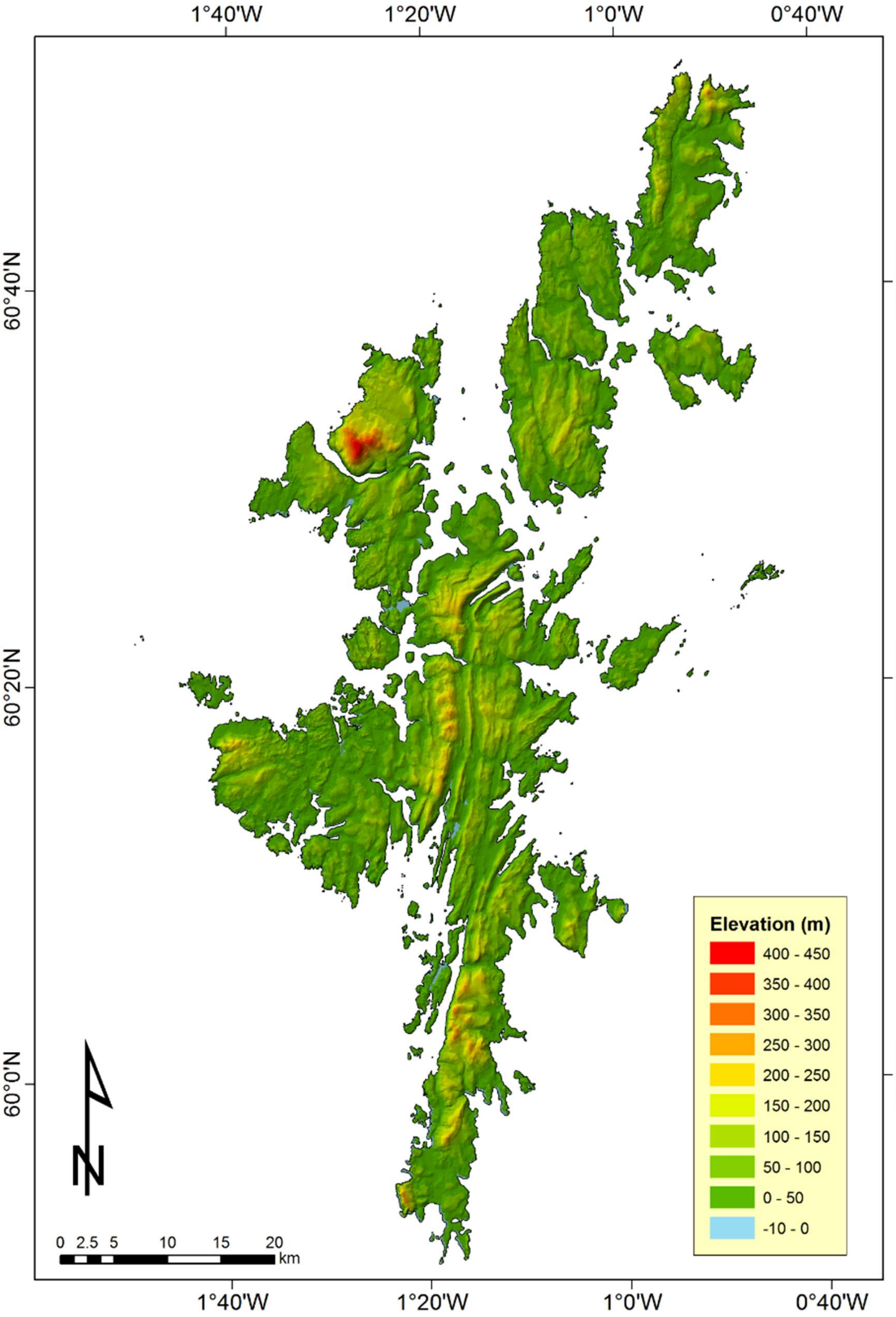

Figure 2.15 The topography of Shetland illustrating the gentle relief of the islands except for Ronas Hill (top left). Glacial scouring of the islands can be observed through the middle of Mainland (OS Terrain 5, scale 1:10000).

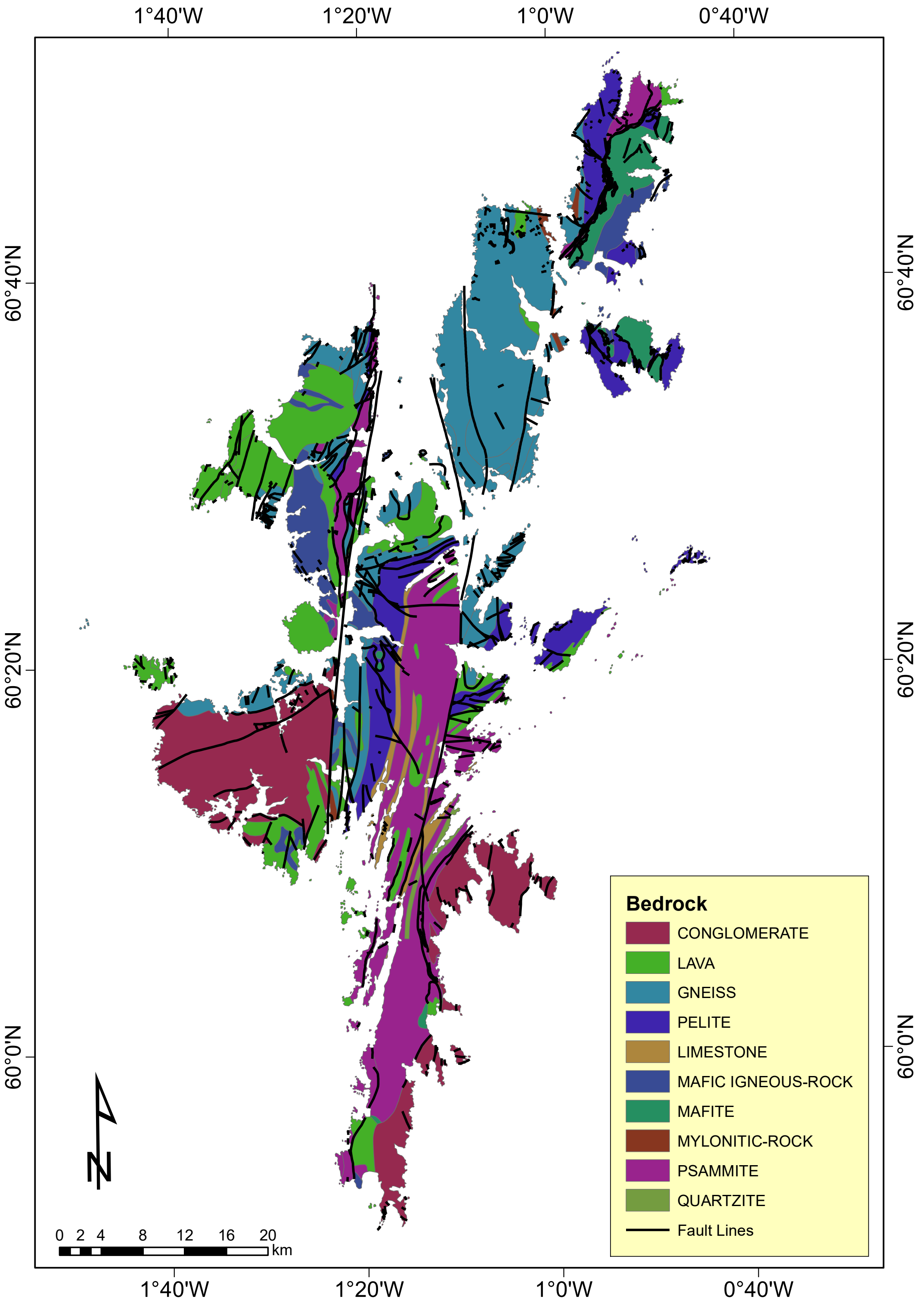

Figure 2.16 The complex geology of Shetland is illustrated in this map (EDINA, 2016). The main fault running through the islands from north to south is the Great Glen Fault and Walls Boundary Fault.

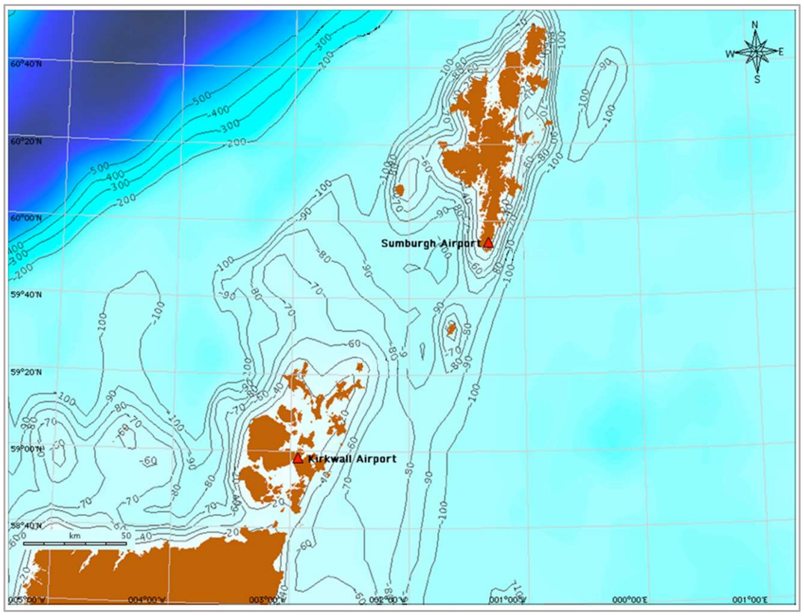

Figure 2.17 Bathymetry of the waters surrounding Shetland (modified from Lipa et al. (2014)).

Presently, extensive areas of Shetland are covered with peat along with peaty gleys and peaty podzols (Fig 2.18). The land cover of Shetland is made up of bog (grass and heather dominated, 31.3%), dwarf shrub heath (28.8%), rough low-productivity grassland (23.5%), and acid grassland (2.9%) while improved grassland only makes up 4.2% of total area (EDINA, 2007), which are concentrated in the valleys of central Mainland and south Mainland. The majority of Shetland is only suitable for rough grazing with limited areas of improved grassland while crops are limited to few low lying areas (Gillespies, 1998). Trees are only found presently on Shetland in sheltered locations, however there is a plantation located in the Weisdale Valley (Gillespies, 1998), south Mainland.

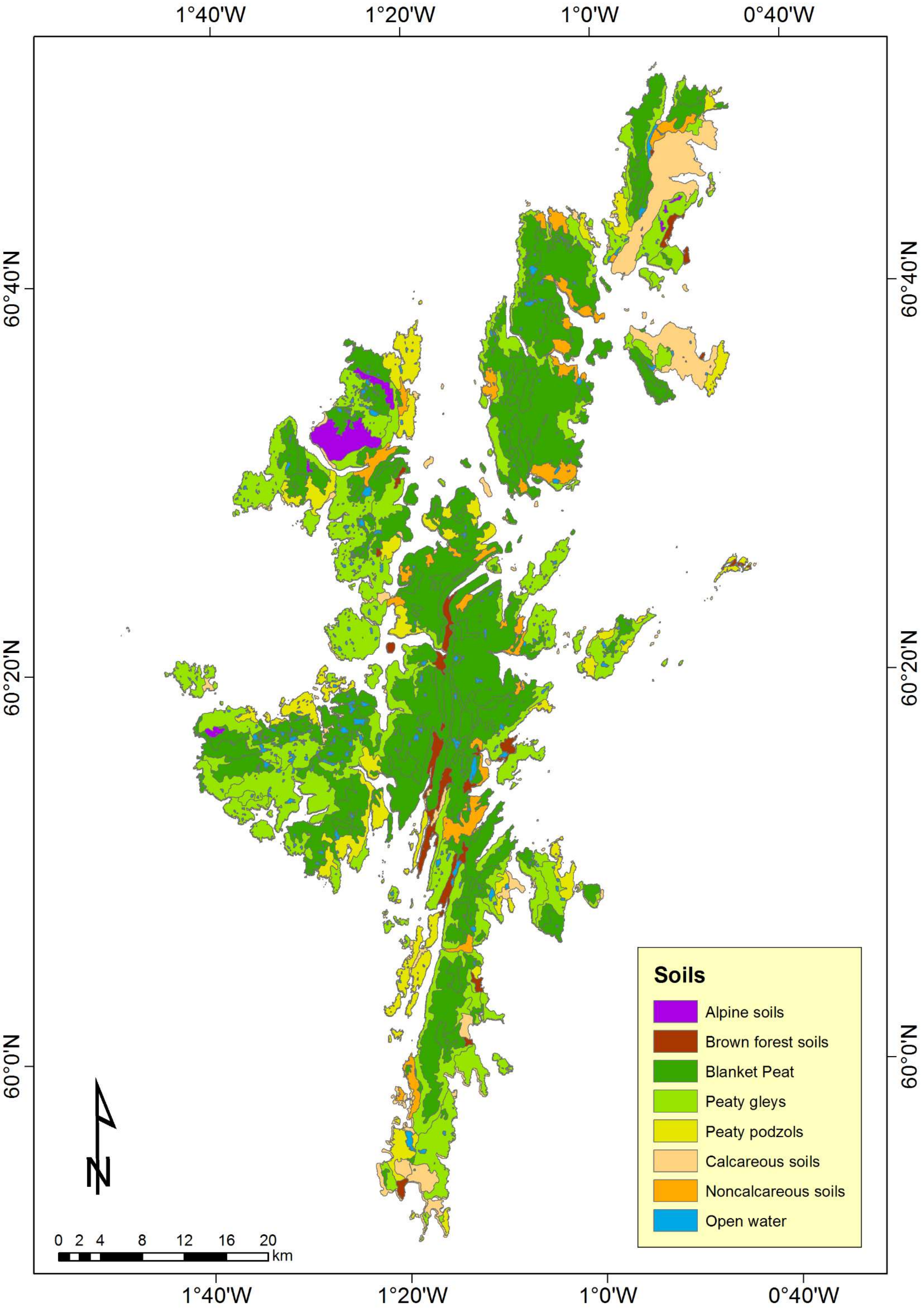

Figure 2.18 Soil profile of Shetland illustrating the extensive peat coverage on the islands (EDINA, 2007).

2.5.2. The Freshwater Aquatic Environment

There are over 1500 freshwater lochs scattered across the islands that make up Shetland, although they only account for 2.3% of land cover (EDINA, 2007). Due to Shetland's unique geology these lochs exhibit a wide variety of types more than what is found in any other region in Scotland. In general, the lochs on Shetland exhibit low biodiversity compared to similar lochs on mainland Scotland due it its isolation in the middle of the North Atlantic and North Sea. The harsh marine environment with high winds and salt spray also limits the number of species that can tolerate these unique conditions. Oligotrophic lakes comprise ~ 75% of the lochs which are characterised by low diversity (Maitland et al., 1994). There are only a few lochs that would be considered eutrophic, such as Loch of Spiggie. Other lochs on Shetland are classified as dystrophic due to the presence of peat which causes the water to become stained by humic acids leaching from the peats and thus reduces the amount of light penetration of the water column (Maitland et al., 1994). These lakes also tend to be low in nutrients and low in diversity with no marl lakes present.

2.5.3. Previous Palaeoecological Research

The palaeoenvironmental record on Shetland is quite sparse and there are very few complete high-resolution records that extend from the LGIT through the Holocene. Most records either just span the Lateglacial or sections of the Holocene that relate to associated archaeological sites. There are only three well-dated Lateglacial records from Shetland located at Lang Lochs, Clettnadal and Aith Voe (Birnie, 1991, 2000; Hulme and Durno, 1980; Hulme and Shirriffs, 1994; Robinson, 2004; Whittington et al., 2003) (Fig 2.14). Two other Lateglacial sites are located at Spiggie Water and Grunna Water, however, these are undated (Birnie, 1993, 1981; Edwards et al., 1993). These studies all used pollen as their principle proxy while Clettnadal, Aith Voe and Spiggie Water also included diatoms in the analysis. Lang Lochs also included macrofossils alongside pollen analysis (Hulme and Shirriffs, 1994) while Clettnadal included invertebrate analysis (Whittington et al., 2003).

When comparing these Lateglacial records, a few issues have been identified with the datasets. The records from Clettnadal and Aith Voe are very coarse in their resolution with ~ >500 years between samples. However, the pollen record from Lang Lochs is at a higher resolution with ~ 200 years between samples. Also, it is evident from the records that there are site-specific variations and therefore it is difficult to identify the nature and timing of the regional sequence of environmental change on Shetland. For example, Clettnadal is located on the west coast of the island of West Bura and reside on psammites and crystalline limestone and its sediment record is dominated by sand while Lang Lochs is located on Shetland Mainland near the east coast and is a mire-loch complex on metamorphic rocks. Aith Voe is also located on the eastern coast of Mainland and lies on Old Red Sandstone. The paucity of high-resolution, well-dated records from this region illustrates the lack of knowledge of how the development from the end of the last glaciation into the Holocene in this part of Shetland progressed in terms of regional variation and the nature and timing of these changes in comparison to one another and in the wider context of the North Atlantic.

2.5.4. Deglaciation & the Bølling – Allerød (> c. 14,700 – c. 12,900 cal BP)

Using the results from the dated records from Aith Voe, Clettnadal, and Lang Lochs a general description can be made about the sequence of environmental change in Shetland during the LGIT. Glacial retreat occurred sometime before *c.* 13,000 cal BP. However, no distinct interstadial organic layer was recorded at Lang Lochs. This may have been due to Lang Lochs being more inland and at a higher altitude than other sites such as Aith Voe and Grunna Water (Hulme and Shirriffs, 1994). However, further analysis confirmed a slight increase in organic content did follow deglaciation at Lang Lochs that could indicate interstadial conditions (but now visual change in sediment). In regards to the vegetation record, at Lang Lochs, the interstadial was dominated by *Rumex* whereas at Aith Voe, Poaceae, *Filipendula* and *Myriophyllum* dominated the assemblage (Birnie, 2000), while the interstadial at Grunna Water is dominated by Cyperaceae and small amounts of Poaceae (Birnie, 2000). This suggests that during this period, the landscape was open with short herb species dominating. The Lateglacial diatom records from Aith Voe indicate that the basin began barren with no diatoms in the basal sediment. As the lake matured, epiphytic diatoms colonized along with *Fragilaria (sensu lato,* (*s.l.*)). which is characteristic of interstadial assemblages (Birnie, 2000; Fritz and Anderson, 2013). This was followed by a reduction in diatoms with the return of colder conditions of the Younger Dryas (Birnie, 2000).

2.5.5. Younger Dryas Stadial (c. 12,900 – 11,700 cal BP)

The Lateglacial records from both Lang Lochs and Aith Voe do not show a clear shift from the Allerød interstadial to the Younger Dryas stadial in the pollen and diatom records. The pollen and diatom evidence at Aith Voe (Birnie, 2000) demonstrates that terrestrial vegetation disappeared during the Younger Dryas while the aquatic assemblage maintained a low amount of productivity which suggests that there might have been increased snow cover in the catchment (Birnie, 2000). As described in the previous section at Lang Loch, there was only a slight increase in organic matter that could indicate an interstadial. These conditions continue into the Younger Dryas, where there is no relative change in the pollen assemblage. Based on the ages provided, it is difficult to determine a clear Younger Dryas signal in the pollen record at both of these sites (Birnie, 2000; Hulme and Shirriffs, 1994).

2.5.6.Early to mid-Holocene (c. 11,700 cal BP – c. 6000 cal BP)

The Holocene record of the Shetland Islands is comprised of records that are found across the Mainland and includes Lang Lochs. Most Holocene studies on Shetland have been undertaken at sites with archaeological significance such as at Clickimin (Edwards et al., 2005), Catta Ness (Bennett et al., 1992), and West Mainland (Edwards and Whittington, 1998) while others focused on the vegetation history and the timing of tree decline on the islands (i.e. Johansen 1975; Hulme and Shirriffs 1994). Other studies looked for evidence of tsunamis on the coast of Shetland which is summarized in Bondevik et al. (2005). Similar to Orkney, the main proxy for palaeoenvironmental reconstruction on Shetland is pollen while the only studies that analysed diatoms through the Holocene used them to determine the sea level rise/ marine inundation at Norwick and Burragarth (Smith, 1993).

Of the well dated records for the Holocene on Shetland only a few commence at the end of the Younger Dryas. These include the records of Murraster (*c.* 12,270 cal BP), and Lang Lochs record which begins in the Lateglacial and continues into the Holocene while Dallican Water (*c.* 10,550 cal BP) and Gunnister (*c.* 11,200 cal BP) begin slightly later (Bennett et al., 1992, 1993; Hulme and Shirriffs, 1994; Johansen, 1975)) while the record at Brunatwatt in West Mainland covers the period from *c.* 6,800 cal BP to the present (Edwards and Whittington, 1998). There are few continuous records that encompass the LGIT into the Holocene and as mentioned previously, there are distinct regional environmental differences in their responses to climatic change across sites in Shetland.

There is some similarity between the records at Murraster (Bennett, 1993), Gunnister (Bennett et al., 1993), Lang Lochs (Hulme and Shirriffs, 1994) and Dallican Water (Bennett et al., 1992). The sediment at the start of the Holocene at each of these sites changes from minerogenic to more organic with the vegetation dominated by herb species and dwarf shrubs such as Poaceae, *Rumex* and Cyperaceae. At Murraster and Lang Lochs, Poaceae dominates the assemblage while at Gunnister the dominant pollen was *Rumex*, *Artemisia*, along with *Salix*, which is typical of early colonizers of a recently deglaciated environment. The appearance of woodland occurred at *c.* 9260 cal BP at Dallican Water, at *c.* 9400 cal BP at Gunnister, at *c.* 9700 cal BP at Lang Lochs and at *c.* 8900 BP at Murraster when *Corylus avellana* type appears. At each site, there is evidence of woodland decline and the formation of peat. The record at Dallican

Water shows several episodes of tree decline. The first was at *c.* 4000 cal BP and again at *c.* 3120 cal BP which is later than the evidence for the presence of people at *c.* 4500 cal BP (Bennett et al., 1992). While at Murraster there is landscape change that can be attributed to human influence at *c.* 4650 BP based on the presence of *Plantago lanceolata* which can be an indication of grazing by animals introduced to the islands by humans (Bennett, 1993; Johansen, 1975). At Gunnister there is an increase in charcoal presence in the record along with an increase in *Calluna vulgaris* occurring at *c.* 8670 cal BP, 7000 cal BP, and 2920 cal BP which is comparable to the record at Dallican Water (Bennett et al., 1992). At *c.* 2920 cal BP, there is an abrupt decrease in tree pollen at the same time at Gunnister (Bennett et al., 1993). The record from Lang Lochs indicates peat formation beginning *c.* 7000 cal BP and trees disappearing by *c.* 6000 cal BP with evidence of Neolithic activity at *c.* 3000 cal BP, suggested by the presence of grazing indicator *Plantago lanceolata*. The short record from Brunatwatt shows woodland reduction occurring at *c.* 5070 BP and blanket peat spreading by *c.* 4740 cal BP (Edwards and Whittington, 1998). The overall patterns of change emerging across Shetland during the Holocene demonstrate a degree of consistency. However, there are still localized differences such as the timing of tree decline across Shetland (Table 2.1) that need to be explained in the greater context of the region.

Table 2.1 The differences in the timing of key events in the pollen records of the Shetland Islands. All dates have been calibrated from their original sources using IntCal 7.1 (Stuiver et al. 2017) and are in cal BP from: [1]Hulme and Shirriffs (1994); [2]Bennett et al. (1992); [3]Bennett (1993); [4]Whittington et al. (2003) and Robinson (2004); [5]Edwards and Whittington (1998).

Event	Lang Lochs[1]	Dallican[2]	Murraster[3]	Clettnadal[4]	Loch of Brunatwatt[5]
Betula decrease	5240	-	5400	10,370	3770
Corylus avellana type decrease	3300	-	3360	-	3770
Calluna increase	8500	5100	5400	3760	5820
Plantago increase	3180	4400	5400	3760	5740

2.5.7.Summary

The extant palaeoecological studies from Shetland suggest there may be a pattern in the lengths of the records obtained and their geographical location. Studies that occurred north of Lerwick (60.15° N) are limited to the Holocene with reporting of an impenetrable till or gravel base limiting the length of the core, including at Murraster, Clickimin, Dallican Water, Garth's Voe Burragarth, Gunnister and Basta Voe (Bennett et al., 1992, 1993; Edwards et al., 2005; Johansen, 1975) (Fig. 2.13). Most Lateglacial records that occur on Shetland are found to the

South of Lerwick (Fig. 2.13) such as Spiggie Water, Lang Lochs, Aith Voe and Clettnadal (Birnie, 1991, 2000, 1981; Hulme and Durno, 1980; Hulme and Shirriffs, 1994; Robinson, 2004; Whittington et al., 2003). Grunna Water seems to be the exception, as it is a Lateglacial sequence that is located north of this boundary (Fig. 2.13), perhaps as a result of the topography (Edwards et al., 1993). These observations raise questions regarding the extent of the last ice sheet on Shetland along with the possible glacial coverage of northern Shetland during the Younger Dryas. This may be consistent with ice sheet modelling and tentative field evidence which suggests that following glacier retreat after the LGM there was a re-advance of an independent ice cap on Mainland Shetland during the Younger Dryas (Golledge et al., 2008; Hall, 2013). With the evidence presented above, the ice/snow cover during the Younger Dryas might have limited itself to the northern part of the Shetland islands, and the southern portion of the island below Lerwick had less ice and snow coverage during this time.

2.5.8. Tipping points, and critical transitions

There is pollen, geochemical and stratigraphic evidence of high magnitude and abrupt changes such as the Bølling / Allerød interstadial, the LGIT, and events such as the 8.2 ka event during the Holocene in the North Atlantic region (Schleussner et al., 2015). In comparison to the large environmental fluctuations evidenced at the end of the LGM, the Holocene events may appear as signal noise but they have been shown to have significant effects on humans in north-west European regions and may have contributed to cultural change (Bonsall et al., 2002; Plagányi et al., 2014; Tipping, 2010). Identifying the timing of any possible warning signals prior to an event will be important in defining the timing of any subsequent changes that may have occurred.

Andersen et al. (2009, p. 49) define ecological regime shifts as "abrupt changes on several trophic levels leading to rapid ecosystem reconfiguration between alternative states". These shifts are known as critical transitions which can result in a new ecological state (Randsalu-Wendrup et al., 2016). Anderson et al. (2009) outlined three main responses to changes that regime shifts could undertake: Type I, where the abrupt change is directly related to a response (Fig. 2.19a); Type II, where the ecosystem responds only after a threshold is surpassed and the relationship is non-linear (Fig. 2.19b) and Type III that is considered a bi-stable hysteresis system where the system changes back and forth between two alternate states (Fig. 2.19c) (Andersen et al., 2009). Randsalu-Wendrup et al. (2016) used the classifications of Anderson et al. (2009) and reviewed numerous palaeoecological records to define which type of regime shift occurred based on the observed changes and determined when the shift occurred, the length of the perturbation and the resulting duration of the change. Their review showed that regime change can result from both abrupt and gradual changes. However, it is important to understand the timescales in which these changes are acting and the low temporal resolution of many palaeoecological records can affect the resulting analysis, making it difficult to detect the length of shifts at lesser sampling timescales.

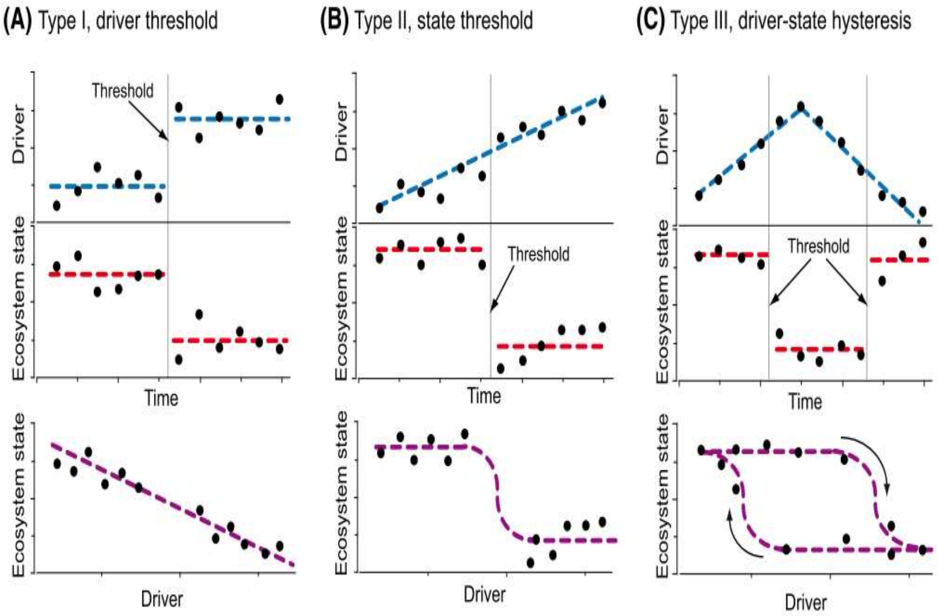

Figure 2.19 Illustration of different scenarios of regime shifts and how they theoretically react to external forces from (Andersen et al., 2009) Column a) illustrates driver threshold where it is a cause and effect relationship. Example b) shows a gradual change where a some point a threshold is reached and the state is changed. Example c) illustrates when a driver flips back and forth resulting in two stable states.

The methods used previously to detect regime shifts have been based on the analysis of variance (the change in variation approaching a tipping point), skewness (asymmetric fluctuations around the mean), kurtosis (how wide the ‘tails’ or outliers are in a distribution), autocorrelation (the similarity between consecutive observations), and critical slowing down (the observation that the closer to a tipping point the longer it takes to get back to equilibrium after a disturbance) (Boettiger et al., 2013; Dakos et al., 2012; Kéfi et al., 2014; Scheffer et al., 2009). However, it has been shown that these methods can only detect a change after it has become established (Eason et al., 2014; Spanbauer et al., 2014). These methods are also only useful for univariate analyses in simple systems, and they tend not to be effective in community-level multivariate systems (Spanbauer et al., 2014). It has also been observed that not all systems exhibit critical slowing down prior to a change (Dakos et al., 2015) making this method less robust under certain conditions. There has been an increase in the development of methods to try and detect these changes based on these theories as the need for this type of analysis is increasing due to the presently changing climate (Dakos et al., 2015; Scheffer et al., 2012; Streeter and Dugmore, 2013). Unfortunately, environmental systems are complex (multivariate) and tend to respond non-linearly to climactic forces acting upon them.

Fisher Information "is a measure of the information present in a dataset being used to fit an unknown parameter" (Eason et al., 2014, p. 773) and a measure of dynamic order (Fath et al., 2003; Mayer et al., 2006) or how much a system varies between perfect order or perfect disorder (Eason et al., 2014; Eason et al., 2016). Perfect disorder is when there is the same probability of measurements being in any state and perfect order is when measurements are consistently in the same state within natural variation (Eason et al., 2014). In analysing FI results several trends can occur. If FI is steadily decreasing the system is said to be losing dynamic order and the system is changing more quickly. Increasing FI results in a system that is becoming more ordered and stable while a sharp decrease between states can signify a regime shift (Ahmad et al., 2016; Eason et al., 2014). The method by which FI is employed is outlined in Chapter 3.

Fisher Information have been predominantly used for human recorded time-series data and applied to ecosystem management practices (Cabezas et al., 2010; Eason et al., 2016) and have been little used in palaeoecological analysis. Fisher Information is slowly making its way from ecology into palaeoecological literature to allow for analysis of longer timescales. Traditionally, in palaeoenvironmental research, constrained cluster analysis (CONISS) is employed to define changes in community composition (Grimm, 1987), and provides a point at which one set of observations is different from the next. However, this is the point at which things have already changed. The data from the three study lochs of this research will be used to explore the applicability and usefulness of Fisher Information and how it relates to CONSIS to explore changes in palaeoenvironmental data.

2.6. Summary

The Scottish Northern Isles are an ideal region to study the nature and timing of climatic changes in the North Atlantic. The islands are strongly influenced by the North Atlantic which influences both terrestrial and aquatic ecology (Ottersen et al., 2001). Despite the differing geology, Orkney and Shetland are considered in the same climatic region. Regardless of being made up of small islands, there is a heterogeneity of their landscapes, and therefore, there will most likely be differences in their response to changing climate. This study will determine the synchroneity of proxies representing terrestrial record (pollen), the catchment (μ-XRF) and aquatic record (diatoms) on Shetland and Orkney and to determine if there are subtle differences in the ecological response of the study lochs and surrounding catchments on these two archipelagos. It has been demonstrated that proxy records other than pollen (such as macrofossils and molluscs) are sparse on Orkney and Shetland and have been used as supplemental data against the main conclusions developed though pollen analysis. However, the nature of terrestrial vegetation, especially in long-lived tree species, produces an inherent lag in their responses to climatic change and, therefore, may not show up in the palaeoenvironmental record until decades after the change. Generalizations regarding the rate and timing of changes observed have also been made which tend to be oversimplified in these instances (Farrell et al., 2012).

There is also a lack of information regarding the nature and timing of changes from the last glaciation through the Holocene from the limited use of proxies that respond more rapidly to changes in the environment (such as diatoms) and therefore, studying aquatic proxies along with terrestrial and geochemical proxies at a higher resolution will provide a record of change at a higher temporal scale than previously studied in this area. This study will highlight the asynchronous nature of some of these changes and help to answer the research questions posed in section 1.2. This type of analysis will also allow for the detection of tipping points and regime shifts of the aquatic communities, especially ones that are at their limit of their optimal habitat. By studying these proxies (diatoms, geochemistry and pollen) more robust inferences can be made regarding the synchronous or asynchronous nature of climate change across Shetland and Orkney along with the North Atlantic region.

Chapter 3

3. Methodology

This chapter will outline the methods and rationale used in this study to address the research questions outlined in Chapter 1. This investigation used pollen and diatom analysis along with physical and chemical sediment characteristics to explore terrestrial and aquatic changes through time. The chapter will begin with outlining the site selection and field techniques and then discuss laboratory theory and methodology concerning geochemical analysis and chronology. Following that, theory and methodology for diatom and pollen analysis will be discussed. The chapter will conclude with an explanation of the data analysis and presentation.

3.1. Site Selection & Field Methods

The rationale for studying the Northern Scottish Islands is provided in Chapter 2. Two lochs were selected from Shetland and one loch from Orkney based on the availability of continuous sediment sequences that extended into the Lateglacial (> *c.* 14,000 cal BP) to allow for high-resolution sampling of the full core length. The lochs also needed to be accessible, and with limited fluvial inputs to avoid transport of pollen and diatoms from the wider hydrological catchment. On Shetland, the only suitable lakes for this study were located in the southern portion of the Mainland. Lochs in the northern portion of the Shetland Islands consistently did not have sediment profiles that continued into the Lateglacial (based on the lack of glacial clays and the inability to core past an impenetrable layer). These criteria for study sites were found in Loch of Sabiston Orkney, which is located on the northwest side of Mainland (Fig. 3.1). Despite repeated attempts to find a suitable loch on Shetland with both a Holocene and Lateglacial sequence, two lochs were chosen to cover this timeframe. Loch of Clumlie, located in the south of Mainland Shetland has a complete Lateglacial sequence, however, field observations determined drying out and possible hiatus in the top most portion of the core. Loch of Grimsetter, from the island of Bressay (Fig. 3.1), provides a Holocene record for Shetland with no evidence of a Lateglacial sequence. Specific details regarding the study lochs and site locations are provided in Chapter 4.

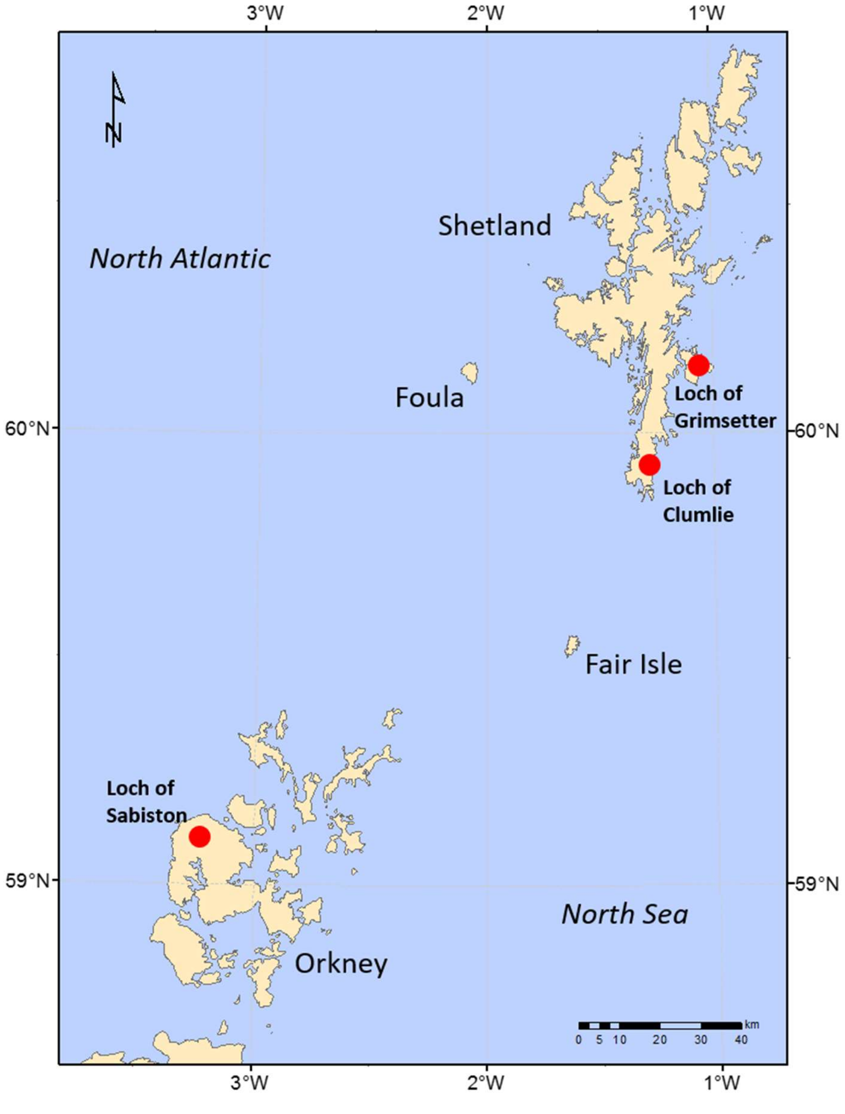

Figure 3.1 The locations of the three study lochs on Shetland and Orkney used in this study.

Preliminary cores were taken with a 1-m long (5-cm diameter) Dutch gouge across the loch from a small inflatable boat to determine the suitability of the sediment based on the presence of Lateglacial bluish-grey clays and silts. After the deepest sediment area was identified, a small raft was anchored over the coring site, (Fig. 3.2). A 1-m long Russian corer with a diameter of 75 mm was used to sample the sediment within the lochs until an impenetrable layer or bottom was reached. The core was sampled in 1-m sections with a 10-cm overlap to ensure a continuous record was recovered. Photographs were taken of the core sections in the field prior to being wrapped with polythene Layflat tubing, and returned to the University of Stirling and stored at a constant 4°C to reduce microbial activity and to retain moisture. At the Shetland sites, a gravity corer was used (core length = 50 cm, inner core diameter = 10 cm) to sample the topmost sediment of the profile due to insufficient sediment consolidation. The gravity cores were sectioned on site at 1-cm intervals using a vertical extruder, placed in labelled zip-lock bags, and then stored at a constant 4°C on return to the University of Stirling. This step was not required for Loch of Sabiston as well-consolidated peat covered the older lake sediment.

Figure 3.2 Example of the raft set up in the middle of the loch from Loch of Grimsetter, Bressay, Shetland. The red arrow points to location of the coring raft.

3.2. Geochemical Analysis

3.2.1. Sediment Characteristics and Loss on ignition (LOI_{550})

The core stratigraphy was described and the results were used to produce a sediment profile for all the biostratigraphical figures in this study. Loss on ignition (LOI_{550}) was used to estimate organic matter (organic carbon) content of lake sediments (Heiri et al., 2001) following the procedure outlined in Dean (1974). Samples for Sabiston and Clumlie were taken at 1-cm contiguous intervals along the length of the cores while Loch of Grimsetter was sampled at 2-cm contiguous intervals. The samples were dried at 105 ºC to remove moisture, cooled and weighed. Next, the samples were heated to 550 ºC for 4 hours to remove organic carbon, cooled, and weighed. The equation used is as follows (Boyle, 2003):

$$\text{Weight percent organic matter} = \frac{100\,(\text{weight at }105\,^{\circ}\text{C} - \text{weight at }550\,^{\circ}\text{C})}{\text{weight at }105\,^{\circ}\text{C}}$$

The results from analysing LOI can provide information on changes in productivity in the lake, which in turn can be used to determine warmer or colder periods and can signify changes occurring in the catchment along with other proxy data. It is accepted that the accuracy of LOI measurements of samples that contain less than 10 % may be affected and organic carbon underestimated (Mackereth, 1966).

3.2.2. micro-XRF

The micro x-ray fluorescence (μ-XRF) core scanner uses x-radiation to cause samples to emit fluorescence energy that is specific to certain elements and is commonly used in analysing the composition of rock and sediment cores for paleo-environmental analysis (Weltje and Tjallingii, 2008). Presently, it has not been used to support palaeoecological studies on Shetland or Orkney. However, wet chemical analysis has been completed for Dallican Water (Bennett et al., 1992) and Gunnister (Bennett et al., 1993) at a low resolution. The μ-XRF scanning for elemental analysis provides a rapid elemental assessment of the cores, which has been used to support interpretations based on the analysis of other proxies, along with providing information concerning tephra layers.

The chemical composition of lake sediments can provide information on the processes occurring in the catchment such as glacial advance and retreat, weathering, and erosion events which can be explained by climatic variability, seismic activity, redox changes, and anthropogenic changes such as pollution, agriculture and deforestation (Davies et al., 2015). The cores taken in lake sediment studies provide an ideal records of long term environmental changes as they can provide a continuous record of the lake and catchment. Originally, Mackereth (1966) outlined steps for interpreting geochemical analysis of sediment profiles and this was followed up by Engstrom and Wright (1984) with a comprehensive review of the development of the science since 1966 and determined that chemical stratigraphy is best used as a supporting technique along with biological methods such as pollen and diatoms as chemical changes are site specific.

The elements used for this study provide information regarding the lithogenic elements, grain size changes along with inferences regarding the biological components of the sediment. For example, the elements Al, Si, K, Ti, Fe, Rb, and Zr are all geochemically stable and are associated with lithogenic inputs and clastic sediments. Ratios such as Rb/K can indicate weathering regimes since K is soluble and Rb is less mobile during weathering (Davies et al., 2015). Changes in grain size can also be determined with the relationship between Fe/Ti where an increase in the ratio equates to smaller grain size and Ti/K, where an increase in this ratio equates to larger grain size, due to Ti being associated with silts and allochthonous input (Davies et al., 2015). The ratio Zr/Rb also has been associated with changes in grain size as Rb absorbs to clay material while Zr is more abundant in silts.

The results from μ-XRF analysis can also be used to infer changes in the biological component of the sediment. For example, Si/Ti can be used to estimate biogenic silica from diatoms, chrysophytes and sponges with higher values indicating high productivity (Brown, 2015). Organic content of the sediment can also be determined by useding the ratio of incoherent (inc) to coherent (coh) scattering. This scatter is produced when the X-ray interacts with elemental carbon and water (inc) and water (coh) and therefore a higher value indicates more organic content in the sediment. Br can also be associated with organic content and productivity of sediment as it forms strong covalent bonds with organic molecules (Gilfedder et al., 2011), however it can also beassociated with marine storminess in some instances (Davies et al., 2015). Ca/Ti can indicate calcium derived from autogenic sources in the lake such as evaporative concentration or biogenic production from vegetation (Davies et al., 2015). Finally, high Fe/Mn values can indicate anaerobic conditions at the sediment – water interface due to stratification of the water column or the result of decay from biological productivity in the lake as Fe and Mn are reduced with Mn being more affected (Davies et al., 2015).

The cores from the three study lochs were non-destructively analysed for a set of elements (Al, Si, P, S, Cl, Ar, K, Ca, Ti, Cr, Mn, Fe, Ni, Cu, Zn, Se, Br, Rb, Sr, Zr, Ba, Ce, and Pb) using a μ-XRF Itrax™ core scanner (Croudace et al., 2006) with a Bartington magnetic susceptibility (MS) instrument at Aberystwyth University. The cores were prepared by removing the top layer of the exposed surface sediment to remove any contamination and then covered with a 1.5 μm film to protect the core and reduce drying during the scanning process. The settings used for each of the cores are listed in Table 3.1. Due to the different sediment types found in Loch of Sabiston, different settings were used for the top core, which consisted primarily of peat. There was an issue with the core 90-190 cm from Loch of Sabiston for which the XRF current and voltage values were switched resulting in an intensity difference in the profile. The µ-XRF data for this section of the core have been corrected as the relationship between elements is consistent.

Prior to the µ-XRF scan, the core was photographed and a radiographic image was produced, and MS was measured continuously at a resolution of 2 mm. After scanning all the core sections, fine-tuning of the alignment between the sections was attained by comparing the µ-XRF elemental results of the overlapping sections and determining suitable tie-in points (distinctive peaks and trends) to produce a continuous sediment record. These tie-in points were then used to align the data for all other proxies.

Table 3.1 The parameters used for the µ-XRF scanning of the cores from each of the study lochs. The bold values highlight differences from the standard settings.

Loch/ Interval	Radiograph (voltage/current/exp time, step size)	XRF (exp time /voltage/current, step size)
Loch of Sabiston		
0-100 cm	60 kV/ 30 mA/ 200 ms, 500 µm step	10 sec/ 30kV/ 30mA, 500 µm step
90-190 cm	60 kV/ 50 mA/ 350 ms, 200 µm step	10 sec/ 50kV/ 30mA, 200 µm step
180-280 cm	60 kV/ 50 mA/ 350 ms, 200 µm step	10 sec/ 30kV/ 50mA, 200 µm step
270-370 cm	60 kV/ 50 mA/ 350 ms, 200 µm step	10 sec/ 30kV/ 50mA, 200 µm step
342-432 cm	60 kV/ 50 mA/ 350 ms, 200 µm step	10 sec/ 30kV/ 50mA, 200 µm step
Loch of Clumlie (all)	60 kV/ 50 mA/ 100 ms, 200 µm step	10 sec/ 30kV/ 50mA, 200 µm step
Loch of Grimsetter (all)	60 kV/ 50 mA/ 100 ms, 200 µm step	10 sec/ 30kV/ 50mA, 500 µm step

3.3. Chronology

3.3.1. Tephrochronology

Tephra is comprised of several different pyroclastic materials ranging in size and can be made up of pieces of rock, glass material and crystals of felsic and mafic (Turney and Lowe, 2001). This material ejected during volcanic eruptions presents a distinct layer in the sediment record and can be correlated by using dating and geochemical methods to volcanic events thus, providing a more robust dating profile of the core. This is possible due to the unique geochemical *fingerprints* of various eruptions determined by the location and timing of the emitting volcanos. In the North Atlantic, most tephra found originate from eruptions from Iceland (Dugmore, 1989), and many are now well known and well dated, such as Vedde ash layer which has been found throughout Scotland, Northern England and Norway (Lane et al., 2012). These well described layers can then be used as isochrones when comparing records.

Once the depth of tephra layers have been identified, the shards were then separated from the sediment using the protocol for acid digestion for diatoms using a 50:50 molar ratio of sulphuric to nitric acid (Rühland et al., 1999) to remove organic content. Tephra shards were mounted on a conductive phenolic resin stubs, and polished for geochemical analysis using Wavelength-Dispersive X-Ray Spectroscopy Electron Probe Micro-Analyser (WDS EPMA) (CAMECA SX-100) at the Tephra Analysis Unit (TAU), at the School of Geosciences, University of Edinburgh. An accelerating voltage of 15 kV, with a beam current of 2 nA for Na, Al, Si, Fe, K, Ca and Mg, and a beam current of 80 nA for F, Mn, Cl, P, S, and Ti with an operational beam diameter of 8.8 μm was used for this analysis (Hayward, 2012). The results were then compared to the elemental ratios of known and well-dated tephra layers for comparison from datasets located on the Tephrabase website (http://www.tephrabase.org/, accessed May 19, 2016; Newton et al. 2007) and corresponding dates found in relevant literature. The results were then used to constrain the age-depth model along with the Accelerator Mass Spectrometry (AMS) ^{14}C radiocarbon dates to provide a robust chronology for the three study lochs.

3.3.2. AMS Radiocarbon Dating

^{14}C AMS assay was carried out at the Natural Environment Research Council NERC Radiocarbon Facility in East Kilbride, UK. Samples were selected from each of the cores to constrain specific events in the stratigraphy such as a change in sediment or a notable change in the pollen or diatom records. Samples were pre-treated with 2 M HCl (to remove carbonates) at 80°C for 8h and then combusted by quartz tube to release CO_2. The samples from Loch of Clumlie and Loch of Sabiston were treated by an acid wash with 2 M HCl and the Loch of Grimsetter was treated using an acid-alkali-acid (AAA) pre-treatment to remove potential charcoal and other contaminants from the bulk sample. The AAA pre-treatment also removes any contamination due to secondary organic acids. The samples were then homogenized, combusted and the resulting CO_2 recovered and then converted to graphite by Fe/Zn reduction.

The resulting ^{14}C dates were calibrated using the computer program CALIB 7.1 with the IntCal13 calibration curve (Stuiver et al., 2013) and an age depth model for each study loch was developed using R ver. 3.3.2 (R Development Core Team, 2016) and the package Clam (Blaauw, 2010) with all ages are reported in cal BP. Clam employs a classical approach to age-depth modelling which takes into account the multimodal and asymmetric nature of calibrated dates (Blaauw, 2010) and was chosen due in part to the limited dating obtained for each core, and it also handles changes in sediment composition and hiatuses more effectively than Bayesian approaches. Clam uses Monte Carlo sampling (repeated random sampling) to determine the probability distribution of a given age at a given depth and then each of these dates in the core are sampled together over many iterations which result in point-age estimates and a curve drawn through these points to produce an age-depth model of calendar age estimates. The dates presented are a best fit date determined by a weighted mean (Blaauw, 2010).

There may be issues with AMS radiocarbon results from the Loch of Sabiston, Orkney due to potential hard water effects (input of carbon from older sources that can contaminate the samples). A hard water effect has been observed at Quoyloo Meadow (Bunting, 1994; Keatinge and Dickson, 1979) which was sampled from an area close to the Loch of Sabiston. Studies on Orkney have used tephrochronology, along with correlating dated sequences from nearby studies using pollen (Keatinge and Dickson, 1979; Moar, 1969) to produce feasible age-depth models. Using present-day techniques including Bayesian statistics coupled with tephra should provide robust age-depth models for all the study sites.

3.4. Diatom Analysis

Diatoms are unicellular eukaryotic algae (Division Bacillariophyta) and consist of two thecae (valves) and a girdle band (Battarbee et al., 2001). Diatom analysis was chosen as a proxy for environmental change in this study owing to their sensitivity to environmental changes due to: 1) their being highly diverse and ubiquitous; 2) their abundance in most aquatic systems (from occasional wetting to well-established water bodies); 3) their short generation times (Mann, 1999); 4) their ability to rapidly colonize new habitats (Heegaard et al., 2006); 6) their siliceous cell wall means they are often well preserved in the sediments of lakes, oceans, rivers and other aquatic environments; and 7) their microscopic size allows for high resolution environmental reconstructions from sediment cores (Smol and Stoermer, 2010). Generally, the diatom species found in each location are representative of that habitat's characteristics (i.e. physical, chemical and biological conditions) and thus the assemblage will change over time, reflecting changes occurring in the aquatic environment.

Diatom species assemblages vary by water depth and its associated aquatic characteristics (i.e. turbidity, dissolved organic carbon, and the extinction coefficient of light) and can provide information on the depth and stratification of a waterbody (Kingsbury et al., 2012). Planktonic species spend their time in the water column while benthic species are associated with various substrates (i.e. epiphytic = plants, epilithic = rocks, episammic = sand, and epipelon = mud) (Smol and Stoermer, 2010). There are also some species considered tychoplantonic, which are considered partly benthic and partly planktonic, spending time in the water column (Smol and Stoermer, 2010). Light transmission is another important factor when considering the effects of ice and snow on a loch. Ice cover alone will allow some light to penetrate allowing for algal growth, however ice and snow cover will obstruct light and impede algal growth. Prolonged ice and/or snow cover will change productivity, species composition, and lake chemistry, and therefore will be recorded in the palaeoecological record (Douglas et al., 1994; Douglas and Smol, 1999; Smol, 1988).

Species assemblages also correlate strongly with in lake variables such as salinity, pH or nutrients, and numerous studies have exploited these relationships to reconstruct past environments and to infer changes in aquatic ecosystems (Battarbee et al 2001). Examples of these applications include studies that track changes in ionic concentration of lakes to determine drought events (i.e. Fritz, 1993; Laird et al., 1996); using diatoms to determine changes in acidification due to atmospheric contaminants and subsequent recovery (i.e. Battarbee et al., 1999; Ginn et al., 2007; Jones et al., 1989); and in determining changes in eutrophication and nutrient input in lake systems (i.e. Bennion et al., 2004; Bennion and Simpson, 2011; Moos et al., 2009; Wiik et al., 2015b)

Diatoms have also been used to infer changes in the terrestrial environment. For example, the movement of treelines across a landscape over time have been inferred using diatom assemblages (Rühland and Smol, 2002). There have been some attempt at developing models to reconstruct temperature changes using diatom species composition, however many other factors of the aquatic environment are affected by temperature (such as chemical processes, ice cover, depth of stratification, pH, and nutrient cycling) and determining a direct relationship between diatom species assemblage change and temperature is problematic (Anderson, 2000). Despite diatoms not being a robust indicator of temperature fluctuations in reconstructions, their response to changing temperature can be indirect, especially during high magnitude changes in the Lateglacial where cold events which caused persistent ice and snow cover or reduced productivity of macrophytes would limit available habitat for certain benthic algae, thus possibly reducing species diversity (K. Rühland et al., 2003).

When using proxies such as diatoms, several limitations or concerns should be addressed. Dissolution and poor preservation can cause bias in the results as different species are affected differently; some are quite robust with heavy silica frustules whereas other species are more delicate (Smol and Stoermer, 2010). Poor preservation may also occur due to variations in pH, the concentration of available silica in the environment, and sedimentation rates. Typically, diatoms preserve best in cold soft water in northern latitudes. However, determining whether the dissolution is constant or if it changed over time can provide information about the environment (Smol et al., 2001). Taxonomic inconsistencies between researchers may result in bias in the counting and identification of the diatom species. There is also the potential for the misidentification of diatom species in large calibration datasets that are combined from different laboratories along with the application of these models. This can be avoided by using calibration datasets that contain photomicrographs to allow for comparison of the most common species to maintain taxonomic consistency (K. M. Rühland et al., 2003). For this study, diatoms are used as a proxy for climate reconstruction and for determining the environmental landscape throughout the LGM and early Holocene in the catchment area of the study lochs. To reduce taxonomic bias, the identification of species reaching a proportion greater than 3% in the assemblage was validated by comparing to published authorities and by another researcher.

Diatoms were prepared from 0.2 – 0.3 g of sediment sub-sampled from 0.5 cm sections of the cores. The interval at which each core was sampled was dependent on the loch and the accumulation rate of the sediment. Loch of Sabiston was sampled every 4 cm, Loch of Clumlie every 2 cm and Loch of Grimsetter every 8 cm. Due to the high amount of calcium carbonate ($CaCO_3$) (marl) deposition in Loch of Sabiston, 10% HCl was used as a pre-treatment to remove $CaCO_3$ as it causes a reaction with the sulphuric acid in the following digestion and produces gypsum which can form a coating on the diatoms making identification difficult (Rühland et al., 1999). The samples were then washed with distilled water several times before beginning the digestion process. After rinsing the samples with distilled water to remove the 10% HCl, 15 ml of a nitric acid (HNO_3) / sulphuric acid (H_2SO_4) mixture (1:1 molar ratio) was added to remove organic matter. The samples were left overnight and afterwards heated in a boiling water bath for at least one hour to accelerate the digestion process. The samples were then left for 24 hours to allow the fine sediment to settle, then rinsed with distilled water, and allowed to settle for 24 hours. This was repeated until a neutral pH was reached. This method is preferred to centrifuging the samples between each rinse to avoid damaging the diatom frustules (Rühland et al., 1999).

Microspheres (7000 series copolymer microsphere solution, 6 µm, concentration = 8.43×10^6 spheres ml^{-1}) were added (Battarbee et al., 2001) to allow for the calculation of diatom concentrations (Battarbee and Kneen, 1982). Known amounts of the rinsed sample slurries were then pipetted onto coverslips, dried, and mounted onto glass microscope slides using Naphrax (refractive index: 1.7). For each diatom sample, a minimum of 400 valves were counted and identified along full transects across the slide using a compound light microscope (Olympus BX43) under phase contrast with polarized optics under oil immersion (1000X magnification) to the highest taxonomic level possible using standard taxonomic references (Carter and Bailey-Watts, 1981; Krammer and Lange-Bertalot, 1991a, 1991b, 1988, 1986). Full diatom names that occur in the text for each of the study lochs are listed in Appendix A along with the naming authority.

A consideration specific to this study is that this region has been heavily influenced by anthropogenic activities for most of the Holocene. Similar studies have been carried out in other northern areas such as the Canadian Arctic (Michelutti et al., 2013), Greenland (Bichet et al., 2013; Perren et al., 2012), Denmark (Bradshaw et al., 2005), and Norfolk, UK (Fritz, 1989) which determined agricultural and anthropogenic impact on the catchment area of lakes. Changes in nutrients such as phosphorous, nitrogen and sulphur (eutrophication) have been used to indicate anthropogenic changes due to farming and animal husbandry and, therefore, were considered in this research. As the timeframe covered in this research is from the LGM to the early Holocene, it is unlikely that human influences have affected the climatic/ environmental responses identified during the LGIT and early Holocene on Shetland and Orkney.

3.5. Pollen Analysis

The study of palynology uses pollen, spores, and other palynomorphs to reconstruct past terrestrial environments and to track changes in vegetation due to human impact or natural processes. Since it is abundant in sediment, easily identifiable, and preserves well under anaerobic conditions such as in lakes, bogs, and fens it is a good indicator for the study of past terrestrial environments (Bennett and Willis, 2001; Birks and Birks, 1980). Pollen is produced in abundance from various trees, shrubs, and herbs, and deposited over the surrounding landscape in a 'rain' that is assumed to fall uniformly over a given area. This feature allows for the assumption that the assemblage is a uniform representation of the vegetation of the area (Birks and Birks, 1980). Pollen also possess a cell wall that contains sporopollenin that is resistant to degradation, which allows the grain to undergo the chemical processes needed for pollen analysis (Smol et al., 2001; Stolze et al., 2013). Fungal spores and charcoal are also preserved when preparing pollen, which can provide additional information regarding the terrestrial environment.

Palynological studies have provided essential information regarding the plants present in an area of interest and have could track changes in vegetation through time. However, one of the caveats of pollen analysis is that it is difficult to quantify community structure and ecological scales based on these records. Pollen analysis provides information on the community present near the lake. Trees produce pollen at different volumes along with different morphological features, which causes dispersal of pollen to be species-dependant. There have been attempts to provide this information (Birks et al., 2016; Fyfe et al., 2013) and models have been developed and used successfully in quantifying these processes (Marquer et al., 2014). This information, along with the state of preservation of pollen grains down-core allows for interpretations regarding the vegetation community structure. There is evidence that pollen can travel long distances and might not necessarily represent the local site vegetation, for example pollen may have been blown from the mainland to Orkney (Birnie, 2000; Bunting, 1996). However, there have been advances in pollen dispersal modelling that incorporate species specific pollen dispersal distances to get a better understanding of the distances that pollen can travel, and therefore, a better indication of the actual community that is being described (Fyfe et al., 2013; Sugita, 1994). This method was applied to the three lochs in this study and it was found that the relevant source areas for pollen of 600 - 800 m (based on the values given in Sugita (1994)) incorporated the area of the lake catchment therefore, discussing the pollen results in terms of catchment processes was appropriate.

Pollen was chosen as a proxy for environmental change in this study to provide a terrestrial record of catchment change to compare to the results of the diatom and μ-XRF analyses from each of the three study lochs. Pollen analysis has been the most common method of determining environmental changes in the Northern Isles to date. Therefore, its undertaking here will provide new records from three previously unstudied sites which will allow for a comparison to other records to be made and to increase the body of knowledge of environmental and climatic change during the LGIT and the early Holocene on Shetland and Orkney.

Samples of 1 cm^3 were taken every 4 cm of the core for Loch of Sabiston and Loch of Clumlie, and every 12 cm for Loch of Grimsetter for pollen analysis. Tablets containing a known quantity of *Lycopodium clavatum* spores (12540 spores/ tablet) were added to the samples to calculate pollen concentrations (Stockmarr, 1971). Samples were then chemically treated with HCl to first remove any $CaCO_3$. Next the samples were treated with NaOH 10% w/v to remove humic acids and sieved to concentrate the pollen grains following the method presented in (Moore et al. 1991). Hydrofluoric acid was added to the samples to remove any obstructing silica content and samples heated in a boiling water bath to accelerate the digestion. Lastly, the pollen samples were treated with acetolysis solution (1:9 ratio of concentrated sulphuric acid to acetic anhydride) to remove polysaccharides from the surface of the pollen grains making them easier to view. Pollen was placed in silicone oil after being dehydrated with tertiary butyl alcohol for analysis. An aliquot of prepared pollen was then placed on a slide with a cover slip anchored at the four corners using nail varnish to allow for the movement of the pollen grains during analysis. Pollen was counted using a light microscope (Olympus BX43) at 400X magnification in full transects across the slide and identified using the key in Moore et al. (1991) along with the reference collection at the University of Stirling. Nomenclature follows Bennett et al. (1994). Pollen was counted to a sum of 300 land pollen grains and values are expressed as percentage of total land pollen (TLP).

The condition of the pollen grains was also observed and graded based on five hierarchical categories (well preserved, broken, crumpled, corroded, and degraded, Fig 3.3 (Cushing, 1967; Tipping, 1987) and is presented as a percentage of TLP. This information can provide evidence of the processes that the pollen grains underwent after release either mechanical, biological, or chemical (Havinga 1984). For example, pollen grains found in a wet, acidic anaerobic environment such as a peat bogs or lake sediments tend to be in good condition while broken and crumpled are considered mechanical degradation attributed to changes in temperature and moisture (Campbell, 1999) or associated with in washed minerogenic sediment (Birks and

Birks, 1980; Tipping, 1987). Biological or chemical changes in pollen grains results in corroded and degraded pollen and can occur in aerobic conditions and chemical oxidation along with the actions of bacteria and fungi present in the sediment (Havinga 1967, 1984).

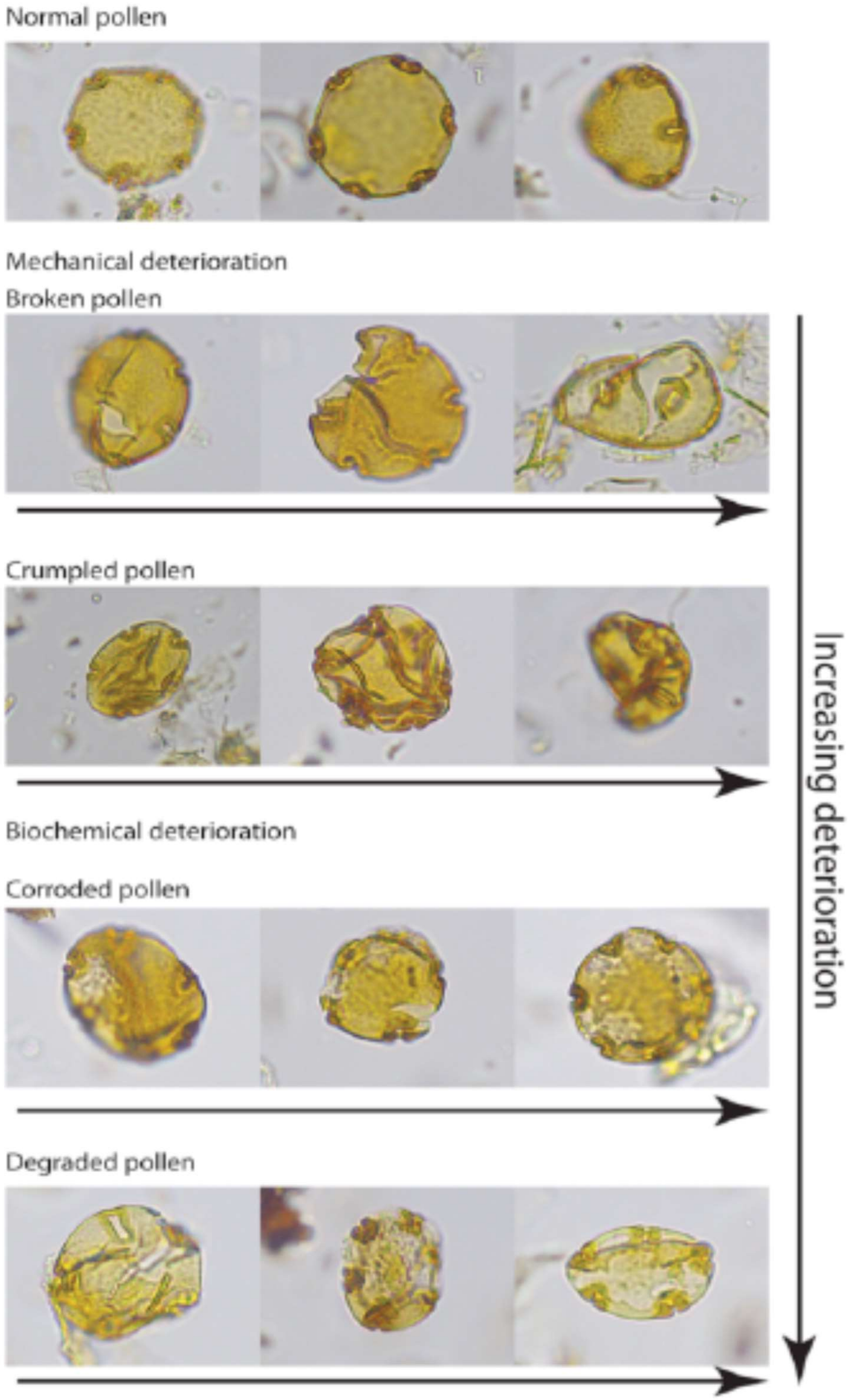

Figure 3.3 Illustration of the various conditions of pollen grains used to classify pollen preservation. Severity of each example increases from left to right, top to bottom (from Mansilla 2015).

3.6. Data Analysis

The diatom diagrams were plotted using the Tiliagraph function in Tilia version 2.02 (Grimm, 1987), and the species were organized by their weighted-average optimum to core-depth (e.g. Birks et al. 1990). For the summary diagrams, diatom taxa, which occurred at greater than 3 percent in at least one of the surface samples were used, and the remaining were combined into genera or ecological groups in each of the lochs. Pollen diagrams were also produced with Tilia version 2.02 and presented as percentages of Total Land Pollen (TLP), pollen concentrations and pollen preservation. To facilitate the description of the pollen and diatom assemblages and to determine points at which the assemblage significantly changes in the record, assemblage zones for both diatoms and pollen were defined by a constrained cluster analysis sum of squares (CONISS within the Tilia program (Grimm, 1987) using a squared-chord distance as a measure of dissimilarity. The threshold for defining zones was completed at the highest possible levels (total sum of squares), to define only the largest separations in assemblages along core-depth in each of the study lakes. The significance and strength of the CONISS zones were determined by the results from broken-stick and optimal splitting sum-of-squares zoning methods (Bennett, 1996).

The following data analyses were performed with R ver. 3.3.0 (R Development Core Team, 2016) and the packages Rioja and Vegan (Juggins, 2015; Oksanen et al., 2015) to get a better understanding of how the species assemblage has changed over time in response to ecological stressors. To simplify the patterns of change in the species assemblages in each lake according to sediment depth, a principle components analysis (PCA) was run based on all species that achieved at least 1% relative abundance. Species diversity measures used in this study include richness (rarefied), Hill's N2; a metric of species evenness (Hill, 2013), and Bray-Curtis similarity coefficient (Bray and Curtis, 1957; Clarke et al., 2006) to represent the species turnover as a function of core-depth. Estimates of total number of diatom species were based on all the counts from each of the lakes after counts were standardized by rarefaction to a valve count of 400 using the package Vegan. The evaluation of species richness (total number of species) and species evenness was carried out separately and not as an integrated diversity index due to the inherent loss of information that can occur with using such indices (James and Rathbun, 1981), especially with assemblages that consist of a few dominant species and many rare species. The perceived patterns reflected in an integrated index may in fact be an artefact of sedimentation rates in sedimentation studies (Smol, 1981).

3.6.1. Tipping points, and critical transitions

Recently, methods that use multivariate and time-series modelling have been developed to account for the dynamic interactions that occur at the community level. Fisher Information (FI) developed by Ronald Fisher as a measure of interdeterminancy is one such method and is used to measure order in a system (Ahmad et al., 2016; Karunanithi et al., 2008). Eason et al. (2014) refined this method which takes into consideration the multivariate nature of ecological communities and the resulting non-linear responses that can occur with a change in the assemblage. It also recognizes that abrupt changes may be the result of sudden tipping points and that prolonged instability can also instigate a regime shift (Spanbauer et al., 2014).

Fisher Information "is a measure of the information present in a dataset being used to fit an unknown parameter" (Eason et al., 2014, p. 773) and a measure of dynamic order (Fath et al., 2003; Mayer et al., 2006) or how much a system varies between perfect order or perfect disorder (Eason et al., 2014; Eason et al., 2016). Perfect disorder is when there is the same probability of measurements being in any state and perfect order is when measurements are consistently in the same state within natural variation (Eason et al., 2014). In analysing FI results several trends can occur. If FI is steadily decreasing the system is said to be losing dynamic order and the system is changing more quickly. Increasing FI results in a system that is becoming more ordered and stable while a sharp decrease between states can signify a regime shift (Ahmad et al., 2016; Eason et al., 2014).

Regime shifts in ecological systems are the result of long term changes in ecological communities that result in a change in status such as the changes that occur during eutrophication (Karunanithi et al., 2008). The theory is that a system will have natural variability and fluctuate within this variation when stable resulting in a stable FI. If the system is acted on by an external force, there will be a loss of order and the FI will decrease signalling change in the system while an increase FI would indicate that the system is becoming more ordered. Finally, a sharp decrease will indicate a possible regime shift in the system (Ahmad et al., 2016; Eason et al., 2014).

For this study, FI will be calculated to detect patterns in stability in the diatom, pollen and geochemical record and to determine what was occurring in the record prior to a shift in assemblage denoted by CONISS using the methods put forward by Spanbauer et al. (2014). For the diatom and pollen analysis, species/ groups achieving a proportion greater than 1% was used. The μ-XRF was corrected using inc + coh prior to analysis. For each proxy, a window of analysis had to be chosen along with step size. Diatom and pollen FI was calculated with a window size of two with a step size of one. The geochemical data, which is more continuous than either the pollen or diatom records, was sampled with a window size of four and a step size of one. In most of the examples given for this method, the time steps are uniform (e.g. every 10 years), however with the nature of palaeoenvironmental data and the different sedimentation rates that can occur in a single core, this research will also look at how FI handles an irregular sampling regime.

3.7. Summary

Using multiple palaeoecological and geochemical indicators as proxies for the terrestrial, catchment, and aquatic records will provide more robust environmental and climatic reconstructions then what has been done in the past. The combination of the dating techniques outlined will ensure a reliable chronological framework, especially due to the underlying problem of old carbon. This combination of techniques will provide novel information regarding the timing of palaeoenvironmental change in the Northern Isles and how they relate to the surrounding region. In the following chapters, these techniques have been applied to three study lochs, one from Orkney (Loch of Sabiston) and two from Shetland (Loch of Clumlie and Loch of Grimsetter), which when combined, will cover the Lateglacial and early Holocene for the two island archipelagos. The data analysis chosen for this research will allow for a quantitative investigation of how the terrestrial and aquatic environments have changed in the Northern Isles from deglaciation through the early Holocene and to determine the synchroneity or asynchrony of these changes within and among the islands groups. The ecological characteristics will provide information on how species assemblage change in relation to each other over time while computing Fisher Information will provide information on the trajectory of those changes.

Chapter 4

4. Results

This chapter covers the results from the three study lochs; Loch of Sabiston on Orkney, and Loch of Clumlie and Loch of Grimsetter on Shetland, along with the application of Fisher Information to the datasets of the three proxies. A comprehensive analysis of the results will be presented in Chapter 5, incorporating these results within the wider context of their respective archipelagos, the Northern Isles, and the North Atlantic region.

4.1. Loch of Sabiston Results

4.1.1.Study Area

The archipelago of Orkney is located 16 km north of Caithness, mainland Scotland, and is made up of about 90 islands, 14 of which are inhabited (Fig. 4.1.1). The topography of Orkney is one of low-lying areas and moderately sloped hills (except for the island of Hoy, which has a steeper landscape) (Chapter 2.4, Fig. 4.1.1). Loch of Sabiston is a freshwater loch situated in the northwest of the Mainland Island of Orkney (N 59° 04’58.6”, W 001° 16’ 35.9”, 15m a.s.l) and lies in a depressed area to the south-west of the Hill of Greenay (Fig. 4.1.2). The loch is currently shallow and alkaline surrounded by actively farmed land that is used primarily for grazing (EDINA, 2007) (Fig.4.1.3). Characteristics of the loch at the time of sampling are listed in Table 4.1.1. There is an outflow consisting of a small stream to the west that flows to the Burn of Wrath. The underlying geology is Old Red Sandstone and the soils consist of non-calcareous, peaty gleys (Fig. 2.6), along with arable pastures, swamp, and sedge mires with bent fescue grassland (Fig. 4.1.2). Presently, the landscape surrounding the loch is fen peat with multiple ditches dissecting it for drainage that date back to the early 1900's. There is also evidence of a probable crannog and stepping stones in the loch located at its southern end (RCAHMS, 1946).

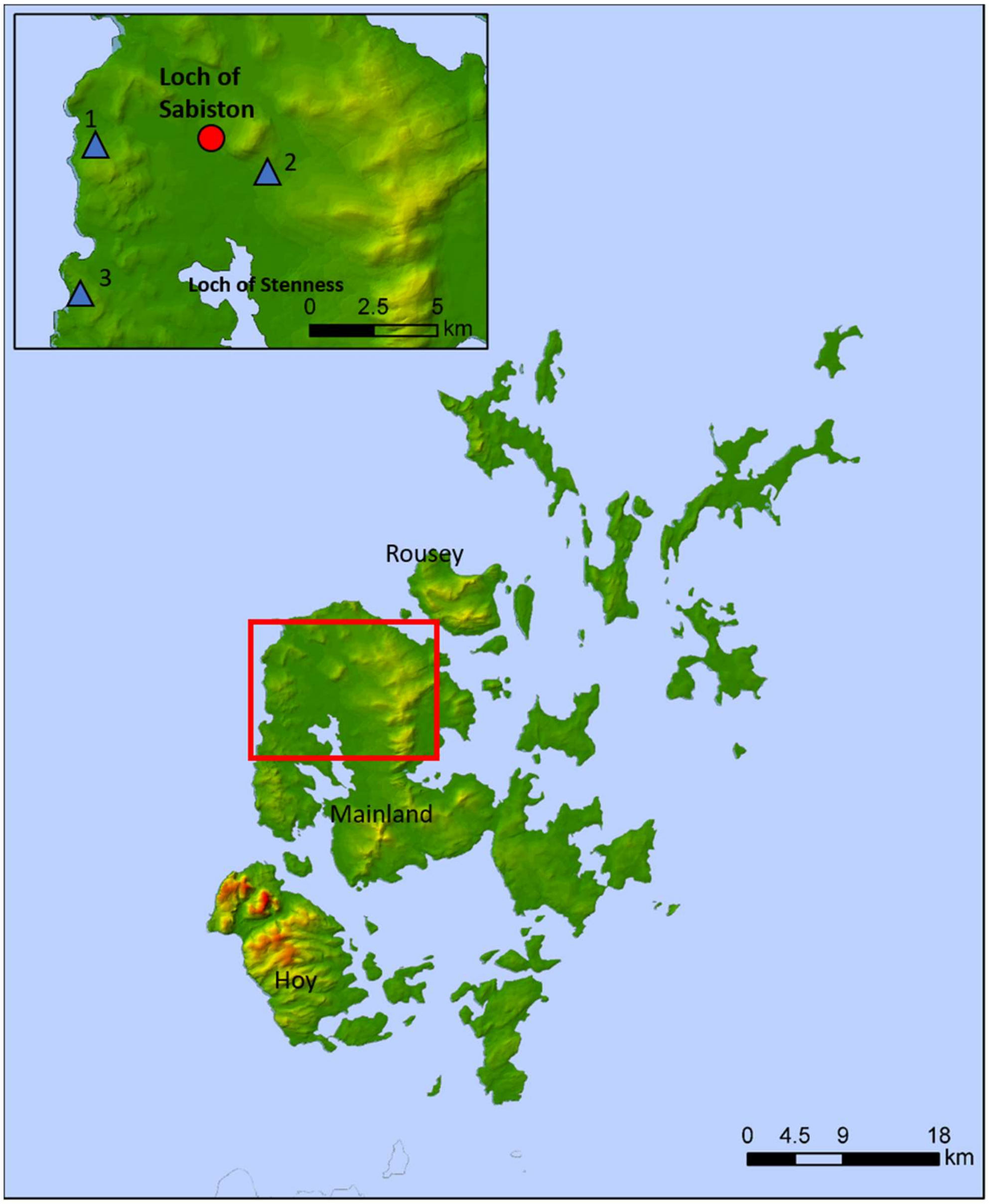

Figure 4.1.1 The island archipelago of Orkney, north of mainland Scotland showing the topography of the area with the inset showing the location of Loch of Sabiston and other palaeoenvironmental study sites mentioned in the text on Mainland Orkney; (1) Quoyloo Meadow (Bunting, 1994); (2) Glims Moss (Keatinge and Dickson, 1979); (3) Crudale Meadow (Bunting, 1994) / Yesnaby (Moar, 1969).

Figure 4.1.2 Land use (EDINA, 2007) around Loch of Sabiston with 10 m contours and the coring location on Sabiston designated with a red diamond. Inset shows the hydrological catchment area for Loch of Sabiston (Hughes et al., 2004).

Figure 4.1.3 The present-day view of Loch of Sabiston taken on April 29th, 2014 from the southeast looking north.

Table 4.1.1 Present-day characteristics of the Loch of Sabiston (taken June 2012).

Depth (max)	0.72 m
Area	~ 24 ha
Temperature	11.9 °C
Conductivity	347 µS cm^{-1}
Altitude	27 m a.s.l.
pH	8.18

4.1.2. Sediment Profile

A 432-cm undisturbed core was obtained from the loch (Fig. 4.1.4). The basal sediments consist of bluish-grey clay that continue to 354 cm. Marl then overlays the basal bluish-grey clays from 354 – 333 cm. At 333 cm, the sediment deposited switches to a silty clay until 324 cm then returns to marl deposition until 315 cm. From 315 cm to 305 cm, organic silty-clay deposition occurs which then changes to bluish-grey clay that lasts until 275 cm. A prolonged period of marl deposition begins at 275 cm and continues to 98 cm (Fig. 4.1.4, summary Table 4.1.2). There is evidence of charophytes, a macrophytic algae, throughout the marl sediment along with shells from freshwater snails of the genus *Lymnaea,* which are part of a sandy 2-cm shell layer occurring at 98 cm. Above the shell layer colluvium is deposited from 98 cm to 78 cm followed by a change in sediment to fen-peat which is then overlain by an organic mud at 20 cm that continues to the sediment / water interface (0 cm).

Table 4.1.2 Summary of sediment profile from Loch of Sabiston.

Depth (cm)	**Properties**
0 - 20	Organic lacustrine mud
20 - 78	Fen-peat
98 - 78	Colluvium
98 - 100	Sand/ silt layer containing shells of the *Lymnaea* species
100 - 120	Gradation from marl to organic silt
120 - 275	Cream-coloured marl with fragments of Chara present
275 - 305	Bluish-grey clay / silt
305 - 315	Gradation from marl to organic clay / silt
315 – 324	Cream-coloured marl
324 – 333	Bluish-grey clay / silt
333 - 354	Cream-coloured marl
354 - 432	Bluish-grey clay / silt

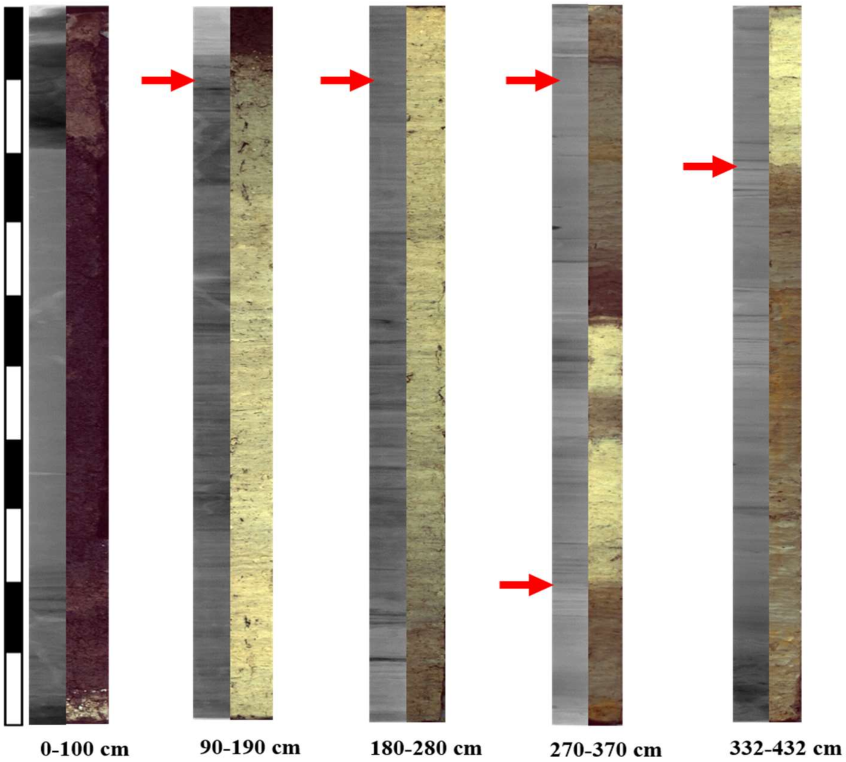

Figure 4.1.4 Sediment cores from Loch of Sabiston. The scale on the left indicates 10 cm intervals and the arrows on each of the cores signal the location of the overlap or tie in point in the final two cores. The grayscale picture to the left of the cores is an x-radiograph of the core illustrating the varying densities of the sediment units observed in the cores.

4.1.3. Loss on Ignition (550)

The organic content of the sediment from Loch of Sabiston does not rise above 10 % throughout the lake sediment portion of the core (from 432 - 100 cm) (Fig. 4.1.5). However, subtle increases of ~ 2 - 4 % do occur along with the onset of clay deposition at 357, 333 and 315 cm. From 280 - 120 cm, during marl deposition, organic content fluctuates between 3.16 - 8.74 % with an average organic content of 4.64 %.

4.1.4. Magnetic Susceptibility

Magnetic susceptibly (MS) was measured continuously every 2 mm for the full length of the core (Fig. 4.1.5). The higher values of magnetic susceptibility (~15 SI units) at the bottom of the core (430 - 355 cm) are characteristic of minerogenic input of silty clays into the loch through erosion due to bank and catchment instability. MS increases during clay/ silt deposition at 370 cm and 325 cm and including from 315 - 280 cm indicated by the sediment profile (Fig. 4.1.5). The

MS readings are low in the adjacent marl layers (from 355 - 315 cm) and stay consistently low through most of the core, except for a peak of 5 SI units at 212 cm and 42 SI units at 292 cm. At 120 cm, there was an increase of MS, which coincides with a change in sediment from marl to organic silt from more allochthonous inputs into the basin.

4.1.5. Micro-XRF Results

Twenty- five elements were detected using the μ-XRF core scanner with an average mean squared error (MSE) of 3.5. The results of the incoherent (inc) and coherent (coh) scans show that the moisture and organic content remains stable throughout the core. The inc/coh ratio can also be used as a proxy for organic content and is consistent with the results of the LOI_{550} analysis (Fig. 4.1.5). Each element was normalized to the sum of inc + coh (Davies et al., 2015) to account for Compton-Rayleigh scattering which reduces the effect of organic content and moisture variations on the μ-XRF profiles (Davies et al., 2015).

Overall, the highest positive correlations (Table 4.1.3) occur between the elements associated with minerogenic detrital input (K, Fe, Ti, and Rb) while Ca is negatively correlated due to autochthonous marl deposition. Strontium is positively correlated with calcium as it has a tendency to combine with carbonate to form $SrCO_3$ and will precipitate out of the water as a constituent of marl (Davies et al., 2015). Phosphorous also is affected by marl production in that it is removed from being bioavailable (Wiik et al., 2015b) (Fig. 4.1.5).

Table 4.1.3 Correlation matrix of R-values for select elements from the Loch of Sabiston. Correlation r-values over 0.75 have been highlighted in bold.

	Si	K	Ca	Ti	Cr	Fe	Se
K	**0.98**	1.00					
Ca	-0.51	-0.47	1.00				
Ti	**0.96**	**0.98**	-0.56	1.00			
Cr	**0.90**	**0.90**	-0.64	**0.94**	1.00		
Fe	**0.95**	**0.97**	-0.53	**0.97**	**0.92**	1.00	
Rb	**0.96**	**0.98**	-0.51	**0.98**	**0.93**	**0.96**	0.67
Sr	-0.41	-0.36	**0.83**	-0.45	-0.59	-0.44	**-0.70**

The relationship between individual elements provides important insights into the changing sedimentation at the site. The Mn/Ti ratio has a negative correlation with Fe/Mn (Fig. 4.1.6) with increases during the beginning of marl deposition and at 80-20 cm.

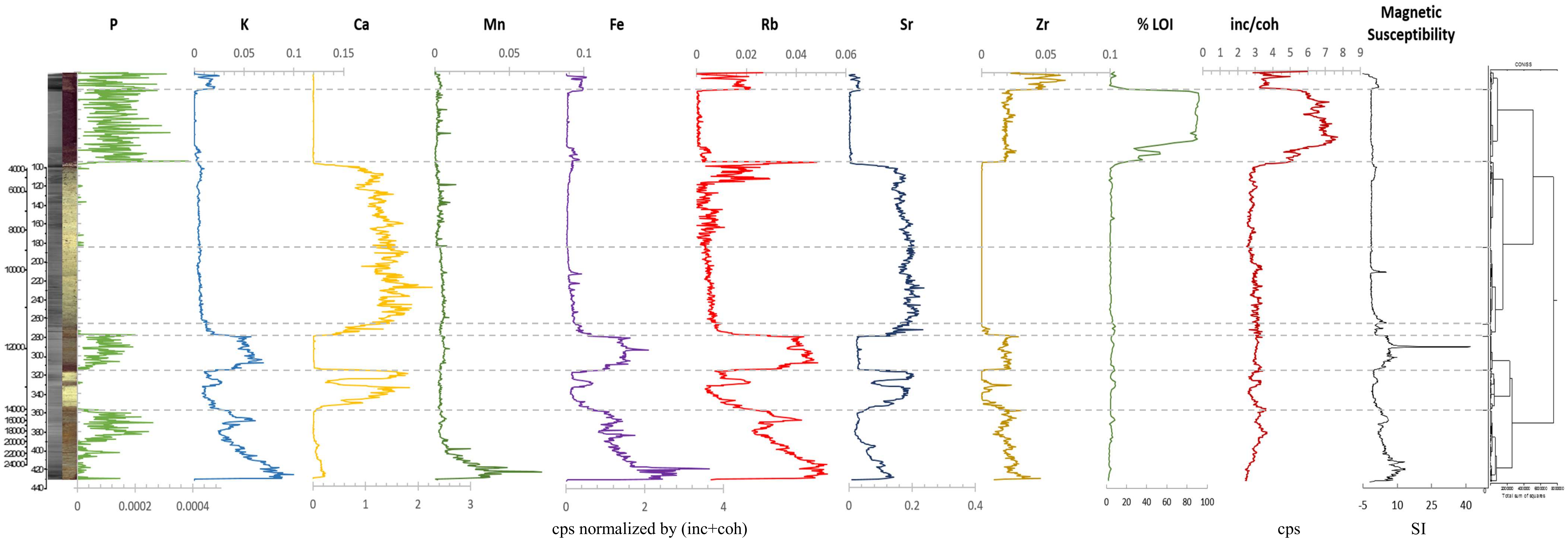

Figure 4.1.5 Lithology, organic matter content (% loss-on-ignition at 550°C (LOI_{550})), magnetic susceptibility and µ-XRF geochemistry from Loch of Sabiston. Selected elements from the µ-XRF scan. Fe, Ti, K and Rb, are associated with various catchment input, Ca is associated with marl deposition and Br is associated with storminess. P and Si are associated with diatom growth, and the deposition of marl limits P. Zones are based on the constrained cluster analysis. Elements have been corrected for organic and moisture content by dividing the elemental results by the sum of incoherent and coherent scatter results.

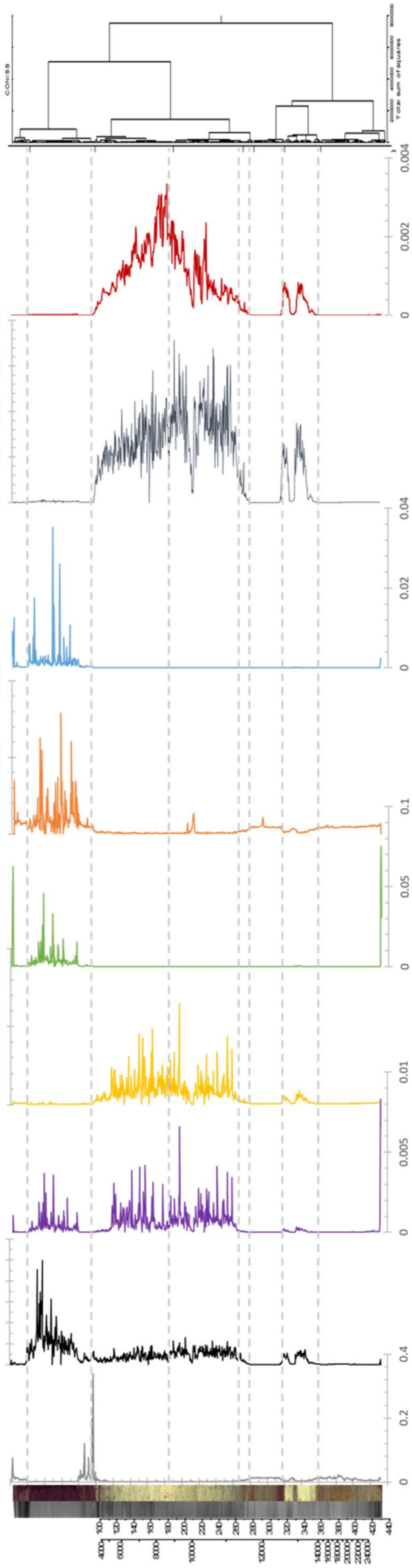

Figure 4.1.6 Selected elemental ratios from µ-XRF scan from Loch of Sabiston. Zones are based on the constrained cluster analysis of the µ-XRF results. Ratios were determined by using elements corrected for organic and moisture content (Davies et al., 2015).

Physical weathering associated with K/Ti and chemical weathering associated with Rb/Sr also correlate well with the changes in sedimentation. Physical weathering is associated with marl deposition while chemical weathering occurs during clay and silt deposition (Fig. 4.1.5). The ratio Fe/Mn which has been used for determining reducing conditions is high at the base of the core gradually decreasing until ~ 250 cm where it drops significantly. This is the same point at which Mn/Ti ratio (proxy for oxic conditions) peaks. At 170 cm, the Ca/Fe ratio indicates that an increase in iron-based accumulation occurred along with an increase in the Zr/Rb and Zr/Ti ratios, which may indicate an increase in grain size (Davies et al., 2015) (Fig. 4.1.6). At 210 cm, there is a reduction in Ca/Fe, Mn/Ti and K/Ti along with a spike in Ti/K which suggests a decrease in pedogenic inputs, a return to anaerobic conditions, a decrease in physical weathering, and an increase in grain size (Fig. 4.1.6).

There is also correlation between Fe/Si, K/Al, and Fe/Si ratios with Zr/K, Zr/Rb, and Zr/Ti. The first three are indicators of fine silt/ clay deposition and physical weathering in the catchment. These elemental ratios increase at ~ 75 cm while the latter Zr-based ratios decrease at the same time which are indicative of coarser grain size of the sediment. Ca/Si has been used to track a change in water temperature (Jouve et al., 2013), which in this core is showing a steady increase from the base to the top of the core (Davies et al., 2015). Finally, the Br/Ti results, which can indicate increased salinity or marine influence, are showing peaks throughout the record (Davies et al., 2015).

4.1.6. Chronology

4.1.7. Tephra Analysis

Two tephra layers were located at 212 cm and 292 cm respectively using the peaks found in the magnetic susceptibility results along with peaks in the Ti/K ratio of the µ-XRF results. Eleven glass shards were geochemically analysed from 212 cm and thirteen glass shards from 294 cm at the Tephra Analysis Unit at the School of Geosciences University of Edinburgh. The results of the geochemical analysis (Table 4.1.4, Fig. 4.1.7) suggest that the tephra located at 212 cm matches that of Saksunarvatn tephra dated at 10,176 ± 49 cal BP (Bronk Ramsey et al., 2014) and the tephra layer at 292 cm matches the tephra from Vedde which is dated 12,023 ± 43 cal BP (Bronk Ramsey et al., 2014). The Vedde ash was made up of primarily the basaltic component of the record instead of the more commonly found rhyolitic shards (Davies et al., 2012; Mortensen et al., 2005). The eruptions have been robustly dated and provide anchor points to construct the age-depth model along with AMS radiocarbon dates.

Table 4.1.4 Major oxide concentrations of basaltic shards from Loch of Sabiston using WDS EPMA (CAMECA SX-100) at the Tephra Analysis Unit (TAU), School of Geosciences University of Edinburgh.

Al_2O_3	CaO	FeO	K_2O	MgO	MnO	Na_2O	SiO_2	TiO_2
Sabiston (212 cm)								
13.1674	9.7715	14.7037	0.5100	5.3384	0.2385	2.4126	49.3770	3.1567
13.3203	9.5953	14.8097	0.4764	5.3800	0.2328	2.8747	50.0082	3.1790
12.8910	9.4113	15.7715	0.6122	5.0732	0.2585	2.8755	49.8516	3.2697
13.5144	9.8517	14.5721	0.4613	5.1112	0.2329	2.5986	49.2312	3.1434
13.5210	9.4934	15.1235	0.3050	4.8533	0.2416	3.2108	49.7954	2.9368
13.0603	10.1545	14.1978	0.4428	5.6188	0.2263	2.7557	49.1098	2.9669
12.9663	9.7240	14.7023	0.5229	5.2310	0.2359	2.8676	49.7929	3.1649
14.0323	12.0738	12.6974	0.2672	7.1859	0.2252	2.3674	48.7430	1.8937
13.2041	9.57010	14.9893	0.4969	5.5016	0.2244	2.8752	49.9374	3.1415
12.8465	9.6439	14.3165	0.4615	5.2918	0.2322	2.6590	49.4053	3.1409
Sabiston (292 cm)								
13.0574	9.8276	14.5046	0.7500	4.9067	0.2318	3.0315	47.1367	4.5579
12.9349	9.8204	14.7466	0.7385	5.3299	0.2177	3.0479	46.8954	4.5694
12.9256	10.0541	14.9153	0.7835	5.1700	0.2366	3.0307	48.0500	4.6077
13.5731	10.0539	15.792	0.8232	5.2252	0.2293	2.9542	46.7872	4.5769
13.5493	9.7268	15.5298	0.7994	5.1270	0.2376	2.542	48.1986	4.2762
12.5296	11.2183	14.6436	0.6268	6.5495	0.1985	2.4112	47.7109	4.5396
12.6053	8.8148	14.5164	0.8002	4.9088	0.2147	3.3330	47.7880	4.5042
13.5405	9.8496	15.0592	0.8809	4.9300	0.2339	2.9475	48.7610	4.7298
12.7716	9.7921	14.7262	0.8121	5.0614	0.2168	3.1308	46.6675	4.6822
12.9862	1.2168	3.7640	3.4604	0.1935	0.1335	5.0268	68.8283	0.2711
12.8916	10.0781	14.3763	0.8176	4.8172	0.2280	2.8413	47.5897	4.4875
13.3288	10.2782	14.7005	0.6586	4.7086	0.2375	3.1773	48.3428	4.9960

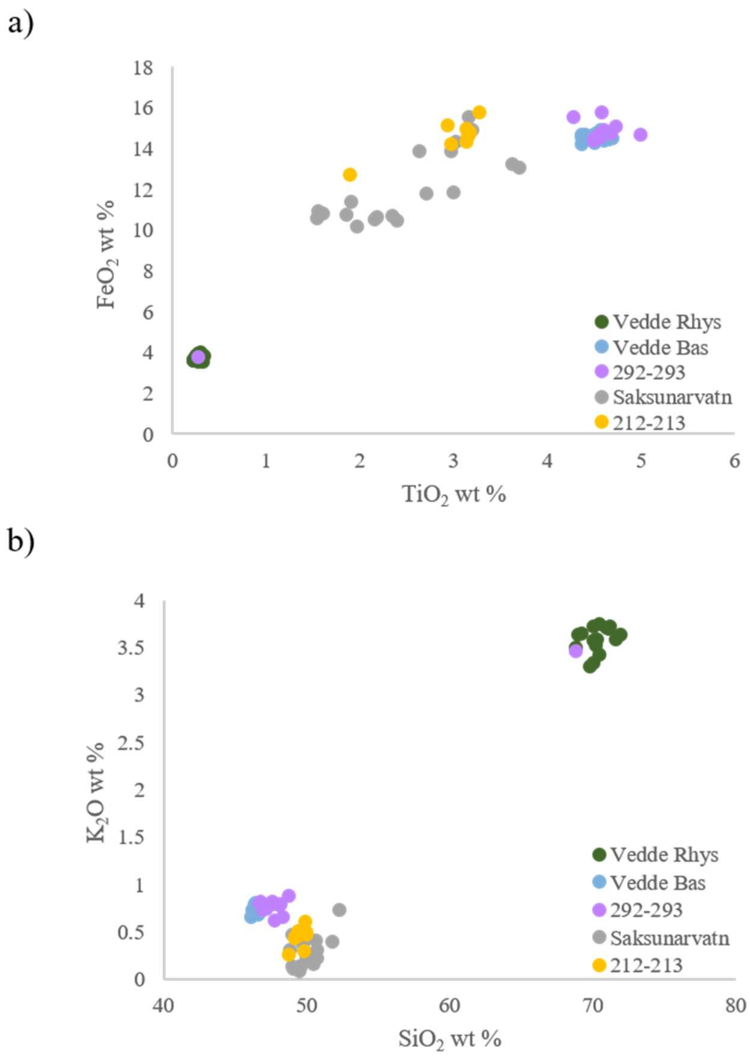

Figure 4.1.7 Geochemical results of the tephra analysis showing chemical bi-plots of major chemical fractions. Sabiston data is compared with reference data from the Tephrabase website accessed May 12, 2016 (Newton et al., 2007).

4.1.8. Radiocarbon Dating

Historically, ^{14}C dating on Orkney has been problematic due to the carbonate rich nature of the sediment and soil and therefore, constructing a robust chronology is difficult (Whittington et al., 2015). Nine radiocarbon dates have been obtained from the Loch of Sabiston core and two well dated tephra layers supplement these dates (outlined above). The radiocarbon dates for eight of the samples were macro-plant fibres sieved from sub-sampled core material, while the ninth radiocarbon age was obtained from a bulk sample of fine detrital organic rich mud due to the lack of suitable macro-plant fibres. The results of the analysis of the nine samples and two tephra samples are summarized in Table 4.1.5. Sample SUERC-67382 was omitted as it was deemed too old based on the tephra dates while SUERC-67383 appears to be also from aquatic carbon sources based on the δ^{13}C results. These two dates were outliers and have not been included in the age-depth model as they are out of sequence (Table 4.1.5). Sample SUERC-67384 and SUERC-67379 were probably contaminated by younger material, either by root penetration from younger overlying layers or through laboratory contamination during the sub-sampling and sieving to isolate the plant fibres. The age estimates are anomalously younger and out of sequence and are also excluded from the age-depth model. The satisfactory dates along with the tephra were then used to construct an age-depth model (Fig 4.1.8) using R (R Development Core Team, 2016) and the package Clam (Blaauw, 2010).

Table 4.1.5 Radiocarbon dates submitted to NERC March 2016, calibrated using CALIB 7.1 (Stuiver et al., 2013) and the IntCal13 atmospheric calibration curve (Reimer, 2013). Samples omitted from the age-depth model have been marked with *. The dates for the tephra layers were obtained from Bronk Ramsey et al. (2014).

Sample ID	Depth (cm)	Material	^{14}C age ± error	δ^{13} C$_{VPDB}$ ‰ ±0.5	Calib cal age BP (2σ)	Calib wmean cal BP
SUERC-67389	99 - 101	plant fibres	3,460 ± 36	-25.1	3,638 – 3,832	3,730
SUERC-67388	114 - 116	plant fibres	4,790 ± 38	-24.5	5,465 – 5,601	5,519
SUERC-67384*	155 - 156.5	plant fibres	1,517 ± 37	-24.0	1,335 – 1,522	1,404
SASK-TEPH	212 - 213	tephra	-	-	10,257 - 10,056	-
SUERC-67383*	239.5 - 240.5	plant fibres	5,210 ± 37	-18.0	5,905 – 6,019	5,963
SUERC-67382*	269.5 - 271	plant fibres	10,881 ± 44	-16.6	12,696 – 12,818	12,750
VEDDE-TEPH	292 - 293	tephra	-	-	12,102 - 11,914	-
SUERC-67379*	312 - 314	plant fibres	9,083 ± 39	-26.3	10,184 – 10,294	10,235
SUERC-67381	329 - 331	plant fibres	11,074 ± 44	-28.0	12,807 – 13,061	12,941
SUERC-67380	354.5 – 356	plant fibres	11,988 ± 48	-27.1	13,731 – 14,003	13,838
BETA-447582	402.5-403.5	bulk clay/silt	18380 ± 70	-24.8	22,410 – 22,115	22,310

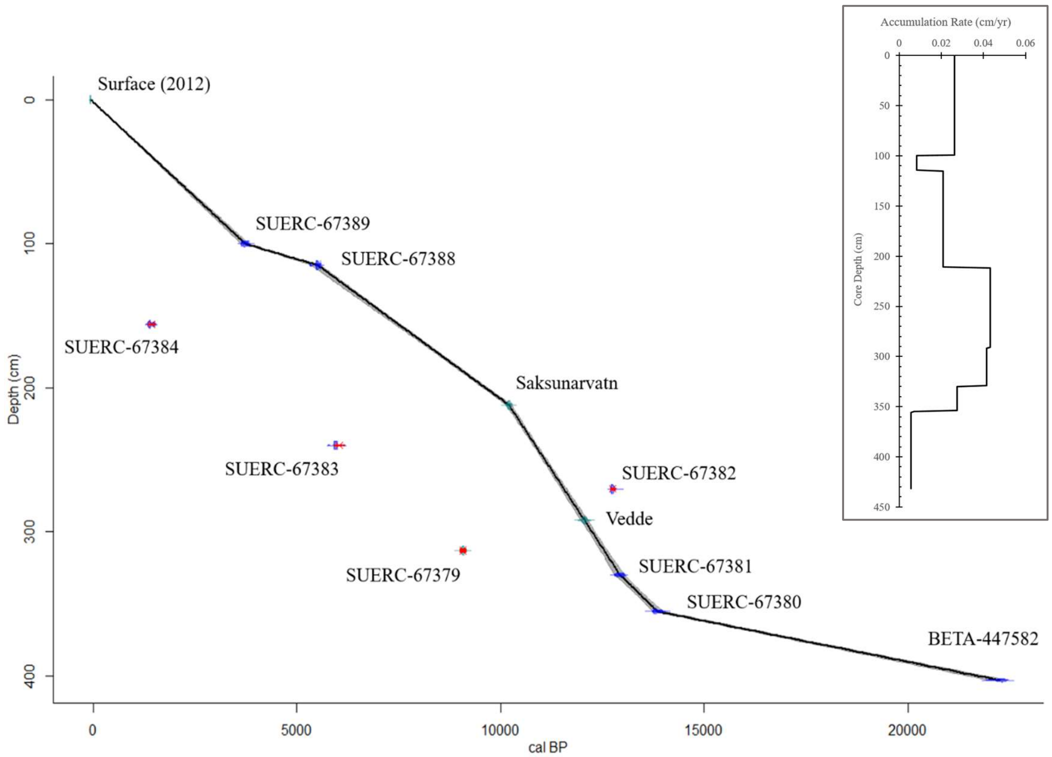

Figure 4.1.8 The selected calibrated AMS ^{14}C dates used along with the dates acquired from the tephra analysis to construct an age-depth model for Loch of Sabiston generated using R v. 3.3 and the package Clam (Blaauw, 2010). Accumulation rate is also presented (inset). The dates designated with blue dots are the ones rejected for the model and have been passively added for illustration purposes.

4.1.9.Diatoms

Seventy-two diatom samples were analysed from Loch of Sabiston, with 163 diatom species identified throughout the record. The diatoms were only sampled from the lake sediments as the acidic nature of peat tends to decrease the preservation of diatoms and thus decrease the reliability of the assemblage which has occurred in the peat layer of the Loch of Sabiston core. Seven distinct temporal zones with two of the zones divided into two sub-zones were identified using CONISS (Grimm, 1987) and optimal zonation (Bennett, 1996) in the diatom assemblage from the end of the basal clay layer and through the extended marl deposition. The primary CONISS break with the highest variance reduction was detected at 270 cm (*c.* 11,510 cal BP) where the sediment changes from bluish-grey clay/silt to marl (Fig. 4.1.9). The break with the second highest level of variance occurred at 114 cm (*c.* 5400 cal BP) that coincides with the change in sediment from the marl to silt accumulation. However, the third highest break is not associated with any change in sediment deposition and is located at 153 cm (*c.* 7340 cal BP) during marl deposition. The fourth highest break occurs at 330 cm (*c.* 12,937 cal BP) again associated with a change in sediment. The subsequent breaks are associated with the changing sediment types that occurred at the end of bluish-grey clay/silt deposition, and within two periods of marl accumulation and not associated with a change in sediment (Fig. 4.1.9). The seven zones and four sub-zones have been labelled SABD-1 to SABD-7 and sub zones as 'a' and 'b' (5a/b and 6a/b) (Fig. 4.1.9) and are described in detail below. The authorities for the diatom species names for all the lochs are listed in Appendix A.

SABD-1 (412 - 355 cm; *c.* 23,828 – 13,840 cal BP): The basal zone is divided into two sub-zones, based on the change of species richness. SABD-1 is dominated by *Staurosira construens, Neidiomorpha biodonis, Staurosirella pinnata, Staurosira venter* and *Pseudostaurosira brevistriata*. The diatom concentration fluctuates in this sub-zone between 6.7 x10^8 and 3.22 x 10^9 valves/g wet sediment (Fig. 4.1.9).

SABD-2 (355-330 cm; 13,840 - *c.* 12940 cal BP): begins at ~ 355 cm marked by an increase in the epiphytic species *Navicula* (*s.l.*), *Denticula tenuis*, *Cymbell*a (*s.l.*), *Brachysira vitrea* and *B. zellensis* along with *Gomphonema angustatum* and the planktonic species *Cyclotella meneghiniana*. *Mastogloia lacustris*, which is a characteristic species of marl lakes (Gaiser et al., 2010), also appears for the first time in this zone. *Fragilaria* (*s.l.*), which was dominant in Zone SABD-1, have drastically decreased along with the diatom concentration in SABD-2 (Fig. 4.1.9).

SABD-3 (330 – 302 cm; *c.* 12,940 – 12,260 cal BP): The beginning of the diatom zone SABD-3 sees a disappearance of *Gomphonema angustatum*, *Cyclotella meneghiniana, Cymbella* (*s.l.*) species and *Brachysira vitera* and *B. zellensis*; *M. lacustris* are also in decline. There is a sharp increase of *Amphora pediculus* to 40 % relative abundance along with an increase in *Denticula tenuis*, *Navicula* (*s.l.*), *Nitzschia* and *Achnanthidium minutissimum*. *Fragilaria* (*s.l.*) also increase returning to an assemblage which is similar in composition to that of SABD-1 (Fig. 4.1.9).

SABD-4 (302 – 270 cm; *c.* 12,260 – 11,510 cal BP): A disappearance of many of the species present in the lower zones occurs in SABD-4. *Fragilaria* (*s.l.*) again dominates (> 80 %) along with *S. pinnata* along with *P. brevistrata., S. venter* which also reaches maximum abundance in this zone.

SABD-5a (270 – 238 cm; *c.* 11,510 – 10,780 cal BP): *Amphora pediculus* decreases to ~ 10 % and *Achnanthes* (*s.l.*) reappear. The diatom concentrations reach their highest value of 6.76 x 10^9 valves/g wet sediment in this zone (Fig. 4.1.9). The start of zone SABD-5a is the point at which the largest CONISS break occurs along with changes in the sediment record. *Cymbellafalsa diluviana* appears for the first time along with a return of *M. lacustris, C.*

meneghiniana and *G. angustatum* but at a higher relative abundance. Other species that returned are characteristic of those found in Zone SABD-2 including *Navicula* taxa, *Denticula tenuis*, *Cymbella* (*s.l.*) species and *Brachysira vitrea* and *B. zellensis* along with *G. angustatum* and the planktonic species *C. meneghiniana*. *Fragilaria* (*s.l.*) are also still present, however they follow a decreasing trend towards the top of the zone. Diatom concentrations start to decrease drastically in this zone and stay low through to the top of the record core (Fig. 4.1.9).

SABD-5b (238 – 190 cm; *c.* 10,7780 – 9120 cal BP): At the start of the zone SABD-5b the diatom assemblages change to salt-tolerant species such as *Epithemia goeppertiana* and subsequently *Fragilaria* (*s.l.*) reaches very low abundances. *Encyonopsis microcephala* reaches its highest abundance of ~ 15 %, along with *B. vitrea* (15 %). *Cyclotella meneghiniana*, found in brackish and freshwater environments, reaches its highest abundance of ~ 35 %. The high abundance of *M. lacustris* (~ 20 %) corresponds with marl deposition (Fig. 4.1.9).

SABD-6a (190 – 153 cm; *c.* 9120 – 7340 cal BP): Zone SABD-6a is driven by the increase in *C. meneghiniana*, a planktonic species, while *M. lacustris* and *Amphora libyca* begin to reach their maximum abundance in this zone (Fig. 4.1.9).

SABD-6b (153 - 114 cm; *c.* 9120 – 5400 cal BP): Within the diatom assemblage of zone SABD-6b, there is an increase of *C. diluviana, P. brevistriata*, and *Amphora pediculus* and a decrease of *Encyonema silesiacum* and *G. angustatum* (Fig. 4.1.9). The diatom assemblages abruptly change to a dominance of *C. diluviana* and *A. pediculus* in zone.

SABD-7 (114 – 90 cm; *c.* 5,400 – 3,350 cal BP): This corresponds with a reduction in the three-dominant species in the previous zones. *M. lacustris*, *C. meneghiniana* and *E. goeppertiana* are drastically reduced or disappeared. This change corresponds to a change in sediment from marl to organic mud, just prior to the shelly layer (Fig. 4.1.9).

The changes in species richness and evenness throughout the core are consistent with the changes observed in sediment that has accumulated (Fig. 4.1.9) with higher diversity in the marl sediment than that found in the clay / silt sediment. At each boundary between zones, there is a change in both species richness and evenness particularly within the transition from the blueish-grey clay/ silt layer to the marl at the base of the core. However, Zone SABD-3 is characterized by highly variable species richness and evenness. At 270 cm (*c.* 11,510 cal BP), there is a sharp rise in species richness which fluctuates between 35 and 40 species, until 190 cm (*c.* 9120 cal BP) where species richness begins to decline to a ~ 20 species. At 120 cm (*c.* 5760 cal BP), species numbers increase again quickly to ~ 40 species followed by a sharp decrease to ~18 species at the point where the loch changes from marl to peat deposition.

The results of the PCA analysis on the diatom assemblage shows that PCA axis-1 explains 38.1 % of the observed variation in species composition (λ = 0.381) related to sample depth (Fig. 4.1.10 a/b). Species positively related to axis-1 such as *S. construens*, *S. pinnata* and *P. brevistriata*, and negatively related species include *M. lacustris*, *G. angustum*, *E. silesiacum* and *C. cymbiformis*. PCA axis-2 explains 11.7 % of species variation (λ = 0.117) and species positively correlated include *C. divulana, A thumensis, A. libyca,* while *C. microcephala, E. smithii* and *B. vitera* are negatively correlated with PCA axis-2 (Fig. 4.1.10a/b).

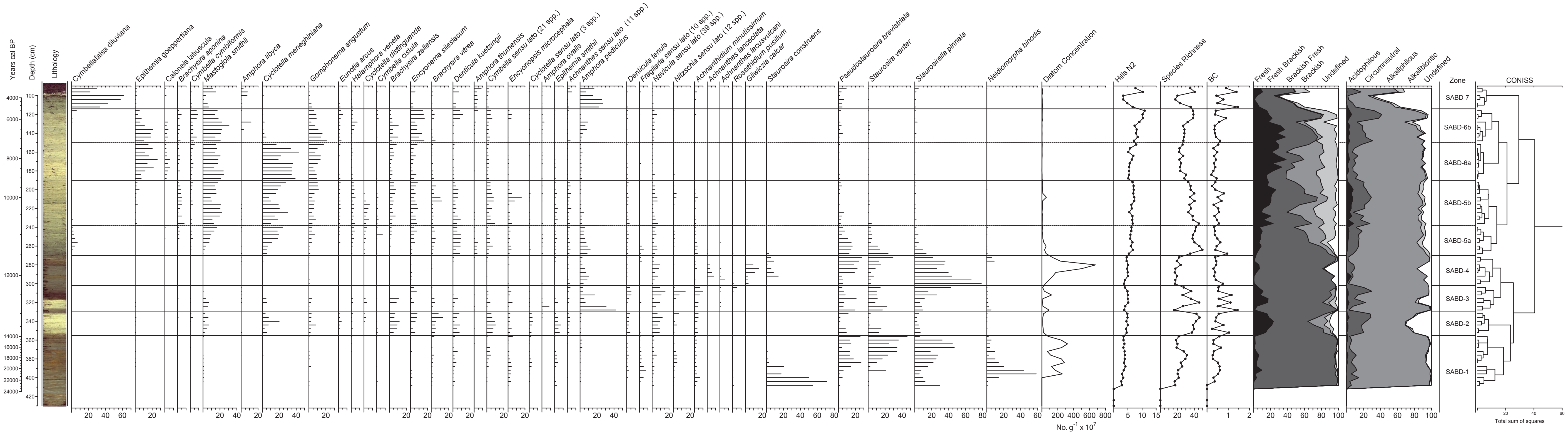

Figure 4.1.9 The dominant (>3 % relative abundance) diatom taxa found in Loch of Sabiston. Species with a relative abundance under 3 % have been grouped based on genus. Solid lines indicate major zones and dashed lines indicate sub-zones determined by constrained cluster analysis (CONISS) (Grimm 1987). The modelled dates appear on the left side by the y-axis along with a lithology of the core. Species evenness, richness and Bray-Curtis similarity are also presented. The diatom assemblage has also been summarized in respect to water pH and salinity of the water (Van Dam et al. 1994)

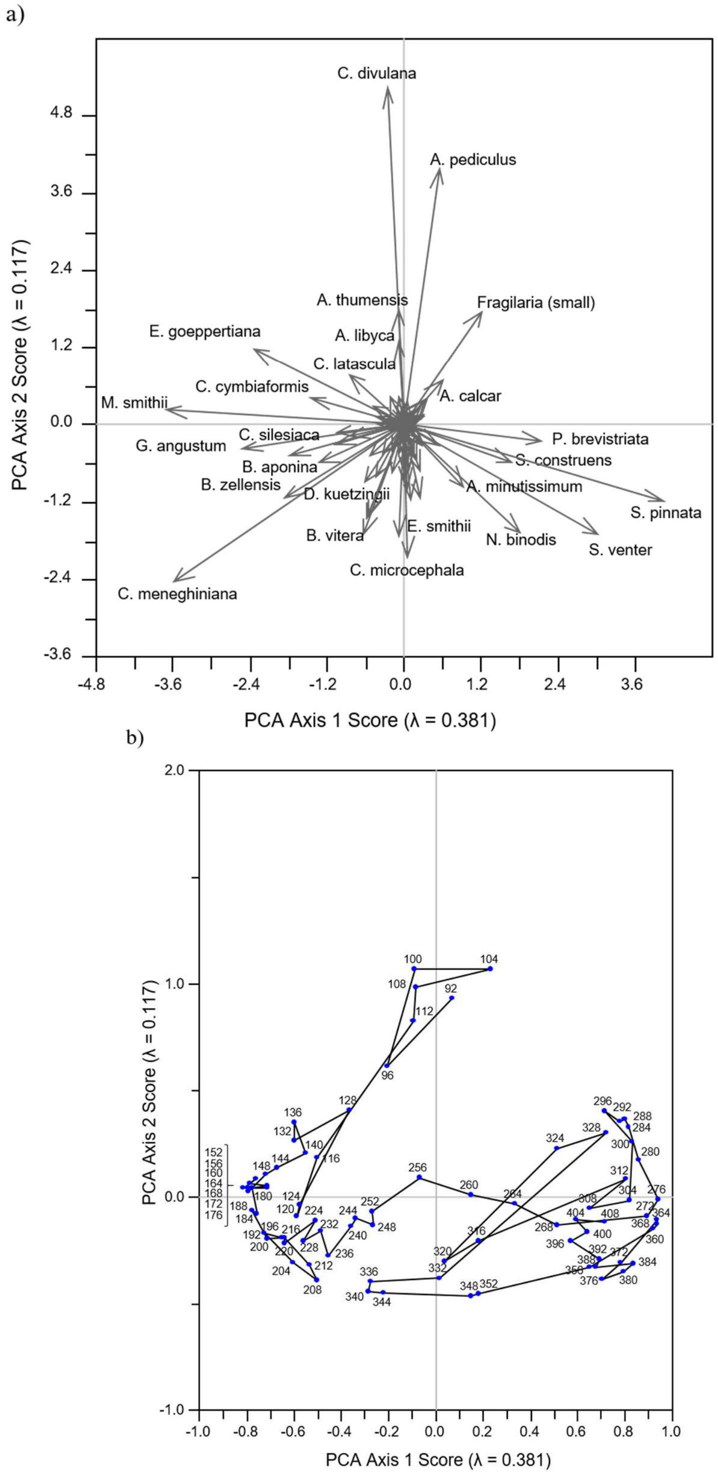

Figure 4.1.10 PCA analysis of the diatom species in relation to sample depth from Loch of Sabiston. PCA axis-1 describes 38.1 % of the observed variation in species composition related to sample depth and PCA Axis 2 explains 11.7 % of species variation. Graph ‘a’ shows the relationship of species to the PCA axis and ‘b’ shows the relationship of the samples to the PCA axis.

4.1.10. Pollen

The core was sampled for pollen from 369 cm to 91 cm at 4-cm intervals for 68 samples. There are several intervals where there was insufficient pollen to count (~ 180 cm, ~ 295 cm and between 335 - 324 cm). The following section outlines the main trends in the pollen assemblage. Five distinct temporal zones and three sub-zones were identified using CONISS (Grimm, 1987) and optimal zonation (Bennett, 1996) in the pollen assemblage from the start of the basal clay layer and through the extended marl deposition. The primary CONISS break with the highest variance reduction was detected at 237 cm (*c.* 10,750 cal BP) where the sediment has changed from silty clay to marl (Fig. 4.1.11). The break with the second highest level of variance occurred at 164 cm (*c.* 7870 cal BP) which is located in the middle of the marl deposition and is not associated with any change in sediment. The third highest break occurs at 270 cm (*c.* 11,514 cal BP), and is associated with a change in sediment from bluish-grey clay to marl. The fourth break is associated with another change in sediment that occurred at the end is in bluish-grey clay/silt deposition and not associated with a change in sediment (Fig. 4.1.11). The five zones and three sub-zones have been labelled SABP-1 to SABP-5 and sub zones as ‘a’ and ‘b’ (1, 2/ab, 3, 4 a/b, and 5a/b) (Fig. 4.1.11) and are described in detail below.

LPAZ SABP-1 (380 – 302 cm *c.* 18,202 – 12,270 cal BP): The basal pollen LAPZ SABP-1 begins at the top of Lateglacial clay sediment and continues to 302 cm. Grasses (65 %) and other herb species such as *Empetrum*, Asteraceae Subfam: Cichorioideae, and *Calluna* dominate the assemblage. *Betula* and *Pinus* are present in the record with 25 % and 15 % respectively. SABP-1 encompasses the change in sediment from clay to marl, with the grasses peaking at the end of the first marl occurrence. Unfortunately, insufficient countable pollen was present within the clay sediments at 333-324 cm (*c.* 13,050 - 12,790 cal BP) (Fig. 4.1.11). The pollen concentrations increase in the subsequent marl period at ~ 320 cm (*c.* 12,700 cal BP).

LPAZ SABP-2 (302 – 270 cm; *c.* 12,270 – 11,510 ca BP): The LPAZ SABP-2 is divided into two sub sections however this is most likely due to the samples that did not have enough pollen to count (Fig.4.1.11). The start of LAPZ SABP-2 is at the same depth as the beginning of the accumulation of bluish-grey clay/silt at 310 cm (*c.* 12,460 cal BP) with only a small portion of samples with sufficient pollen concentrations to be countable. The assemblage during this LAPZ dominated by grasses along with Cyperaceae, *Corylus avellana* type which begins to increase along with an increase in *Pinus* and *Betula*. There is a decrease in Asteraceae Subfam: Cichorioideae to 4 % which reached 20 % by the top of LPAZ SABP-1 (Fig. 4.1.11). At the top of LPAZ SABP-2 the onset of marl sedimentation that marks the beginning of the Holocene. Algae begins to decrease and grasses begin to increase at 280 cm (*c.* 11,750 cal BP), prior to the

LPAZ boundary which occurs at 270 cm (*c.* 11,510 cal BP) (Fig. 4.1.11), but at the same time as changes occurring in the sediment.

LPAZ SABP-3 (270 – 237 cm; *c.* 11,510 – 10,750 cal BP): The lower boundary of LAPZ SABP-3 occurs ~10 cm (~ 230 years) after the change in sediment deposition from bluish-grey clay/silt to marl. There is a marked increase in *Betula* (from 17 to 33 %) and a continuing increase in *Pinus* and *Empetrum* with grasses still dominant (~ 30 %) (Fig. 4.1.11).

LPAZ SABP-4a (237 – 213 cm; *c.* 10,750 - 10,200 cal BP): By the beginning of LAPZ SABP-4a, grasses still dominate with *Corylus avellana* type increasing from ~ 5 % to 30 %. *Betula* and *Pinus* continue to be the dominant tree groups while *Filipendula*, which started to slightly increase in the previous zone to ~ 10 %, is stable though this LPAZ at ~ 6 %. *Empetrum* disappears by the top of this LPAZ while *Calluna* increases into SABP-4b.

LPAZ SABP-4b (213 - 164 cm; *c.* 10,200 - 7870 cal BP): *Calluna* continues to slightly increase to ~ 10% while there is a slight increase in *Corylus avellana* type and decrease in Poaceae which distinguished this LPAZ from SABP-4a. At ~ 180 cm (*c.* 8640 cal BP) near the top of SABP-4b (Fig. 4.1.11) the pollen samples did not contain enough pollen to be countable. However, there is no stratigraphic evidence to suggest any break in the sediment accumulation.

LPAZ SABP-5a (164 - 116 cm; *c.* 7870 – 5560 cal BP): SABP-5 begins at 164 cm and is characterized by an increase of *Corylus avellana* type during the LPAZ along with a corresponding decrease in Poaceae. *Pinus* remains stable though this LPAZ at ~ 11 %, while *Betula* reaches its maximum abundance of 40 % at 155 cm and then begins to decrease to a low of 15 % at 130 cm. By the top of the zone, *Betula* has recovered to ~ 30 % abundance and Polypodiaceae doubled in abundance though the LAPZ (Fig. 4.1.11).

LPAZ SABP-5b (116 – 92 cm; *c.* 5560 – 3430 cal BP): The lower boundary of LPAZ SABP-6 begins at 110 cm, just after a change in sediment from marl to organic mud-silt composition. There is an increase in Poaceae at 100 cm (*c.* 3730 cal BP) along with a decrease in *Betula* and *Corylus avellana* type pollen. At the same time charcoal is present in the pollen record and Polypodiaceae reaches its maximum abundance of 44 % (Fig. 4.1.11).

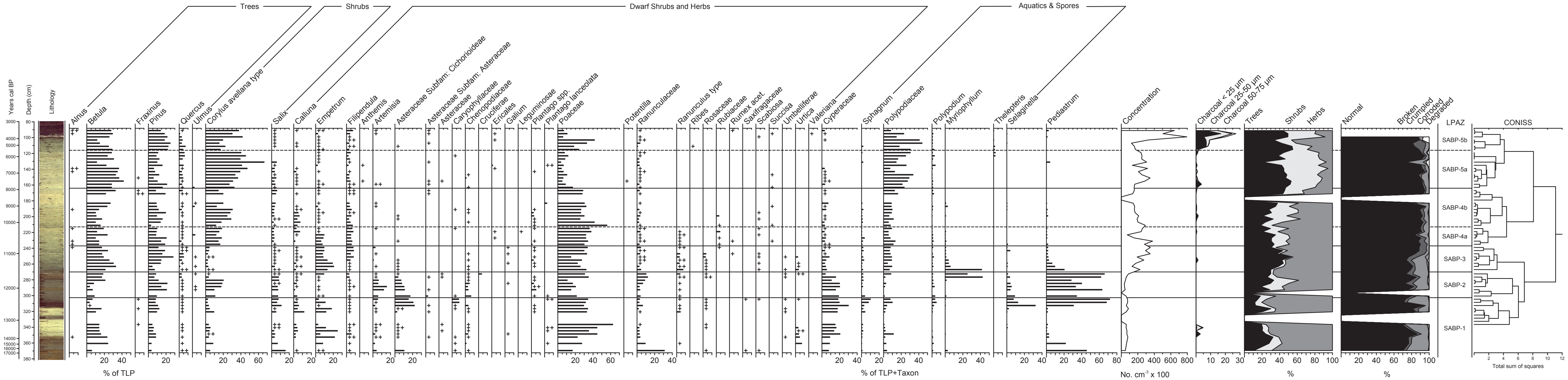

Figure 4.1.11 Percentage pollen and spore diagram of the dominant taxa found in Loch of Sabiston. Taxa under 1% are designated by '+' and under 2% by '++'. Solid lines indicate the major zones and dashed lines indicate the sub-zones determined by constrained cluster analysis (CONISS) (Grimm 1987). The modelled dates appear on the left side by the y-axis along with a lithology of the core.

4.2. Loch of Clumlie Results

4.2.1.Study Area

Loch of Clumlie is found in the southern part of Mainland Shetland on the eastern coast (59° 56’21.5”, 001° 16’ 35.9”), at 23 m a.s.l. (Fig. 4.2.1). It lies in a low area surrounded by hills to the northwest that reach a maximum height of 256 m a.s.l. (Fig. 4.2.2). The loch was shallow and slightly alkaline at the time of sampling (full characteristics listed in Table 4.2.1). A strand line of vegetation debris circling the loch several metres above the waterline demonstrates intermittent high-water events. There are old 19th century mills located along the outlet, which shows the history of the area and the loch’s use for grain processing. A combination of heather, improved grassland, and rough, low productivity grassland dominate the present-day land cover (Figs 4.2.2 and 4.2.3).

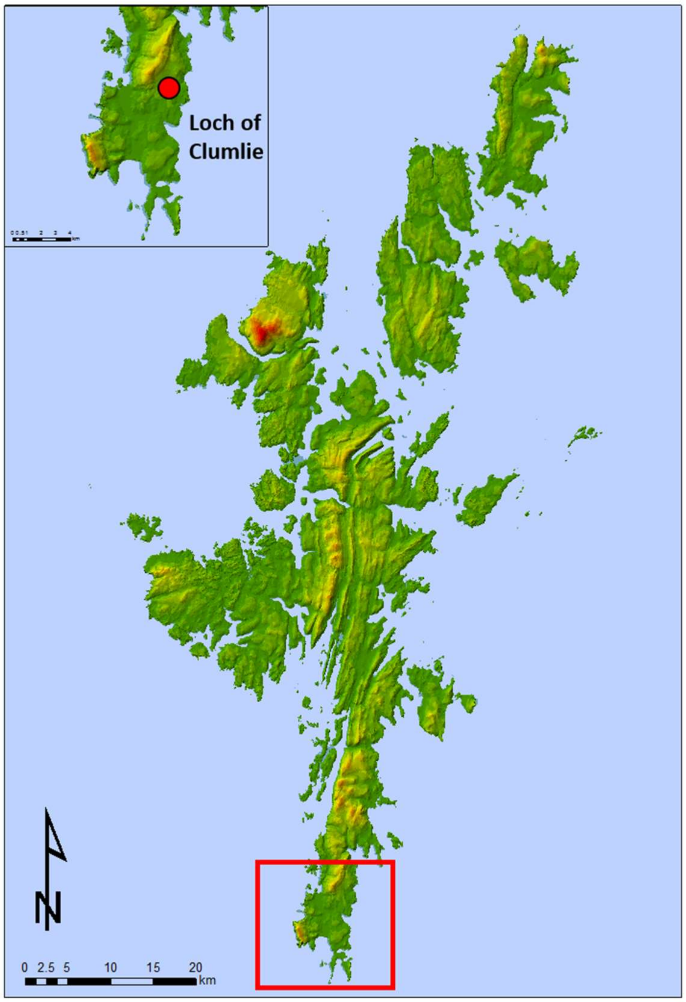

Figure 4.2.1 Shetland with the location of Loch of Clumlie highlighted and the location of the loch (inset).

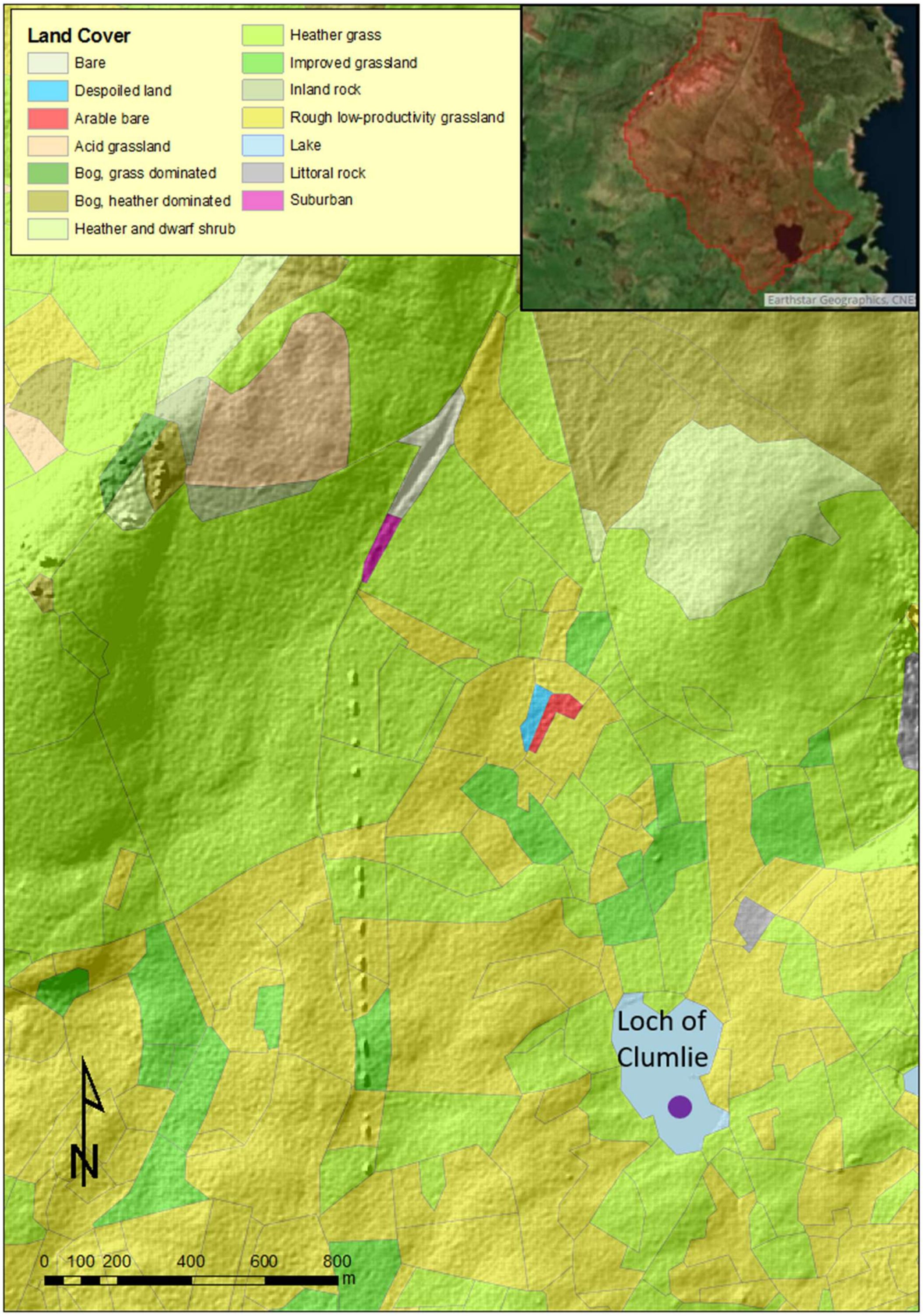

Figure 4.2.2 Present land cover (EDINA, 2007) and topography (NEXTMAP Britain digital elevation data (Intermap Technologies)) in the Loch of Clumlie catchment. The purple point designates the coring location in Loch of Clumlie. The catchment area for Loch of Clumlie (Hughes et al., 2004) is shown in the inset.

Table 4.2.1 Present day characteristics of the Loch of Clumlie at the time of sampling.

Depth (max)	1.04 m
Area	14.5 ha
Temperature	15.4 °C
Conductivity	305 µS
Altitude	23 m a.s.l.
pH	8.15

Figure 4.2.3 Loch of Clumlie on the day of sampling (September 3, 2014).

4.2.2.Sediment Profile

The loch was cored to an impenetrable layer and a 200-cm core was retrieved including ~ 20-cm long gravity core to sample the top surface sediment-water interface. The composition of the sediment is outlined in Table 4.2.2 and illustrated in Figure 4.2.4.

Table 4.2.2 General characteristics of the sediment make up from Loch of Clumlie.

Depth (cm)	**Properties**
0-35	Organic mud
35-52	Iron pan layer with organic mud
52	Coarse sediment/ sand layer
52 – 80	Iron pan layer with organic silt
80 – 119	Bluish-grey clay
119 – 128	Silty organic clays with coarse sediment/ sand layers at 119 and 121 cm
128-130	Coarse sediment/ sand layer
130-137	Silty organic clays with coarse sediment/sand layers at 132 cm
137-141	Bluish-grey clay
141-180	Silty clay
180-200	Bluish-grey clay

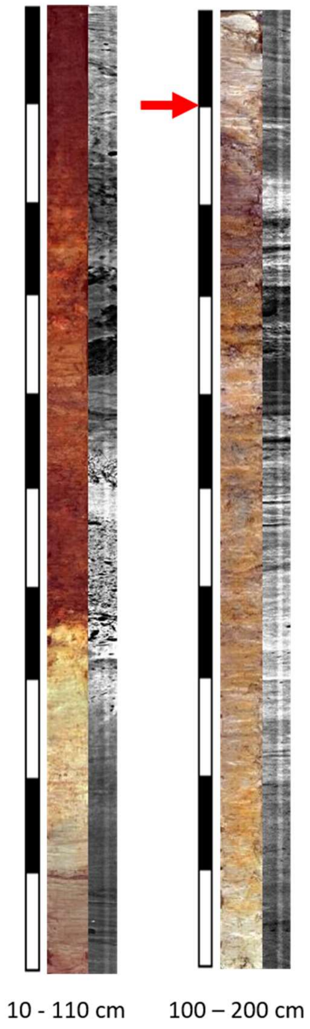

Figure 4.2.4 The 1-m core sections retrieved from Loch of Clumlie. The scale on the left indicates 1-m sections with 10 cm intervals. The red arrow designates the location of the 10-cm overlap. The grayscale picture to the left of the core image is an x-radiograph of each section illustrating the different densities of the sediment. The pictures have been enhanced to enable the detection of variations in the sediment and the colour of the core does not represent the original colour of the sediment at the time of coring.

There is evidence that the basal sediment is Lateglacial in origin due to the presence bluish-grey clay from 200 cm to 180 cm. At 180 cm, the sediment becomes more silty and organic content increases. This lasts until 141 cm at which point bluish-grey clay covers the silty clay which lasts until 137 cm. The sediment accumulation returns to a silty organic layer from 137 - 130 cm and is intersected with a coarse sediment layer at 132 cm. Another coarse sediment layer consisting of silt and sand is deposited from 130 - 128 cm. (Table 4.2.2). Between 128 and 119 cm silty organic clays with coarse sediment sand layers at 119 and 121 cm make up this part of the core. There is a return to bluish-grey clays between 119 - 80 cm core depth. At this point the organic content has increased in the sediment with a change to organic silt with evidence of oxidation and iron pan which lasts until 52 cm where another coarse sediment sand layer is deposited. The oxidation continues and the sediment consists of an organic mud until 35 cm where the oxidation disappears and there is organic mud until the top of the core.

4.2.3. Loss on Ignition (550)

The organic content of the sediment follows closely the changes in sediment composition. (Fig. 4.2.5). At the base of the core, organic content begins at ~ 4 % and reaches 5 % at the transition to a siltier sediment at 177 cm. During the accumulation of the silty-clay, organic content continues to rise to ~ 10 % at 159 cm. Then organic content gradually decreases to ~ 3 % at 133 cm. There is a step-wise increase in organic content with peaks at 127 cm (8 %) and 120 cm (11 %). From 120 - 78 cm, there is a consistent decrease of organic content to a low of 1.5 %. From 78 - 55 cm, another rise and fall in organic content occurs with 18 % organic content at 66cm. From 55 cm to the top of the core, the organic content is highly variable and fluctuates from 5 - 18 %, peaking at 34 % at the top.

4.2.4. Magnetic Susceptibility

Magnetic susceptibly (MS) was measured continuously every 2 mm for the full length of the core (Fig. 4.2.5). High values of magnetic susceptibility were observed at the bottom of the core (200 - 180 cm) and remains stable until the occurrence of peaks at 42, 90, 95, 100, 110 and 124 cm. These peaks were investigated for the evidence of tephra layers, however, these locations in the core did not reveal any tephra shards. The peaks in MS are more likely associated with minerogenic in-wash of sediment.

4.2.5. Micro-XRF Results

Twenty-five elements were selected during the µ-XRF process with an average mean squared error (MSE of 3.5). Overall, the highest positive correlations (Table 4.2.3) occur between elements associated with minerogenic inputs (K, Ti, Rb, Zr, and Sr, Fig. 4.2.5) while manganese is found to be negatively correlated with potassium and titanium. The relative changes in elements throughout the core correlate with changes in lithology (Fig. 4.2.5). There are several occurrences in the core where these minerogenic elements increase. These take place at 140 cm, from 120 to

80 cm and from 55 to 30 cm. Silica also follows this trend and to a lesser extent iron (Fig. 4.2.5). Bromine is slightly negatively correlated with K, Ti, Rb and Si.

Table 4.2.3 Correlation matrix of select elements analysed using μ-XRF. Correlations with a probability greater than 0.5 are highlighted in bold, negative correlations are also underlined.

	Si	K	Ca	Ti	Rb
K	**0.716**				
Ca	0.456	**0.589**			
Ti	**0.633**	**0.866**	**0.616**		
Mn	-0.484	<u>**-0.565**</u>	-0.228	<u>**-0.584**</u>	
Rb	**0.654**	**0.930**	**0.527**	**0.797**	
Sr	**0.667**	**0.730**	0.421	**0.687**	**0.728**
Fe	-0.389	-0.191	0.023	-0.108	-0.288
Br	-0.313	-0.378	-0.152	-0.363	-0.330
Zr	0.367	0.373	0.210	0.446	**0.405**

In Loch of Clumlie, the relationship between Ca/Ti, Zr/K, and Rb/K show fluctuations from 160 cm to 120 cm with a decrease from 142 - 136 cm. Between 120 cm and 80 cm values remain stable until a maximum peak is reached ~ 70 cm (Fig. 4.2.6). There is high variability in these elemental relationships at the top of the core (from ~ 60 cm onwards). The Fe/Si ratio follows the same pattern as the one described above, however, the changes observed from 160 - 120 cm are subtle and less variable. The relationship between Si and Ti shows stability at the base of the core until ~ 136 cm where values peak (Fig. 4.2.6). Between 120 - 80 cm, values slightly increase by 100 cm, but then decrease by ~ 90 cm. By 80 cm, the ratio increases again to a peak at 70 cm. Fe/Mn and Mn/Ti are showing a different pattern to those observed above. The Fe/Mn shows a steady decline from the base of the core with high variability. Between ~ 90 cm and 75 cm, it begins to rapidly decrease to virtually zero which persists for until 55 cm (Fig. 4.2.6). Mn/Ti mirrors this pattern with low values for most the core, and then reaches its highest value ~ 75 cm. The peaks in Mn/Ti mirror the lows in the Fe/Mn values (Fig. 4.2.6).

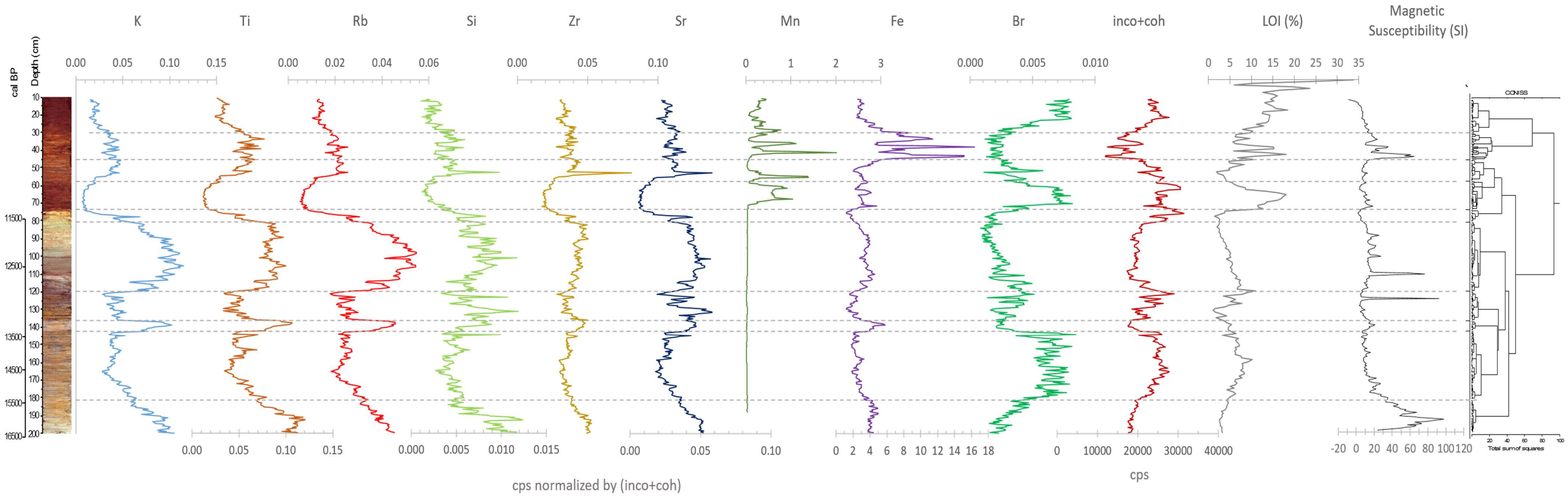

Figure 4.2.5 Lithology, organic matter content (% organics based on loss-on-ignition at 550°C (LOI$_{550}$)), magnetic susceptibility and selected µ-XRF geochemistry from Loch of Clumlie. Zones are based on the constrained cluster analysis of the µ-XRF results and elements have been corrected for organic and moisture content by dividing the elemental results by the sum of incoherent and coherent (inco+coh) scatter results (Davies et al. 2015).

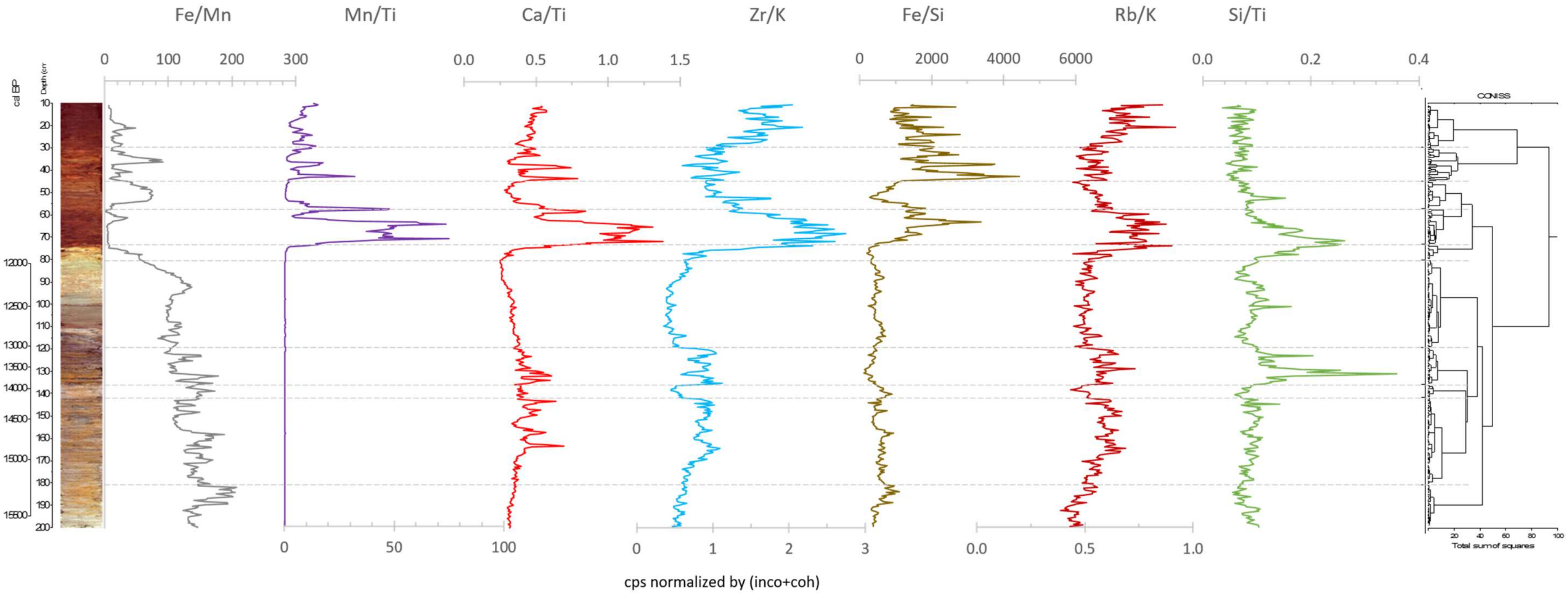

Figure 4.2.6 Selected elemental ratios from µ-XRF scan from Loch of Clumlie. Zones are based on the constrained cluster analysis of the µ-XRF results. Ratios were determined by using elements corrected for organic and moisture content (Davies et al., 2015).

4.2.1. Chronology

Five dates were obtained from the NERC AMS radiocarbon facility at East Kilbride (labelled SUREC) while the Beta sample was analysed at Beta Analytic Inc., Miami, Florida. The results of the analysis are summarized in Table 4.2.4. During sampling for suitable material, it was noted the presence of roots penetrating the lower sediment from the fen-peat layer at the top of the core which is probably the cause of contamination from younger carbon, and therefore, sample SUERC-68213 was omitted from the age depth model as it was out of sequence and believed to be too young. Sample SUREC-68211 was determined to be too old based on the sediment stratigraphy. In the topmost organic portion of the core, there is evidence of drying out, leaching and oxidation of the sediment along with evidence of a hiatus from 78, cm to the top of the core. Due to the nature of the sediment, the chronology will only incorporate the Lateglacial period below 78 cm.

These dates were then used to construct an age-depth model using R (R Development Core Team, 2016) and the package Clam (Blaauw, 2010). Based on evidence from the pollen, diatoms, and geochemistry results, the age-depth model constructed was determined to be too old as the start of the IACP has been dated to *c.* 13,311 cal BP while the date from the Loch of Clumlie age-depth model was c. 14,210 cal BP) (Fig. 4.2.7). To compensate for this, three dates from the Greenland ice core records (Rasmussen et al., 2014) were supplemented into the model along with the basal radiocarbon age from Loch of Clumlie (SUERC-68210) and sample SUREC-68212. The Greenland dates define the beginning of the Holocene at 80 cm (11,703 cal BP ± 4), the start of the Younger Dryas at 120 cm (12,896 cal BP ± 4), and the start of the IACP at 142 cm (*c.* 13,311 cal BP). The depths at which to place these sample were determined by the results of the μ-XRF analysis whereby a change in sediment composition determined these points. BETA-447119 was omitted as its variation was such that it caused an age reversal error when running the model. The resulting age-depth model (Fig. 4.2.7) is in good agreement with the lithology of the core. However, the chronology will be used with caution during interpretation as the assumption has been made that these events on Shetland and Greenland are synchronous by using the NGRIP dates.

A search of tephra layers was undertaken based on the results of the magnetic susceptibility and μ-XRF elemental scan using elements that have been associated with tephra (Mn, Fe, TI, Zr, K, Co/Ni) (Davies et al., 2015). However, no tephra shards have been identified.

Table 4.2.4 Radiocarbon dates submitted to NERC / SUERC and Beta, calibrated using CALIB 7.1 (Stuiver et al., 2013) and the IntCal13 atmospheric calibration curve (Reimer, 2013). The samples marked with an * were used in the development of the age-depth model due to age reversal.

Sample ID	Depth (cm)	Material	14C age ± error	δ13C^{VPDB} ‰ ±0.5	Calib cal age BP (2σ)	Calib wmean cal BP
SUERC-68210	177–179*	plant fibres	12,784 ± 49	-27.9	15,071 – 15,431	15,230
SUERC-68211	141.5–142.5	plant fibres	12,287 ± 47	-24.9	14,034 – 14,520	14,210
SUERC-68212	118.5–119.5	plant fibres	10,982 ± 44	-26.3	12,727 – 12,981	12,832
BETA-447119	116–117*	plant fibres	11,030 ± 30	-26.5	12,995 – 12,795	12,855
SUERC-68213	89.5–91	plant fibres	4,700 ± 35	-21.4	5,320 – 5,580	5,406

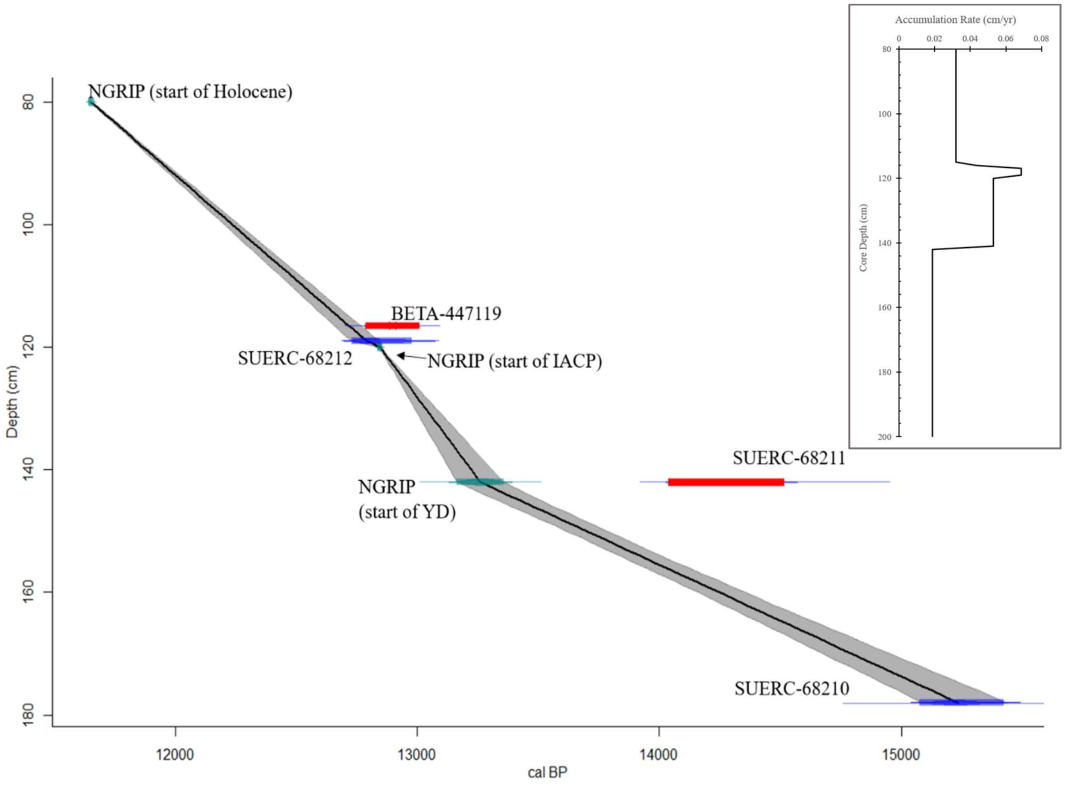

Figure 4.2.7 The selected calibrated AMS ^{14}C dates used along with the dates from Rasmussen et al. (2014) to construct an age-depth model for Loch of Clumlie generated using R v. 3.3 and the package Clam (Blaauw, 2010). Accumulation rate is also presented (inset). The depth chosen for the Greenland dates was based on the change in sediment composition from the μ-XRF data. The red dates are the ones rejected for the model.

4.2.2.Diatoms

Ninety-two samples were counted from Loch of Clumlie, with 178 diatom species identified. Eight distinct temporal zones were identified by CONISS (Grimm, 1987) and optimal zonation (Bennett, 1996) in the diatom assemblage. The primary CONISS break with the highest variance reduction was detected at 142 cm (*c.* 13,310 cal BP) which occurs just before a change in sediment from silty to bluish-grey clay (Fig. 4.2.8). The break with the second highest level of variance occurred at 90 cm (*c.* 12,020 cal BP) and the third highest break was located at 119 cm (*c.* 12,882 cal BP). The eight significant zones have been labelled CLMD-1 to CLMD-7b (Fig. 4.2.8) and are described in detail below.

CLMD-1 (200 - 166 cm; *c.* 16,400 – 14,590 cal BP): This basal zone of the core incorporates the transition from glacial bluish-grey clay to a more organic silty-clay (Fig. 4.2.4) and is characteristic of a Lateglacial diatom assemblage with a dominance of *Staurosira construens* and *Staurosirella pinnata*. However, despite the dominance of *S. construens* and *S. pinnata*, the overall species richness begins at 20 species and then steadily climbs throughout the zone reaching a maximum abundance of 45 species (Fig. 4.2.8) indicating an established community including *Nitzschia* (3 species), *Achnanthes suchlandtii* (which reaches its highest abundance in this zone), *Rossithidium pusillum, Staurosira venter*, *Pseudostaurosira brevistriata*, in quantities under 10 %. The similarity between the samples is stable within the zone (Fig. 4.2.8), while evenness steadily increases along with species richness (Fig. 4.2.8). By the top of the zone (167 cm; *c.* 14,650 cal BP) *F. construens* has decreased from over 60 % of the diatom community to around 20 %. Diatom concentrations start quite low and then fluctuate in a series of peaks at 169 cm (*c.* 14,750 cal BP), 177 cm (*c.* 15,180 cal BP) and 187 cm (*c.* 15,710 cal BP) (Fig. 4.2.8).

CLMD-2 (166 - 142 cm; *c.* 14,590 – 13,310 cal BP): Zone CLMD-2 continues through the silty-clay sediment and its top boundary is at 142 cm, ~ 4 cm below the transition back to bluish-grey clay. Across the transition from CLMD-1 to CLMD-2 *S. construens* proportionally continue to decrease while *S. pinnata* increased by 15 % along with an increase in *P. brevistriata* and a slight increase of *Fragilariforma exigua*. *Achnanthes lanceolata* also appears at the end of CLMD-1 and continues through CLMD-2 (Fig. 4.2.8). Diatom concentrations increase and are again variable in this zone with peaks at 160 cm (*c.* 14,270 cal BP), 155 cm (*c.* 14,000 cal BP), and 143 cm (*c.* 13,360 cal BP). Hills N2 shows a lower value at the beginning of the zone but then increases to 10 by the top of the zone. This trend is also found in species richness with an increase occurring at the top of the zone, although the similarity between the samples remains quite consistent (Fig. 4.2.8).

CLMD-3 (142 - 119 cm; *c.* 13,310 – 12,880 cal BP): The boundary between CLMD-2 and CLMD-3 at 142 cm has the highest level of variance in the diatom assemblage (as determined by CONISS) marked by the decrease in *S. construens* and *P. brevistriata* and the rapid increase of *F. exigua,* along with an increase in species diversity (Fig. 4.2.8), with the appearance of *Navicula* (*s.l.)* and *Pinnularia* species. Diatom concentrations are lower at the start of CLMD-3 as the sediment changes to bluish-grey clay and organic content decreases but then increases along with diatom concentrations through the zone. Evenness begins to fluctuate in a series of peaks during this zone, from a low index of 2 to a high of 21, while there is a decrease in species richness around 135 cm (*c.* 13,180 cal BP) from 48 to 30, followed by an increase to 47 again by the top of the zone. Bray-Curtis reflects this dynamic by increasing in this zone and then fluctuating between 0.61 and 1.75 indicating that adjacent samples are less similar to each other (Fig. 4.2.8).

CLMD-4 (119 - 98 cm; *c.* 12,880 – 12,270 cal BP): Low diatom concentrations are found throughout this zone (Fig. 4.2.8). *Navicula* (*s.l.*) and *Pinnularia* species reach their maximum proportions in this zone along with *Fragilaria capucina*, *S. construens* almost disappears in this zone and *S. pinnata* drops below 5 % (Fig. 4.2.8). Species richness also decreases in this zone along with Hills N2 which corresponds with the accumulation of bluish-grey clay. Bray-Curtis also indicates variability at the beginning of the zone, however, the samples are more similar in the upper half of the zone.

CLMD-5 (98 – 90 cm; *c.* 12,270 – 12,020 cal BP): There is a reduction in species diversity along with a sharp increase of *F. exigua*, which dominates the diatom assemblage at the transition from CLMD-4 to CLMD-5 (Fig. 4.2.8). CLMD-5 is a very short zone (10 cm) characterized by a rapid change in species assemblages and Hills N2. The cold tolerant species (e.g. *A. helveticum, Naviculadicta digitulus, Navicula molestiformis. Pinnularia biceps*) that were dominant in zone CLMD-4 now have declined.

CLMD-6 (90 - 80 cm; *c.* 12,020 – 11,700 cal BP): The boundary between CLMD-5 to CLMD-6 was the second highest level of variance indicated by CONISS. The zone is characterised by a fast turnover of species with a rapid decrease in *F. exigua* and a return of *S. pinnata*, *P. brevistriata* and *S. construens* (Fig. 4.2.8). Species richness also decreased with the disappearance of *A. helveticum, N. digitulus, N. molestiformis,* and *Pinnularia biceps* from the diatom community.

CLMD-7a (80 – 24 cm; *c.* 11700 cal BP >): The top 80 cm of the core has two depths that did not have adequate diatoms to count at ~ 55 cm and from 32 to 47 cm. There was evidence of dissolution of the diatom content probably due to a drying out of the loch resulting in a hiatus. There is a peak in concentration values at the transition between CLMD-6 and CLMD-7a but a drastic reduction at 60 cm follows which continues to the top of the core (Fig. 4.2.8). Although there is an increase in species diversity in this zone dominated by *Fragilaria* (*s.l.*) species, it is difficult to comment on the diatom diversity and similarity near the top of the core during the probable hiatus.

CLMD-7b (24 - 0 cm): The topmost zone CLMD-7b consists of the top 15 cm of the core. However, this assemblage is analogous to the CLMD-7a and is likely affected by a hiatus or multiple hiatuses. The lack of diatoms present limits the effectiveness of the cluster analysis in this section of the core (Fig. 4.2.8).

PCA axis-1 describes 36.7 % of the observed variation in species composition ($\lambda = 0.367$) related to sample depth and PCA axis-2 explains 16.6 % ($\lambda = 0.166$) of species variation (Fig. 4.2.9). Axis-1 is positively correlated to species associated with small *Navicula* (*s.l.*) and is negatively correlated to pioneer diatom species such as *S. construens*, *F, constricta*, and *A. pediculus* (Fig. 4.2.9). PCA axis-2 is negatively correlated with diatoms species such as *Pseudostaurosira. elliptica, R. pusillum*, A. *minutissimum*, and *P. brevistriata* (Fig. 4.2.9).

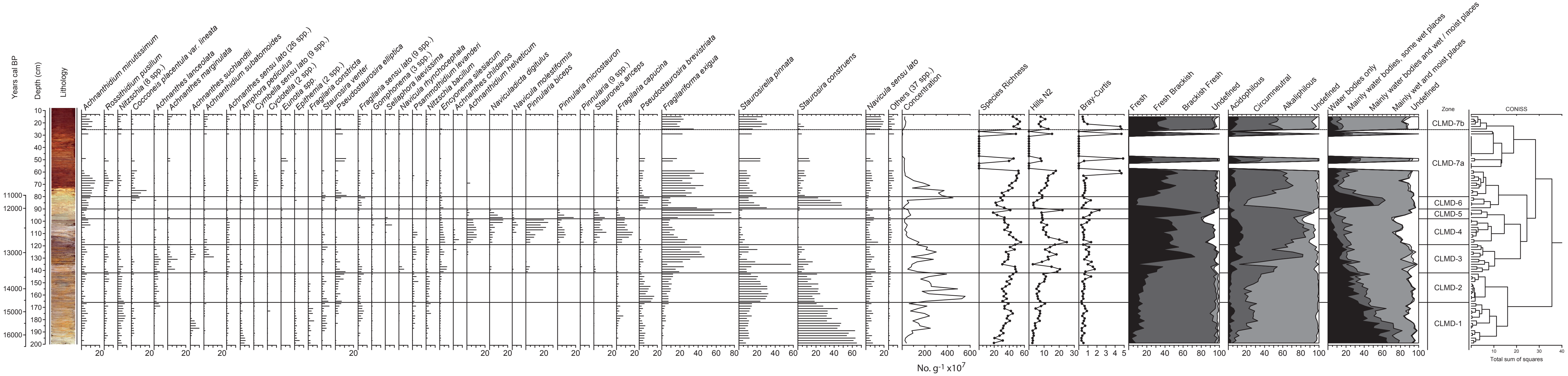

Figure 4.2.8 The dominant (>3 %) diatom taxa found in Loch of Clumlie. Species with a relative abundance below 3% were grouped based on genus. The record has been divided up into zones based on a constrained cluster analysis (CONISS) (Grimm 1987). Species evenness, richness and Bray-Curtis similarity are presented. The diatom assemblage has also been summarized in respect to water pH and salinity of the water (Van Dam et al. 1994).

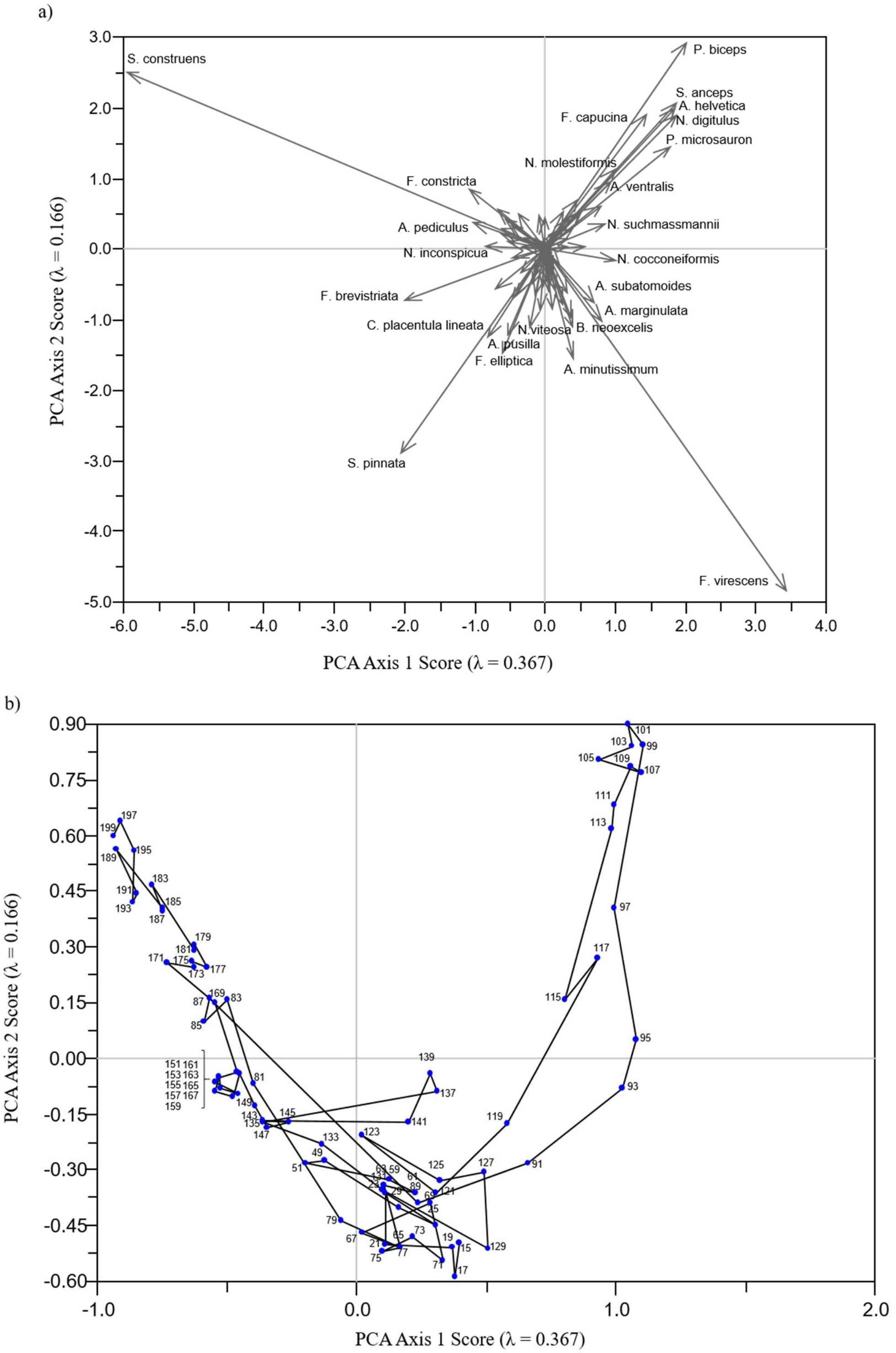

Figure 4.2.9 PCA analysis of the diatom species from Loch of Clumlie in relation to sample depth. PCA axis-1 describes ~ 36.7 % and PCA axis-2 explains ~ 16.6 % of the observed variation in species composition related to sample depth. Graph 'a' shows the relationship of species to the PCA axis and 'b' shows the relationship of the samples to the PCA axis.

4.2.3.Pollen

The core was sampled for pollen analysis at 4-cm intervals (1-cm^3 samples) for 50 samples along the full length of the core. Five significant local pollen assemblage zones (LPAZ) were identified by CONISS (Grimm, 1987) and optimal zonation (Bennett, 1996). The primary CONISS break with the highest variance reduction was detected at 52 cm (Fig. 4.2.10). The break with the second highest level of variance occurred at 78 cm (*c.* 11,640 cal BP). The third highest break occurs at 122 cm (*c.* 12,930 cal BP), while the fourth occurs at 166 cm (*c.* 14,590 cal BP) (Fig. 4.2.10). The zones have been labelled CLMP-1 to CLMP-5.

LPAZ CLMP-1 (180 - 166 cm; *c.* 15,340 – 14,590 cal BP): The beginning of the pollen record occurs at 180 cm (*c.*15,340 cal BP) and coincides with the beginning of silty-clay sediment. The early record is dominated by Poaceae*,* *Ranunculus* type herb, *Rosaceae* and Cyperaceae (sedges). *Salix* which begins at ~ 5 % at the beginning of the record steadily increases to the top of the zone (Fig. 4.2.10). The algae *Pediastrum* was also abundant during this time in the loch.

LPAZ CLMP-2 (166 - 122 cm; *c.* 14,590 – 12,93 cal BP): The LPAZ CLMP-1 assemblage continues into LPAZ CLMP-2 where the percentage of herb species decline corresponding with the increase in trees and shrubs. However, Poaceae remains the dominant taxon at ~ 50 % with lesser proportions of *Salix* ~ 20 %. *Pediastrum* remains dominant in the aquatic environment until ~ 140 cm when it decreases to ~ 10 % which corresponds to a change in the sediment to silty clay and a decrease in organic content (Fig. 4.2.10). *Pediastrum* subsequently increases slightly at ~ 130 cm to 20 % however, steadily declines until the top of the LPAZ.

LPAZ CLMP-3 (122 - 78 cm; *c.* 12,930 – 11,640 cal BP): During LPAZ CLMP-3 Cyperaceae increases to ~ 40 % and *Salix* reaches its maximum abundance of 35 % (Fig. 4.2.10). However, the proportion of *Salix* decreases to under 5 % by the top of the LPAZ. Poaceae slightly decreases in the middle of the LPAZ while *Salix* reaches its maximum abundance. Poaceae increases again to reach its maximum abundance (~ 60 %) at the top of the LPAZ while *Salix* decreases (Fig. 4.2.10). The percentage of trees and shrubs in this zone remain quite low with *Corylus avellana* type having the highest dominance in this groups at ~ 5 %.

LPAZ CLMP-4 (78 – 52 cm; 11,640 cal BP >): In this LPAZ, Poaceae dominates this zone comprising of over 60% of the TLP. *Salix,* which consisted of almost 40% of TLP has decreased to below 5 % in this LPAZ (Fig. 4.2.10). Again, there is minor change in the trees and shrubs (excluding *Salix*) in this LPAZ and their proportions continue to be below ~ 25 %. Cyperaceae increases in this zone and aquatic pollen and spores reach their maximum abundance in this zone (Fig. 4.2.10).

LPAZ CLMP-5 (52 – 0 cm): At ~ 50 cm where a hiatus has been suggested based on the diatom record in CLMP-5. At this point, the pollen assemblage switches to heath and peat development along with a reduction in aquatic pollen and spores and Cyperaceae. The terrestrial vegetation is now dominated by Poaceae along with *Calluna* and herbs. Trees and shrubs have been reduced to below 5 % of the assemblage (Fig. 4.2.10).

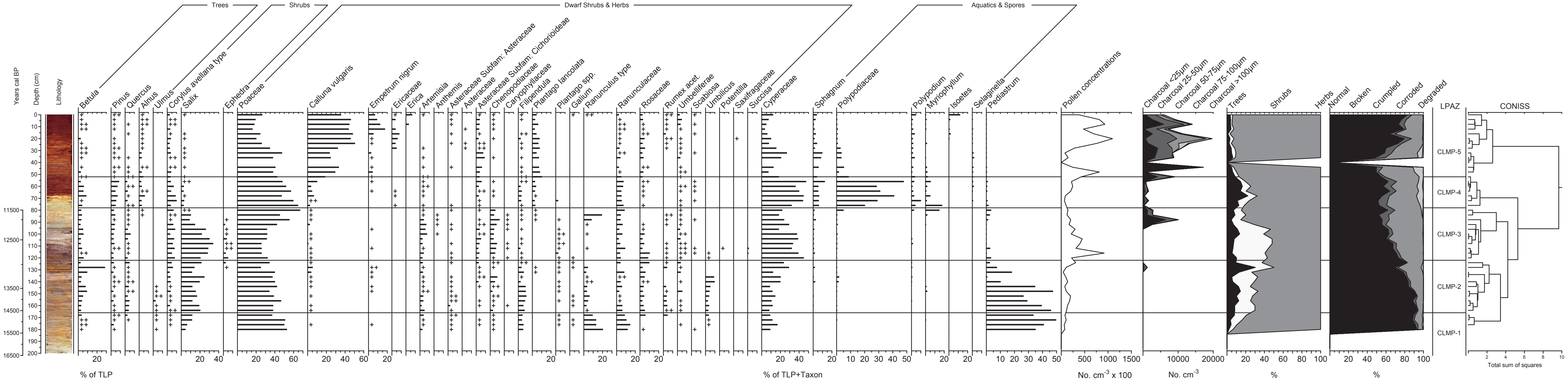

Figure 4.2.10 Percentage pollen and spore diagram of the dominant taxa found in Loch of Clumlie. Taxa under 1% are designated by '+' and under 2% by '++'. Solid lines indicate the major zones and dashed lines indicate the sub-zones determined by constrained cluster analysis (CONISS) (Grimm 1987). The modelled dates appear on the left side by the y-axis along with a lithology of the core.

4.3. Loch of Grimsetter Results

4.3.1. Study Area

Loch of Grimsetter is located on the eastern side of the island of Bressay (60° 08'22.2", 001° 04' 10.9") which lies off the eastern coast of Mainland Shetland opposite the town of Lerwick (Fig.4.3.1). The loch has one inlet to the north-west and one outlet to the south-east and is surrounded by hills with West Hill reaching a height of ~150 m a.s.l. to the south-west (Fig. 4.3.2). There is evidence of a Norse settlement as well as the remains of 18th century crofts in the catchment. To the south, a Bronze Age burnt mound and souterrain are located at Wadbister (Fig. 4.3.2). The present-day characteristics of the Loch are listed in Table 4.3.1 at the time of sampling. The catchment is presently used for sheep grazing (Fig. 4.3.3).

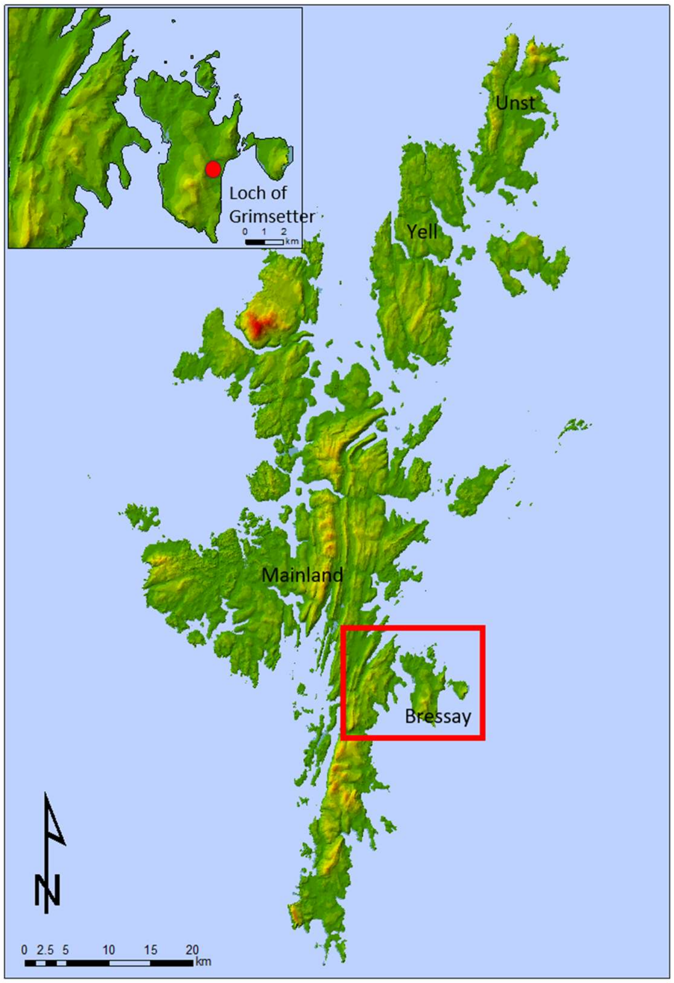

Figure 4.3.1 Shetland highlighting the island of Bressay and the location of Loch of Grimsetter (inset).

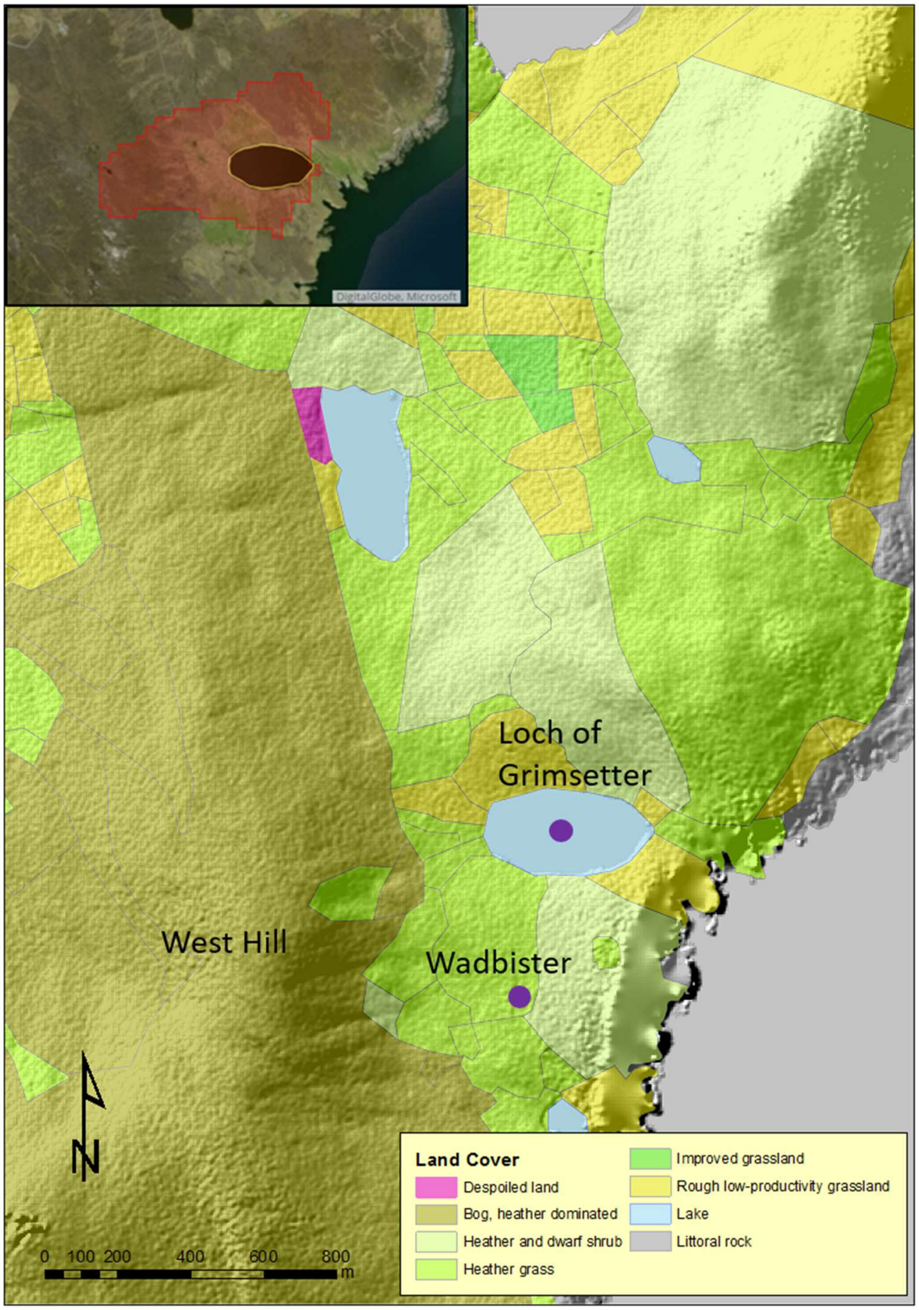

Figure 4.3.2 Present land cover (EDINA, 2007) and topography (NEXTMAP Britain digital elevation data (Intermap Technologies)) around Loch of Grimsetter on the island of Bressay. The purple point in the loch designates the coring location. The catchment area (inset) for Loch of Grimsetter (Hughes et al., 2004).

Table 4.3.1 Present day characteristics of the Loch of Grimsetter at the time of sampling.

Depth	2.30 m
Area	10.9 ha
Temperature	15 °C
Conductivity	1136 µMcm^{-1}
Altitude	27 m a.s.l.
pH	8.03

Figure 4.3.3 Loch of Grimsetter on the day of sampling (September 1, 2014). Note the location of the raft in the middle of the loch where coring was undertaken.

4.3.2.Sediment Profile

The sediment of the loch consists of soft, organic rich gyttja throughout the core. Due to the soft nature of the sediment, the first ~ 20 cm (from the sediment-water interface into the sediment) was sampled with a gravity corer to reach a depth where the sediment would be more consolidated and could be retrieved successfully with a Russian corer to an impenetrable base. A total of 568 cm was retrieved (Fig. 4.3.4) with 10 cm overlap of each of the core sections to guarantee a continuous record.

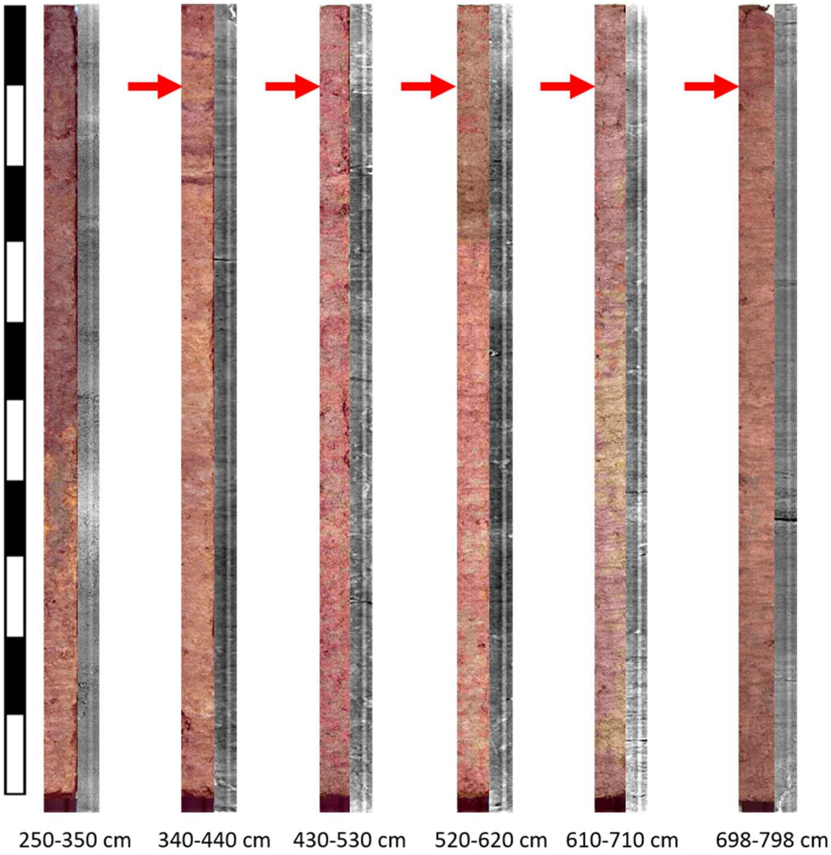

Figure 4.3.4 The 1-m core sections retrieved from Loch of Grimsetter. The scale on the left indicates 1-m section with 10-cm intervals. The red arrows designate the location of the 10-cm overlap. The grayscale picture to the left of the core image is an x-radiograph of each section illustrating the different densities of the sediment. The pictures have been enhanced to enable the detection of variations in the sediment and the colour of the core does not necessarily represent the original colour of the sediment at the time of coring.

4.3.3.Loss on Ignition (550)

The results of the LOI_{550} show that the organic content of Loch of Grimsetter gradually increased from the bottom of the core to the top from 20 % to 40 % (Fig. 4.3.5). At the bottom of the core, organic content starts around 22 % and then has a sharp decrease at 545 cm to ~ 12 % and then returns to previous values at 520 cm. There are other decreases in organic content occurring at 475 cm and 430 cm. At 320 cm, a peak is present with an increase in organic content of 28 %, which stays consistent until 130 cm. Organic content gradually increases until 70 cm to

~ 32 %. At 70 cm, organic content increases again reaching a maximum of 42 % at the top of the core.

4.3.4. Magnetic Susceptibility

Magnetic susceptibility was not performed during the XRF scanning process for Loch of Grimsetter. The MS detector rests on the core during the scan which resulted in significantly disturbing the soft sediment and risked destroying the cores.

4.3.5. Micro-XRF Results

The results of the μ-XRF analysis are based on the core taken with the Russian corer with the sediment record beginning at ~ 20 cm below the sediment-water interface. Twenty-five elements were selected during the μ-XRF process with an average mean squared error (MSE) of 3.5. Each element was standardized using the sum of inc + coh (Davies et al., 2015) to account for Compton-Rayleigh scattering which reduces the effect of organic content and moisture on the profiles (Davies et al., 2015). The results of the incoherent (inc) and coherent (coh) scans show that the moisture and organic content remains stable throughout the core. The ratio of inc/coh can be used as a proxy for organic content and is consistent with the results of the LOI analysis (Fig. 4.3.5).

The elemental changes within the sediment from Loch of Grimsetter are subtle, suggesting that the catchment has been relatively stable over time (Fig.4.3.5). There is strong correlation between K, Ti, Fe, Rb, and Sr whereas Si correlates with Ca and Br (Table 4.3.2). At 530 cm, there are peaks in Ti, K, Rb, and Sr with corresponding decreases in LOI and inc/coh. These elements then stabilized throughout the remainder of the core until ~ 320 cm where there is a slight increase. By ~ 78 cm the elements increased again and remain elevated to the top of the core. Manganese has three zones where it increases with peaks at 230 - 200 cm, 120 – 100 cm and ~ 60 cm.

Table 4.3.2. Matrix of elements that are highly correlated. Bold = >0.50.

	Si	K	Ca	Ti	Fe	Rb
Ca	0.57	0.12				
Ti	-0.10	0.93	0.05			
Fe	-0.35	0.71	-0.18	0.77		
Br	0.59	0.31	0.46	0.22	-0.08	
Rb	0.08	0.77	0.17	0.71	0.45	
Sr	-0.03	0.79	0.09	0.76	0.51	0.74
Mn	-0.37	0.18	-0.19	0.31	0.27	0.08
Zr	-0.25	0.06	-0.12	0.14	0.12	-0.02

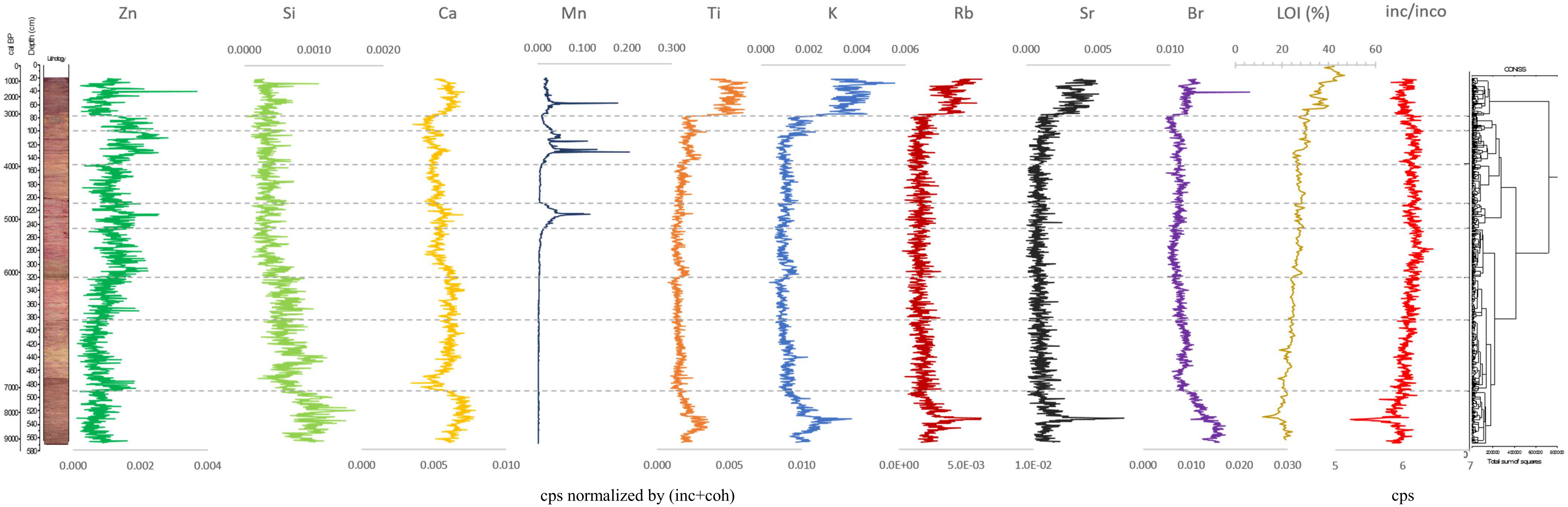

Figure 4.3.5 Lithology, organic matter content (% loss-on-ignition at 550°C (LOI$_{550}$)), and selected µ-XRF elements from Loch of Grimsetter. Zones are based on the constrained cluster analysis of the µ-XRF results and elements have been corrected for organic and moisture content by dividing the elemental results by the sum of incoherent and coherent scatter results (Davies et al. 2015).

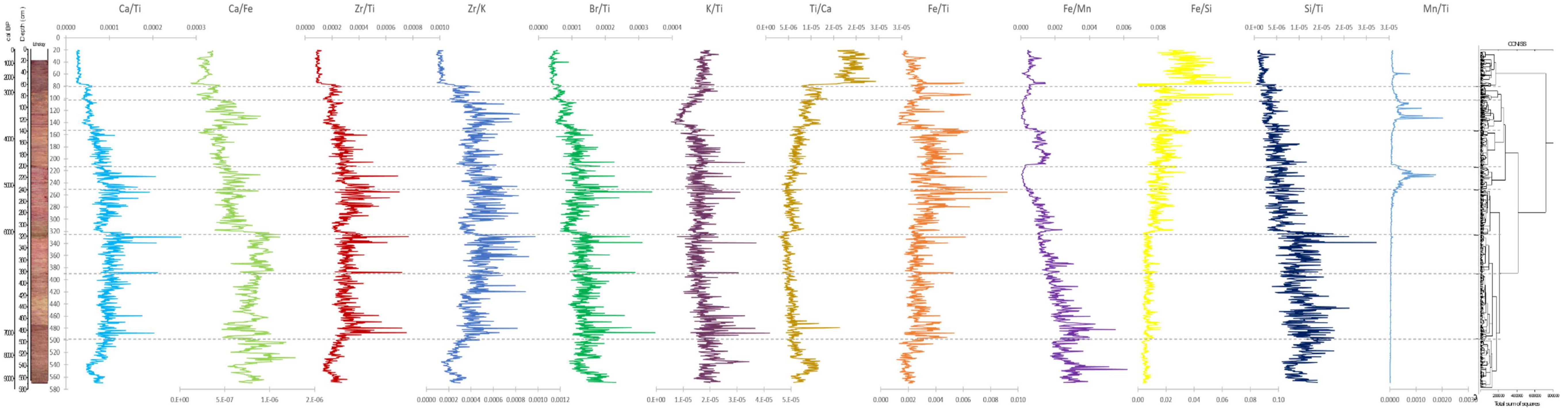

Figure 4.3.6 Selected elemental ratios from µ-XRF scan from Loch of Grimsetter. Zones are based on the constrained cluster analysis of the µ-XRF results. Ratios were determined by using elements corrected for organic and moisture content (Davies et al. 2015).

The relationship between individual elements provides important insights into the changing sedimentation at the site. The ratio between Fe/Mn is high at the base of the core and then gradually decreases until ~ 250 cm where there is a significant drop while Mn/Ti peaks. Subsequent peaks in Mn/Ti correspond with a decrease in Fe/Mn at the top of the core (Fig. 4.3.6). There are several subtle changes that occur in the geochemical analysis of the core from Loch of Grimsetter. At 321 cm, there is a consistent decrease in many of the elemental relationships (Fig. 4.3.6) such as Ca/Ti, Ca/Fe, Zr/Ti, Zr/K, and Br/Ti. All but Ca/Fe return to previous levels by ~ 295 cm while Ca/Fe remains lower. At the top of the core ~ 100 cm the most notable change occurs (based on the results of CONISS) where there is has been another decrease in many of the ratios along with the variation that was present in the rest of the record. There is an increase in the K/Ti and Ti/Ca relationship which began ~ 150 cm.

4.3.6. Chronology

The chronology of the core is constructed from four radiocarbon dates (3 from NERC and 1 from BETA). Bulk samples were used due to the lack of plant macrofossils found in the sediment and high organic content. The dates from NERC were prepared and analysed at the AMS radiocarbon facility at East Kilbride, Scotland. The Beta sample was analysed at Beta Analytic Inc., Miami Florida, USA (Table 4.3.3). The dates were calibrated using CALIB 7.1 (Stuiver et al., 2013) and the IntCal13 atmospheric calibration curve (Reimer, 2013). Due to the lack of magnetic susceptibility results (see below), a search of tephra layers was undertaken based on the results of the µ-XRF elemental scan using elements that have been associated with tephra (Mn, Fe, TI, Zr, K, Co/Ni) (Davies et al., 2015). Minute concentrations of tephra were located however spread out over a large section of the core and therefore not ideal for use as an age constraint at this time. The acceptable dates were then used to construct an age-depth model using R (R Development Core Team, 2016) and the package Clam (Blaauw, 2010). The resulting age-depth model based on the four radiocarbon dates is presented (Fig. 4.3.7).

Table 4.3.3 Radiocarbon dates submitted to NERC and Beta, calibrated using CALIB 7.1 (Stuiver et al., 2013) and the IntCal13 atmospheric calibration curve (Reimer, 2013).

Sample ID	Depth (cm)	Material	^{14}C age ± error	$\delta^{13}C_{VPDB‰}$ ±0.5	Calib cal age BP (2 sigma)	Calib wmean cal BP
SUERC-68217	564.5-565.5	Bulk	8147 ± 40	-20.8	9,006 –9,144	9,083
SUERC-68218	478.5-479.5	Bulk	5985 ± 38	-22.1	6,734 –6,934	6,824
SUERC-68219	69.5-70.5	Bulk	2841 ± 37	-21.9	2,860 –3,062	2,949
Beta-447118	319-320	Bulk	5320 ± 30	-29.5	6,200 –5,995	6,094

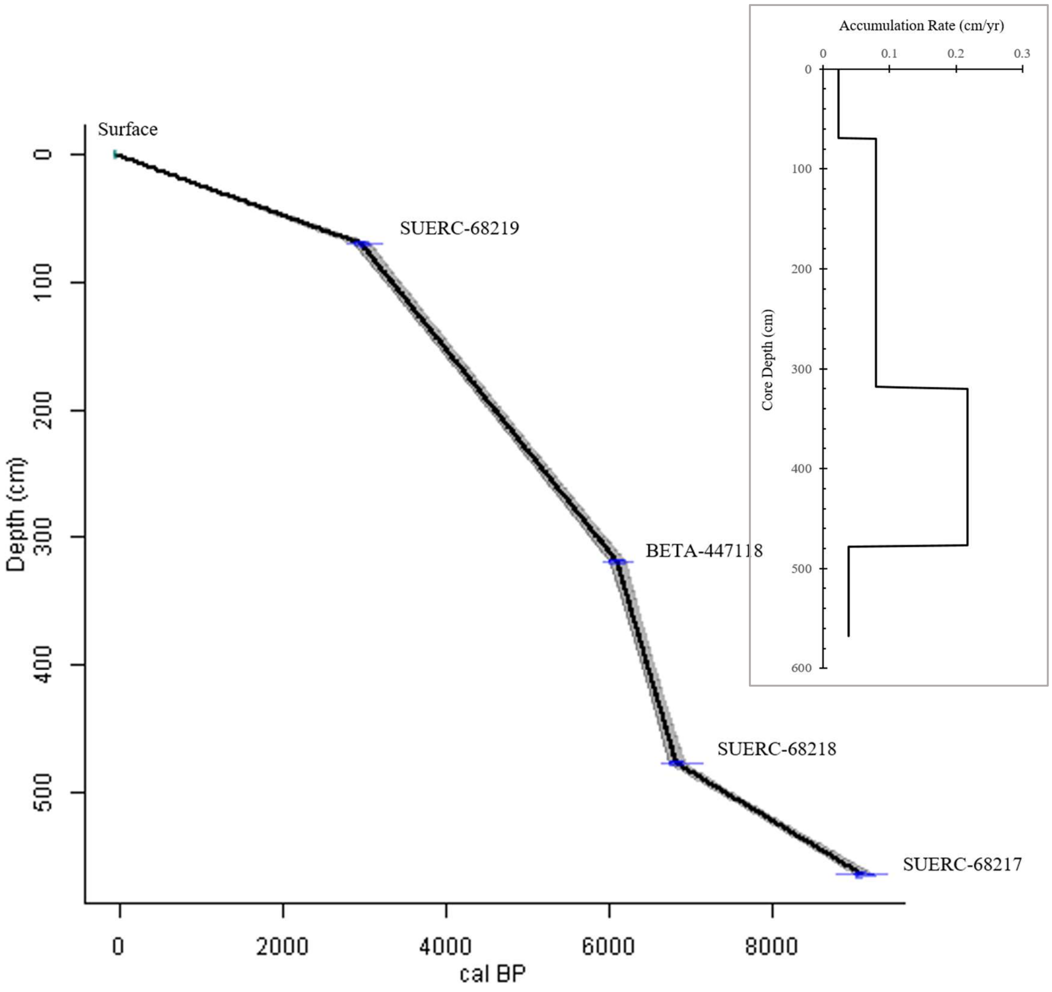

Figure 4.3.7 The selected calibrated AMS ^{14}C dates used to construct an age-depth model for Loch of Grimsetter generated using R v. 3.3 and the package Clam (Blaauw, 2010). Accumulation rate is also presented (inset).

4.3.7.Diatoms

Eighty-three samples were counted from the base of the core to 20 cm from Loch of Grimsetter, with 150 diatom species identified throughout the record. The diatom assemblages are divided into six distinct temporal zones and four sub-zones identified by CONISS (Grimm, 1987) and supported by optimal zonation (Bennett, 1996). The primary CONISS break with the highest variance reduction was detected at 101 cm (*c*. 3350 cal BP). The break with the second highest level of variance occurred at *c*. 6880 cal BP (480 cm), and the third highest break located at 317 cm (*c*. 6060 cal BP). The fourth highest break occurs at 69 cm (*c*. 2900 cal BP). The six zones (and four sub-zones) have been labelled GRMD-1 to GRMD-6 (with the four sub zones GRMD-1a/b and GRMD-4a/b) (Fig. 4.3.8) and are described in detail below.

GRMD-1 (a/b) (568 – 528 cm; *c.* 9170 – 8100 cal BP): The basal zone GRMD-1 is dominated by the planktonic species *Cyclotella stelligeroides* along with *Fragilariforma virescens*, *Nitzschia* species, *A. minutissimum, R. pusillum*, and *B. neoexilis*. (Fig. 4.3.8).

GRMD-1b (528 cm - 480cm; *c.* 8100 – 6880 cal BP): There is a slight decrease in *C. stelligeroides* at the break between zones 1a and 1b which then increases throughout GRMD-1b. There is also an increasing trend found in *S. construens* and *Cyclotella bodanica* var *lemanica* from zone GRMD-1a to GRMD-1b. The diatom concentration also decreases at this transition point. Species richness in this zone fluctuates between 31 and 38 species, peaking at the very top of the zone at 43 species. The dissimilarity between the samples reflects these changes as Bray-Curtis index also fluctuates noticeably along with species evenness reflected in Hills N2 values.

GRMD-2 (480 – 372 cm; *c.* 6880 - 6340 cal BP): There is a decrease in the dominant *C. stelligeroides* with an increase in *F. exigua* from the previous zone into GRMD-2 and an increase in *Brachysira neoexilis*. *Staurosira construens* steadily decreases progressing through the zone reaching its lowest concentration at 425 cm (*c.* 6580 cal BP) and then increases again by the top of the zone (Fig. 4.3.8). There is also a corresponding decrease in the proportion of *A. minutissimum* and increase in *C. stelligeroides* at 425 cm. Species richness again fluctuates throughout the zone, however at a lower frequency, but higher amplitude than the basal zone. Bray-Curtis has sharp fluctuations at the beginning of the zone while by the middle of the zone ~ 450 cm (*c.* 6690 cal BP) becomes less drastic. This change in amplitude and change in frequency is also reflected in the Hills N2 where there are sharp changes at the beginning of the zone that in turn become more gradual (Fig. 4.3.8).

GRMD-3 (372 - 317 cm; *c.* 6340 - 6060 cal BP): Zone GRMD-3 represents a short period characterized by an increase of *B. neoexilis* (characteristic of bogs and associated with sphagnum) which reaches its maximum abundance in the core. *S. construens* also disappears at the beginning of the zone and only reappears at the top in low abundance. The assemblage continues to be dominated by *F. exigua* and *C. stelligeroides*. However, *C. stelligeroides* abundance has decreased from a maximum abundance of 30 % in the previous zone to ~ 20 %. Diatom concentrations stay low compared to previous zones (Fig. 4.3.8) and the change from lower amplitude, lower frequency changes in the Bray-Curtis similarity index that began in the middle of the previous zone, continues until 110 cm (*c.* 3460 cal BP). At the same time, the species evenness decreases in this zone while species richness again fluctuates between 36 and 43 species (Fig. 4.3.8).

GRMD-4a (317 - 205 cm; *c.* 6340 - 4650 cal BP): The diatom assemblage in zone GRMD-4a remains quite stable throughout. There is a suggestion of a degree of periodicity corresponding to the rising and falling of *S construens*, *F. exigua*, *C. stelligeroides* and *Achnanthidium minutissimum* every 24 cm (~ 300 years). There is also the emergence of *Sellaphora vitabunda*, *Diatoma tenuis*, and *Staurosirella leptostauron* in this zone. The diatom concentrations also rise and fall with the periodicity in the assemblages (Fig. 4.3.8).

GRMD-4b (205 - 101 cm; *c.* 4650 - 3340 cal BP): The sub zones of *GRMD-4* are defined by a change in rare species such as *Staurosirella leptostauron*, *Nupela vitiosa* and *Tabellaria flocculosa* at 205 cm (*c.* 4650 cal BP) which either initially appears in zone *b* or at a higher concentration than in zone a. There are several periods of sharp increases in species richness in this zone occurring at 281, 249, 177, and 137 cm (*c.* 5610, 5210, 4300, and 3800 cal BP respectively) where richness spikes to over 40 species present. However, Bray-Curtis does not reflect these changes, however Hills N2 does with peaks in the values at those locations. Hills N2 reaches a low point at 200 cm (*c.* 4590 cal BP) that does not have a corresponding event occurring in the richness data (Fig. 4.3.8).

GRMD-5 (101 - 69 cm; *c.* 3340 - 2910 cal BP): GRMD-5 is another short zone consisting of 25 cm of sediment (~ 315 years) which denotes the start of a notable change in diatom species composition. This is the location of the highest CONISS break and sees the appearance of ‘small’ *Fragilaria* (s.*l.*) species. These were too small to be identified to species level but could be *Pseudostaurosira elliptica, S. pinnata* or F. *exigua*. *Nupela vitiosa* also increased in concentration and *Navicula minima* appeared for the first time in the palaeorecord along with a decrease in *C. stelligeroides*. Species diversity also decreases with a disappearance of the rarer species (right side of the diagram Fig. 4.3.8). Species richness peaks slightly in this zone to 40 species, however the assemblage is quite different based on the results of Bray-Curtis analysis which reaches its highest value of 1.9 at 97 cm (*c.* 3290 cal BP).

GRMD-6 (69 - 20 cm; *c.* 2910 - 800 cal BP): Diatom concentrations increase in the uppermost zone GRMD-6. The changes that started in the previous zone continue with a decrease in *S. construens* while the smaller *Fragilaria* (*s.l.*) reach their highest abundance and *C. stelligeroides* reach their lowest abundances (Fig. 4.3.8). There is a turnover of species with the appearance of *Planothidium conspicuum*, *Aulacoseira granulata*, *Hippodonta capitata* and *Eolimna rotunda* and a disappearance of *Achnanthes childanos*, and reduction of *Tabellaria flocculosa*. Species richness peaks in this uppermost zone with 44 species, however it decreases to 26 by the top of the record while evenness remains comparable to the previous zone. The similarity of the samples in this zone are more similar due to the less variance in the Bray-Curtis results.

The results of the PCA show that the first axis explains 31.4 % ($\lambda = 0.314$) of species variation per sample depth and PCA axis-2 16.0 % of species variation ($\lambda = 0.381$). The site scores (Fig. 4.3.9b) show that for the beginning of the loch's history, diatom assemblage began in the upper left quadrant driven by *C. stelligeroides*. For most of the record it shifted back and forth above and below the first axis migrating to below the first axis until ~ 100 cm when the assemblage shifted to a very distinct and unique assemblage to the right of the secondary axis (Fig. 4.3.9).

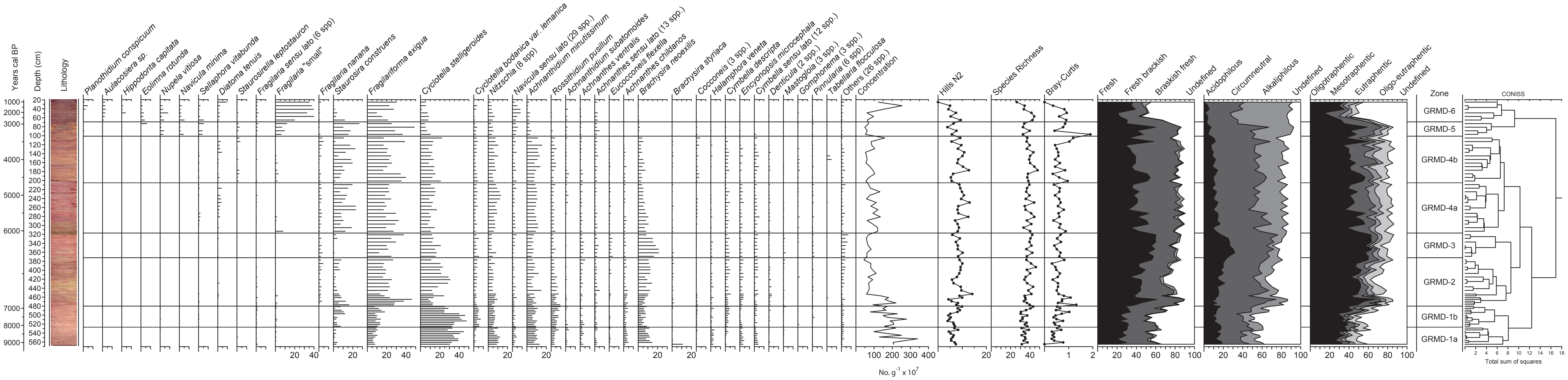

Figure 4.3.8 The dominant (>3 %) diatom taxa found in Loch of Grimsetter. Species with a relative abundance below 3% were grouped based on genus. The record has been divided up into zones based on a constrained cluster analysis (CONISS) (Grimm 1987). Species evenness, richness and Bray-Curtis similarity are presented. The diatom assemblage has also been summarized in respect to water pH and salinity of the water (Van Dam et al. 1994).

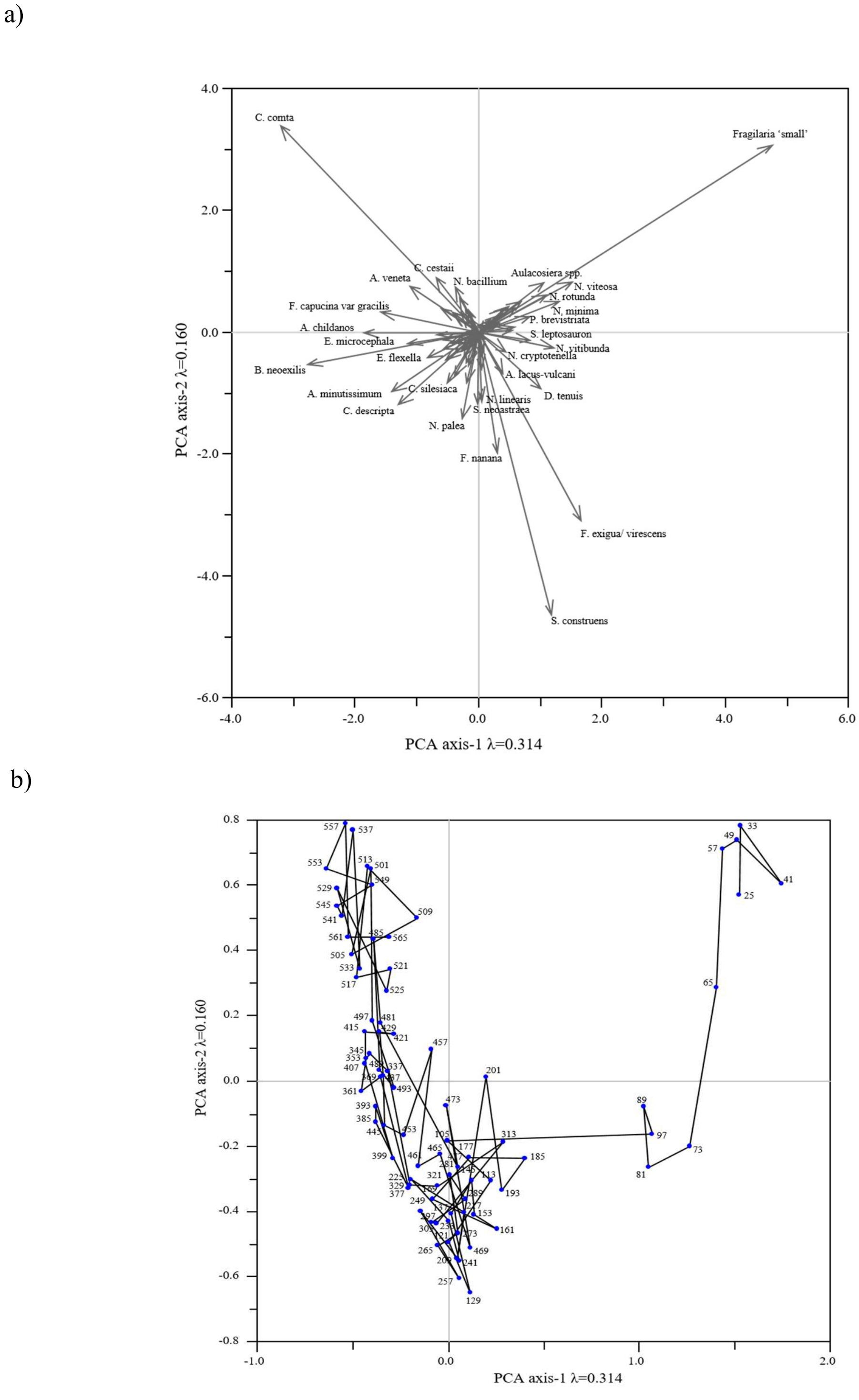

Figure 4.3.9 PCA analysis of the diatom species from Loch of Grimsetter in relation to sample depth. PCA axis-1 explains 31.4 % of the observed variation in species composition related to sample depth and PCA Axis 2 explains 16.0 % of species variation. Graph 'a' shows the relationship of species to the PCA axis and 'b' shows the relationship of the samples to the PCA axis.

4.3.8.Pollen

The core was sampled for pollen analysis at 8-cm intervals (1-cm^3 samples) for 48 samples the full length of the core. Six significant temporal zones were identified by CONISS (Grimm, 1987) and optimal zonation (Bennett, 1996) in the pollen assemblage. The primary CONISS break with the highest variance reduction was detected at 77 cm (*c*. 3040 cal BP) (Fig. 4.3.10). The break with the second highest level of variance occurred at *c*. 5460 cal BP (269 cm). The third highest break occurs at 534 cm (*c*. 8290 cal BP) (Fig. 4.3.10). The fourth and fifth occur at 390 cm (*c*. 6420 cal BP) and 209 cm (*c*. 4700 cal BP) respectively. The zones have been labelled GRMP-1 to GRMP-5.

LPAZ GRMP-1 (568 - 534 cm; *c.* 9170 - 8290 cal BP): The basal LPAZ GRMP-1 is short and characterized by typical *Betula/ Pinus/ Corylus avellana* type canopy with Poaceae/ *Empetrum* dominant understory with some Ranunculaceae. The spore record is dominated by Polypodiaceae. The ratio of trees/herbs/shrubs show around 45 % trees, 20 % shrubs and 35 % herbs (Fig. 4.3.10).

LPAZ GRMP-2 (534 - 390 cm; *c.* 8290 – 6420 cal BP): *Betula, Pinus, Corylus avellana* type and Poaceae still dominate in LPAZ GRMP-2 along with an increase in Ranunculaceae. Species diversity has increased with the appearance of Asteraceae Subfam: Asteraceae*, Caltha, Campanula* and *Scabiosa*, though their abundances are quite low and are classified as rare. Both *Polypodium* and Polypodiaceae decrease from levels seen in zone GRMP-1. The concentrations of the pollen fluctuate through the LPAZ, with peaks at 516, 480, and 432 cm (*c.* 7820, 6880, 6610 cal BP respectively).

LPAZ GRMP-3 (390 - 269 cm; *c.* 6420 - 5460 cal BP): The proportions of Poaceae slightly decrease in LPAZ GRMP-3 and there is a corresponding increase in *Corylus avellana* type. *Betula* levels remain consistent at ~ 20 %, along with *Pinus* and *Alnus* (Fig. 4.3.10). There is also an increase in Polypodiaceae. Pollen concentrations slightly fluctuate through the LPAZ, though they remain stable overall.

LPAZ GRMP-4 (269 - 209 cm; *c.* 5460 - 4700 cal BP): LPAZ GRMP-4 is a short zone (~ 50 cm) which sees a peak of 25 % in Poaceae and a slight reduction of ~ 5 % in *Corylus avellana* type and *Betula*. *Alnus* slightly increases while *Pinus* decreases (Fig. 4.3.10). At the top of this LPAZ (210 cm; *c.* 4720 cal BP) *Plantago lanceolata* begins to increase from ~ 8 % to ~ 12 %. *Rumex* appears for the first time in this zone while Cyperaceae peaks in this LPAZ at ~ 10 %, the highest abundance for it in this core. Pollen concentrations remain similar to the previous LPAZ.

LPAZ GRMP- 5 (209 - 77 cm; *c.* 4700 - 3040 cal BP): At the start of this LPAZ, Poaceae slightly decreases and then remains constant at ~ 15 % while *Empetrum* gradually increases from 10 % to 20 % by the top of the LPAZ (Fig. 4.3.10). *Plantago* has also increased from the previous zone to around 18 %. Midway through the LPAZ, *Erica* starts to increase around 120 cm (*c.* 3580 cal BP) from 1 to 5 %. At the top of the LPAZ at ~ 100 cm (*c.* 3330 cal BP) *Betula* and *Alnus* begin to decrease. Again, the pollen concentration remains comparable to the previous LPAZs. Charcoal starts to appear at the bottom of the zone (Fig. 4.3.10).

LPAZ GRMP- 6 (77 - 20 cm; *c.* 3040 - 800 cal BP): The top-most LPAZ GRMP-6 is defined by CONISS as the highest sum of squares break in the record. This is due to a considerable decrease of *Corylus avellana* type*, Betula, Alnus* and *Pinus*, together with an increase in Poaceae*, Calluna, Erica* and *Empetrum*. Pollen concentrations peak in this LPAZ. Charcoal concentrations also increase to their highest values (4833 charcoal per cm^3) in this LPAZ starting at 65 cm (*c.* 2740 cal BP), around the same time as the proportions of trees and shrubs decrease (Fig. 4.3.10).

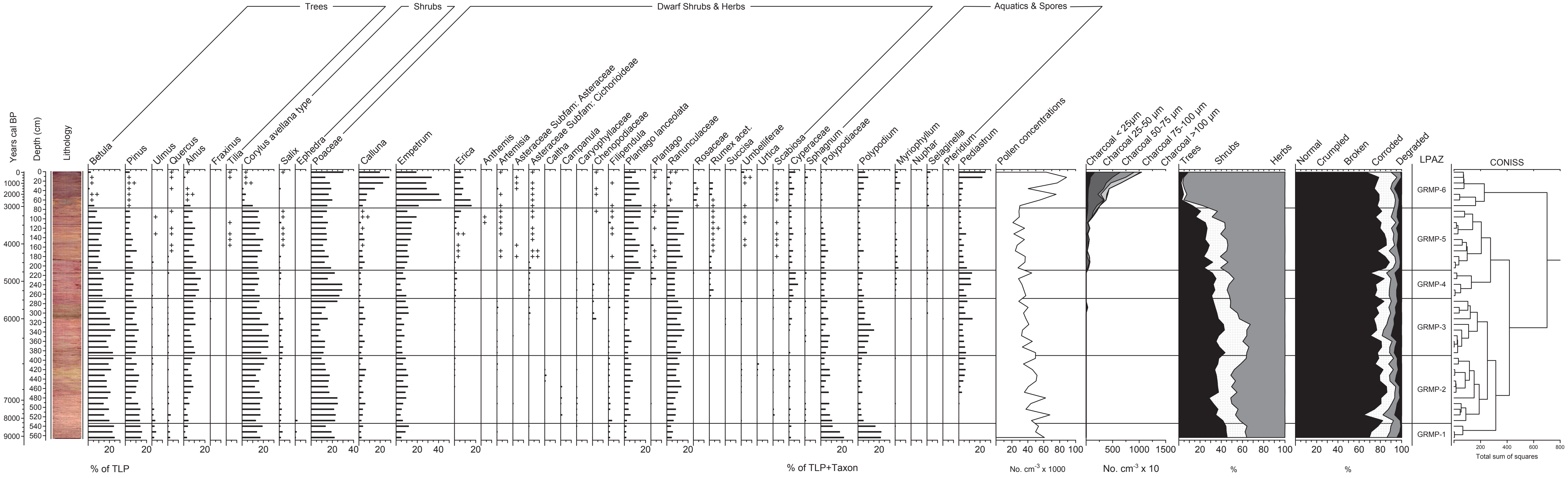

Figure 4.3.10 Percentage pollen and spore diagram of the dominant taxa found in Loch of Grimsetter. Taxa under 1 % are designated by '+' and under 2 % by '++'. Solid lines indicate the major zones and dashed lines indicate the sub-zones determined by constrained cluster analysis (CONISS) (Grimm 1987). The modelled dates appear on the left side by the y-axis along with a lithology of the core.

4.4. Fisher Information

This section will explore the relationship between the change of multiple proxies through time using time-series statistical techniques. The previous sections have shown the differing responses to climatic influences across the three study lochs located on Shetland and Orkney along with differences in the rates of change during the Lateglacial through the early Holocene. It has been demonstrated in palaeoenvironmental records that during the Lateglacial more abrupt changes had occurred in proxy data while during the Holocene changes were deemed subtler or gradual (Shuman et al., 2005).

Fisher Information was applied to diatom, pollen, and geochemical data which are distinct types of data that are routinely used in palaeoenvironmental reconstructions to provide information on the state of the catchment, terrestrial vegetation, and the aquatic environment. The benefit from using multiple proxies in this type of analysis is that each proxy responds to change in diverse ways. Diatoms have a short life span and generation time and respond quickly to changes in the aquatic environment while pollen tends to have a lag in its response to environmental change due to the longevity of certain plant groups (i.e. trees) and provides a terrestrial signal for the catchment. The geochemical data provides information on both the catchment and aquatic environment as the sediment composition is made up of allochthonous and autochthonous input and, depending on the environmental change, can respond quickly (i.e. precipitation, catchment instability) or slowly (i.e. drought).

As outlined in Section 2.6, the technique chosen for this analysis is Fisher Information, which takes into consideration the multivariate nature within ecological communities and the resulting non-linear responses that can occur with changes in the environment. These results can be used to identify trends in the data that can result in a signal prior to a critical transition by combing multiple variables into an index (FI) that can be followed through time (Eason et al., 2014). It also recognizes the fact that abrupt changes can be a result of sudden tipping points and that prolonged instability can also instigate regime shifts (Spanbauer et al., 2014).

Between the three lochs, the Lateglacial and early Holocene is represented for both Orkney and Shetland. These three proxies should provide a comprehensive look at how the vegetation, catchment and loch are responding to changes in the environment and at what temporal resolution and to determine whether there are any trends in the data that might be used as warning signals prior to a change in assemblage. It will also determine if these methods are appropriate for palaeoecological data and highlight the benefits from their use.

4.4.1.Results: Loch of Sabiston

The results of the FI analysis from Loch of Sabiston are presented in Figure 4.4.1. The pollen and diatom records were calculated with a window of 2 samples and progresses through the record one sample at a time while the μ-XRF data had a window size of 4 with the data smoothed to an 8-point average. This was necessary due to the amount of noise in the μ-XRF data (Fig. 4.4.1a). Starting with the μ-XRF data, there seems to be some agreement with the CONISS breaks and sharp changes in FI. At the beginning of the record there is a decrease in FI which signifies that the system is losing order and stability. This continues until *c*. 16,500 cal BP when FI begins to increase until *c*. 15,000 cal BP. At this point, there is a sharp decrease which indicates a regime shift that lasts until *c*. 14,200 cal BP. The record during GI-1d illustrates a gradual shift until this point where the beginning of the interstadial is recorded. Through the Bølling interstadial there is a steady climb of FI to *c*. 13,250 cal BP with a series of increases and decreases which correlate with GI-1d, the Allerød and IACP (GI-1b) and possibly other climatic shifts that have been recorded in the Greenland ice core records (Rasmussen et al., 2014). Again, there is an abrupt decrease in FI at the beginning of the Younger Dryas stadial. The Younger Dryas is relatively variable with increasing and decreasing FI values, followed by an extreme rise in FI transiting out of the Younger Dryas and into the Holocene at *c*. 11,770 cal BP. The rapid rise in FI indicates that the geochemical system is becoming more organized and stable (Ahmad et al., 2016). Therefore, the Holocene appears to have been more stable than the preceding LGIT. The beginning of the Holocene has a high FI and appears to have been stable, however, there is relatively high variability throughout this period. At *c*. 8880 cal BP there is a rapid decrease in FI that is in good agreement with a CONISS zone (Fig. 4.4.1a) which then transition into a prolonged stable period. This stability lasts until *c*. 5000 cal BP where a start of a gradual downward trend that extends to the top of the record indicating a reduction in order and stability of the geochemical record.

The diatom results of the FI analysis (Fig. 4.4.1b) suggests the GI-1d was relatively stable until *c*. 14,700 cal BP after which FI became unstable signifying a tipping point occurring that began the interstadial. There are multiple sharp increases and decreases in FI during the Bølling – Allerød which indicates a very unstable time leading up to the Younger Dryas. During the Younger Dryas, the FI is stable but with an overall decreasing trend that ends in a sharp decline signifying the beginning of the Holocene. The first part of the Holocene is characterised by a steadily increasing FI score until *c*. 9120 cal BP after which FI became stable and orderly until *c*. 7700 cal BP when FI begins to decrease suggesting an increase in instability in the diatom assemblage. At *c*. 6000 BP there is a sharp decline in FI that is not reflected in the zonation systems (Fig. 4.1.8), but is marked by a drop in *Cyclotella meneghiniana* and an increase in

Gomphonema angustum abundance. At *c*. 5200 cal BP the FI scores begins to gradually decrease to the top of the record.

The FI results for the pollen record indicates changes in the assemblages also not reflected in the zonation system for the LGIT (Fig. 4.4.1c). There is a definite instability signal during the Bølling – Allerød interstadial. The Allerød pollen assemblage (which is combined with the Younger Dryas in the results from CONISS) has a low FI that decreases until the transition into the Younger Dryas *c*. 11,700 cal BP when FI increases. The variability during the Younger Dryas and the beginning of the Holocene are similar in indicating a very unstable time. There is a subtle upward trend in FI scores which continues into the Holocene. Until *c*. 10,000 cal BP where it begins to decrease again until *c*. 8600 cal BP when there is a sharp increase in FI to *c*. 8500 cal BP. After this the Holocene record stabilizes and remains consistent until *c*. 4500 cal BP when it begins to decrease to the top of the record.

4.4.2.Results: Loch of Clumlie

There is high variability in the results of the FI analysis for µ-XRF record in Loch of Clumlie (Fig. 4.4.2a). The geochemical FI scores in GI-1d shows instability until *c.* 16,030 cal BP and then FI becomes more stable as it heads into the Bølling interstadial. During the Bølling the FI score remains stable with high variability until *c.* 13,740 cal BP when dynamic order begins to decrease and the system becomes more unstable as it proceeds into the GI-1d. The FI score remains low though this zone and into the Allerød when it increases and peaks at c. 13,030 cal BP. The end of the Allerød is characterized by a step-down trend which could indicate the IACP. During the Younger Dryas, the FI score has high variability with a subtle trend upwards to *c.* 12,330 cal BP. The second half of the Younger Dryas sees another peak in FI score *c.* 11,860 cal BP with subsequent tipping point reached as it heads into the Holocene. There is a prolonged transitional zone in the record that has a low and stable FI prior to the sharp increase into the Holocene. The uppermost portion of the core that contains the Holocene must be interpreted with caution because of the evidence for hiatuses. After ~ 40 cm core depth the FI value steadily increases which suggest a more stable environment at the top of the record.

The diatom FI results are in good agreement with the abrupt changes observed in the assemblage along with the results of the zonation established with CONISS (Fig. 4.4.2b). The GI-1d shows a general downward trend in FI until *c.* 14,750 cal BP when an abrupt but small decrease occurred prior to the CONISS zone at *c.* 14590 cal BP. The FI then suggests a stable Bølling until *c.* 13,900 cal BP when the FI begins to decrease until the after the start of the Allerød. The Allerød shows a steady decrease of FI throughout the period until the start of the Younger Dryas which also agrees with the CONISS zonation. The FI steadily increases during the Younger Dryas until *c.* 12,330 cal BP after which the variability of FI increases. There is a tipping point where the species assemblage changes from one group to another in the second half of the Younger Dryas (at *c.* 12,050 cal BP) which is also indicated by CONISS. There is an increase in FI at the end of the Younger Dryas followed by a downward trend indicating a loss of stability of the record. This downward trend continues into the Holocene until ~ 60 cm core depth. The FI during the Holocene is very stable and orderly, however, caution in interpretation is needed due to the complexity of the top of the core with the probably hiatus.

The pollen FI record for Loch of Clumlie shows there is a slight general trend through the GI-1d into the Bølling - Allerød of decreasing order and resilience in the system (Fig. 4.2.2c). Like the record from Loch of Sabiston, there is no abrupt change during the transition from the end of the Allerød. The result of the FI through the Younger Dryas is variable with a slight decreasing trend during the first half of the Younger Dryas and a slight increasing trend during the second half. By *c.* 12,330 cal BP core depth the FI index remains quite stable and orderly through the Holocene with little variability occurring.

4.4.3. Results: Loch of Grimsetter

The FI results from Loch of Grimsetter illustrate a very stable environment over the length of the record. In the µ-XRF results there is probable evidence for the 8.2 ka event (yellow arrow, Fig. 4.4.3a), but the environment remains relatively stable throughout the record. It could be argued that the perpetration of the 8.2 ka event began at *c.* 8500 cal. BP based on the FI analysis. Another probable event occurred *c.* 6900 cal BP that CONISS does not record. The different sedimentation rates are evident in the XRF data as the middle of the record from *c.* 6800 to *c.* 3000 cal BP has a slower sedimentation rate and therefore the data points will be at a higher resolution than the rest of the core. However, the overall trends can be interpreted.

The FI results also suggest relative stability in the diatom record overall. However, there is a difference in the rate of change from *c.* 6800 to *c.* 6000 cal BP with a higher amount of instability in the record. Based on the large decrease in FI there are two other potential events, the first occurring between *c.* 8400-8150 cal BP, which is most likely the 8.2 ka event. The other begins at *c.* 3700 cal BP, ~ 360 years prior to the CONISS break in the assemblage. The early Holocene pollen record is quite stable in Loch of Grimsetter with a loss of order occurring at *c.* 6330 cal BP after the CONISS break at *c.* 6420 cal. It then returns to a stable state for the remainder of the record. At *c.* 6000 cal BP there was a loss of order followed by a return to a stable state which lasted for ~ 530 years.

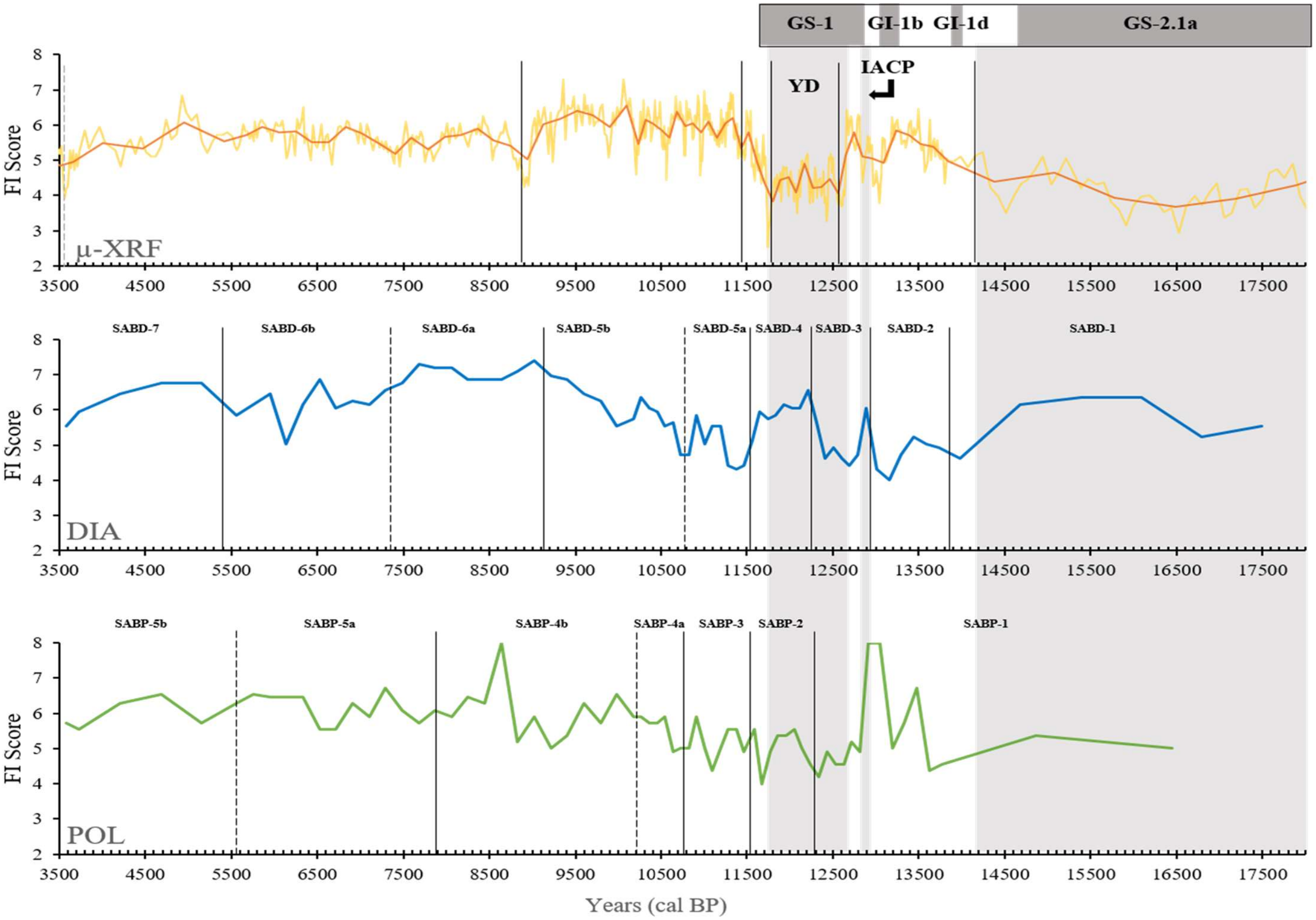

Figure 4.4.1 Result of FI analysis of the three proxies from Loch of Sabiston. The vertical lines denote the CONISS breaks that were presented in section 4.1 for each of the proxies. The dashed vertical lines represent the subzone locations of the CONISS breaks. The red line in the XRF graph represents an 8-point smoothing of the record. The grey shaded regions show where the sediment has returned to clay during the IACP and the YD (labelled), while the white regions consists of marl. At the top of the diagram for reference, is the Greenland ice core events based on the chronology from Rasmussen et al. (2014). The dark grey areas are the cold events labelled from GS 2.1 to GS-1.

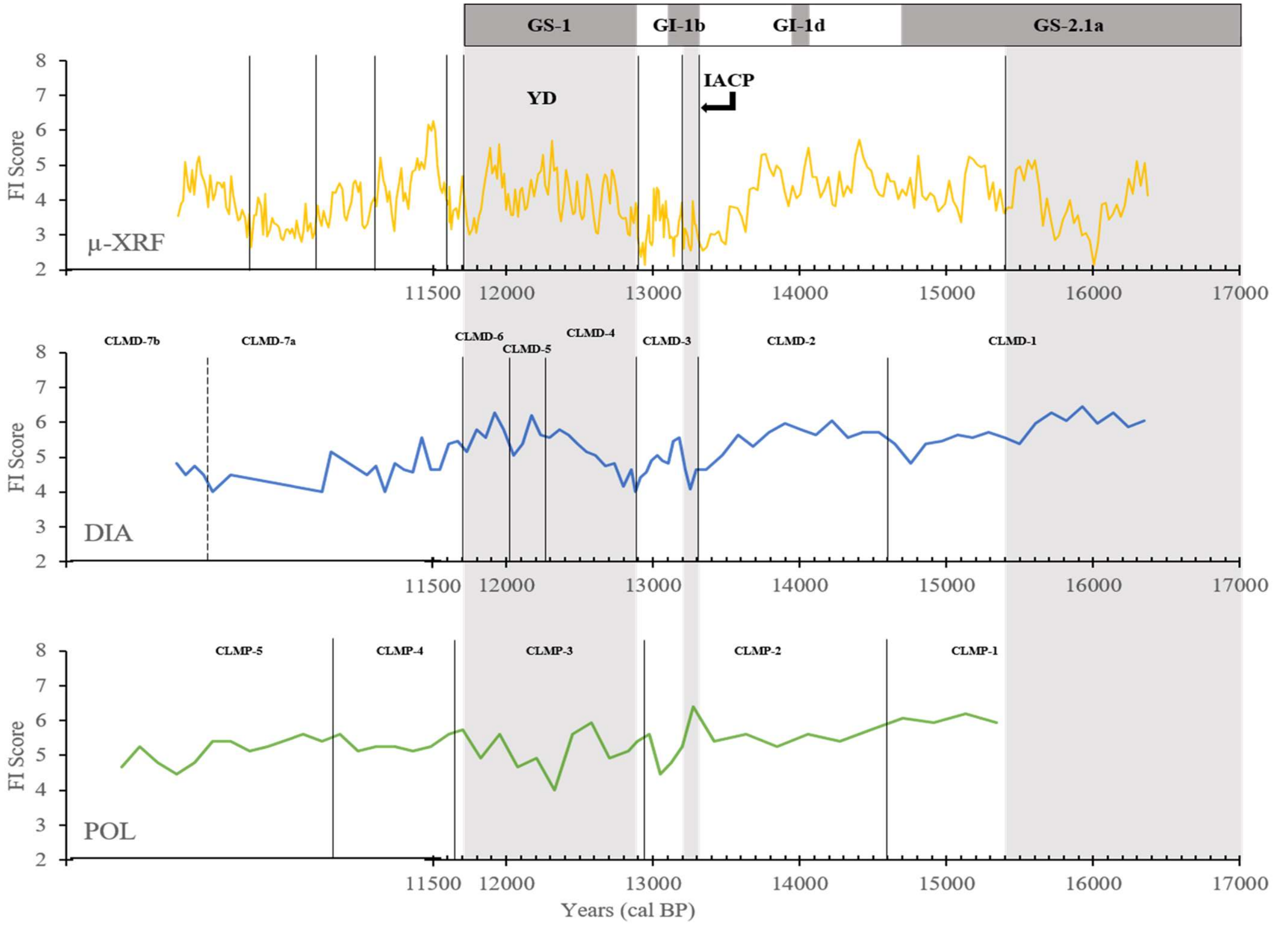

Figure 4.4.2 Results of FI analysis from the three proxies from Loch of Clumlie. The vertical lines denote the CONISS breaks that were presented in section 4.1 for each of the proxies. The dashed vertical lines represent the subzone locations of the CONISS breaks. The grey shaded regions show where the sediment has returned to clay during the IACP and the YD (labelled). At the top of the diagram for reference, is the Greenland ice core events based on the chronology from Rasmussen et al. (2014). The dark grey areas are the cold events labelled from GS 2.1 to GS-1.

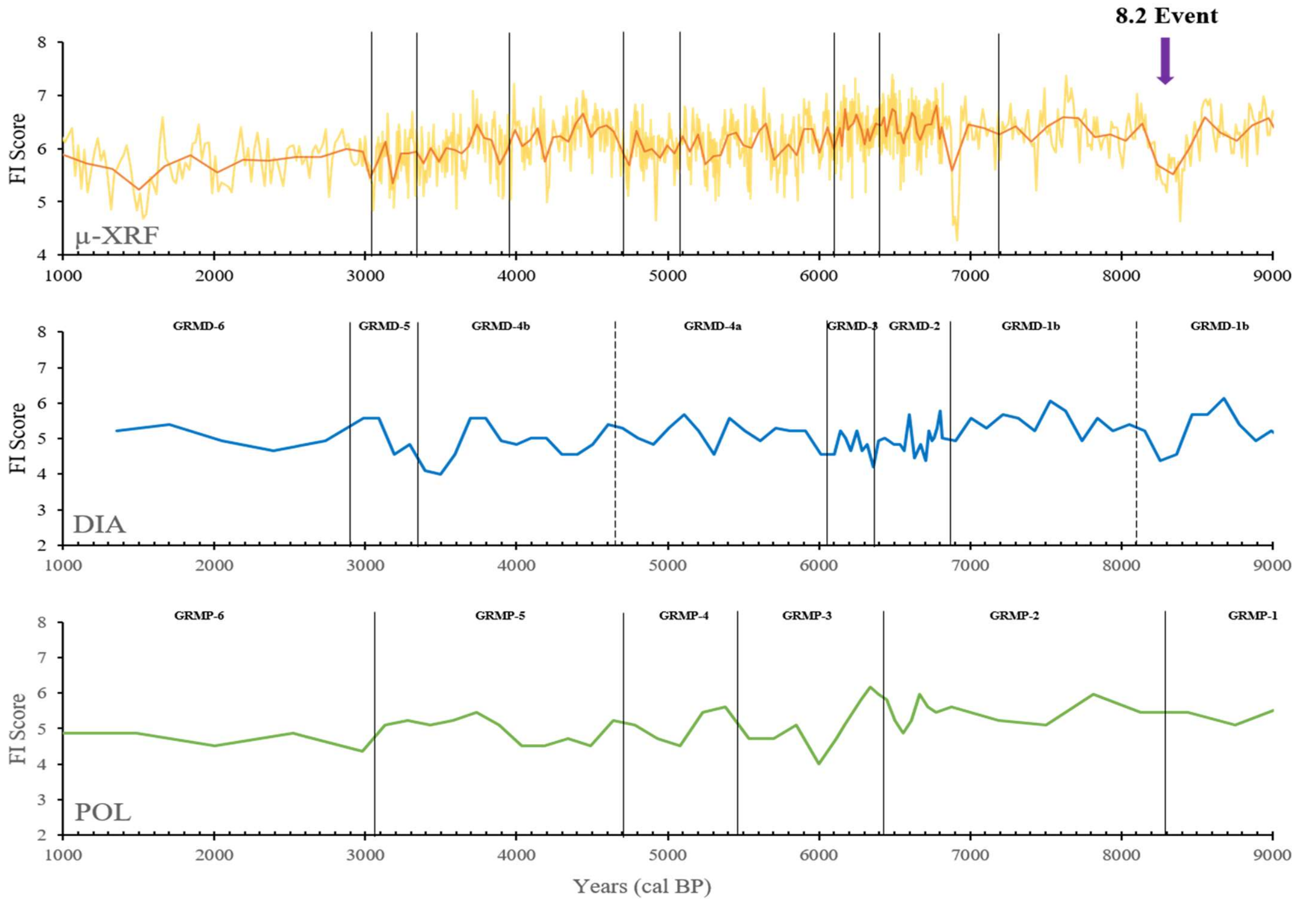

Figure 4.4.3 Results of FI analysis from the three proxies from Loch of Grimsetter. The vertical lines denote the CONISS breaks that were presented in section 4.3 for each of the proxies. The red line in the XRF graph represents an 8-point smoothing of the record. Evidence for the 8.2 ka event is denoted with an arrow at the top right of the figure.

Chapter 5

5. Discussion and Synthesis

The central aim of this research is to determine the nature and the timing of palaeoenvironmental changes that occurred at the end of the LGM through to the early Holocene on the island groups of Shetland and Orkney. Traditional constrained cluster analysis was used along with Fisher Information, a new technique to assess the stability of a system (in this case diatom, pollen, and geochemical systems), to determine the nature and the timing of changes that occurred in each of the proxies and how each analysis relates to the other. Discussion will focus on the relationship between diatom, pollen, and geochemical records from the study lochs on Shetland and Orkney and will examine how these results compare to their respective island archipelagos and to the wider North Atlantic

This synthesis will begin with an examination of the chronological uncertainties that occur when undertaking palaeoenvironmental studies and the considerations that were taken in this present study. This will be followed by an examination of the robustness of applying Fisher Information to palaeoenvironmental data and how well it detects tipping points in the data. General observations of each of the study lochs will be presented followed by the discussion of the palaeoenvironmental evidence for climatic change arranged chronologically from deglaciation, through the Bølling-Allerød interstadial and the Younger Dryas Stadial concluding with the early Holocene.

5.1. Chronological Considerations

The key to understanding any synthesis of climatic and environmental changes that have occurred though time is having a robust chronology. Chronologies developed from Shetland and Orkney are often impacted by dating issues such as *old* carbon or contamination of older sediment from younger sediment above which compromises the accuracy of ages (Marty and Myrbo, 2014; Shotton, 1972; Yu et al., 2007). In locations such as the Northern Isles, fine-tuning of chronologies can be carried out with the application of tephrochronology. Using known and well-dated tephra horizons that are independent of site specific variables can be used to correlate palaeoenvironmental data sets (Bronk Ramsey et al., 2014; Lowe, 2015). Another method is to correlate the sequence of regional climatic events to other records that are well-dated or stratigraphically constrained such as the Greenland ice core records (synchronized NGRIP, GRIP, and GISP2 records and resulting chronology presented in Rasmussen et al. (2014) (used for this study) which are based on the $\delta^{18}O$ isotope ratio that is associated with variations in palaeoprecipitation and is temperature-controlled. However, it is acknowledged that by aligning two different climate records, it prevents the identification of any time transgressive changes across the region without leading to circular arguments (Baillie, 1991; Blaauw, 2012).

The chronology developed for Loch of Sabiston appears to be independently consistent with the ages of the Saksunarvatn and Vedde tephra layers used to constrain the age-depth model (Figure 4.1.8). This strengthens the reliability of the dates between the Younger Dryas and the early Holocene, but dates younger or older than the timespan between the tephra layers may be less reliable. In comparison with the Greenland ice core records (Rasmussen et al., 2014), it is apparent that there is an offset of ~ 200 years between the Sabiston age-depth model and the climate-stratigraphy. It is not clear however, if this offset is an artefact of the age-depth model or if it represents an actual delay in the Orcadian response to events recorded in the ice core records. This offset will be discussed further below. Despite this, the Loch of Sabiston record represents the most precise age-depth model from Orkney to date.

The chronology from Loch of Clumlie was difficult to reconcile due to the drying out of the loch and hiatuses during the early and mid-Holocene. There was visible evidence that root penetration most likely contaminated underlying older sediment with younger material. Also, the accumulation of lake sediment was halted and subjected to pedogenetic processes near the top of the core. The correlation of the Lateglacial climate-stratigraphy from Loch of Clumlie to the Greenland ice core records (the transitions from the LGM to the Bølling - Allerød, the IACP and the Younger Dryas) provides a chronological framework in which to place palaeoenvironmental interpretations. However, this restricts the interpretation of the evidence from Loch of Clumlie to a regional record based on the assumption that events were synchronous between Greenland and

Shetland. Finally, the chronology from Loch of Grimsetter, which covers the Holocene, is considered to be somewhat reliable due to the high organic content of the sediment and the lack of any evidence of contamination from earlier sediment or old carbon input. The identification of crypto tephra layers in the lochs from Shetland would have provided more robust independent age-depth models to enable comparisons to be made across the North Atlantic region but unfortunately no definitive tephra horizons were found in the cores.

5.2. The Nature and Timing of Events During the LGM and the Holocene

Fisher Information was developed to identify the dynamics of multivariate systems and to determine points at which regime shifts occurred (Spanbauer et al., 2014). It also attempts to record the changes and calculate the stability in species assemblages as they are occurring, and therefore it is used for active monitoring regimes (Cabezas et al., 2010). The application of FI to palaeoecological data is relatively new (Spanbauer et al., 2014) and many of the published research using palaeoecological data are theoretical outputs or based on simple or model systems (Ahmad et al., 2016; Eason et al., 2014; Fath et al., 2003; Karunanithi et al., 2008; Mayer et al., 2006). Other zonation procedures, such as constrained cluster analysis (CONISS, (Grimm, 1987)), averages the ecological changes occurring in the assemblages and therefore some species might be responding to events prior to others. This can be observed in the diatom, µ-XRF, and pollen figures in Chapter 4 for the three lochs where the zone boundary can sometimes be found on the peak or trough of a species' relative abundance. Overall for the three lochs, the zonation boundaries appear to correlate with transitional points in the FI results. In some instances, there is agreement regarding the timing of critical transitions with the zone boundary coincident with a FI peak or trough (Fig. 4.4.1 - 4.4.3), however, zones can also lie in the middle of a FI transition (i.e. in between zones SABD-4 and SABD-5, Figure 4.4.1). This indicates the FI is detecting the start of a change, with a decrease in FI (indicating a decrease in stability) occurring while the zonation is detecting the threshold at which the assemblages have changed significantly. There are also changes indicated by the FI index that are occurring in each of the proxies that have not been identified using CONISS. CONISS does not provide detail on the rate and duration of ecological changes (fast or slow, long or short) occurring across or between the zones while FI provides information on the nature of the change (fast or slow, long or short). This is observed in the Sabiston diatom record (Fig. 4.4.1), where there are two instances at *c.* 6100 cal BP and *c.* 10,000 cal BP when changes are evident in the FI index but are not reflected in the CONISS zones which have been deemed significant using Broken Stick (Bennett, 1996).

It is assumed that abrupt climate changes lead to abrupt ecological changes, however, the results of the FI index from this study suggests this is not always the case. By using the types of climate shifts outlined in Section 2.6, FI can be used to characterize the nature of changes in the diatom, pollen, and geochemical data sets observed in the palaeoenvironmental records from the three lochs. Types I and II (smooth regime shifts and threshold responses) were most observed in palaeoenvironmental data from this study with Type II being the most common. Type III (bistable shifts) were associated with anthropogenic activities (for example in Figure 2.19) (Randsalu-Wendrup et al., 2016). This is evident when comparing changes that occurred during the LGIT with those in the Holocene in the three study lochs. Based on the FI index most changes can be considered Type II tipping points based on threshold responses (Randsalu-Wendrup et al., 2016). There are more abrupt changes over shorter periods of time during the LGIT while the Holocene has more gradual shifts over longer periods. Also, the intensity of the shifts is higher during the LGIT with a larger difference between high and low FI values than in the Holocene.

FI demonstrates how an ecological system behaves by establishing how assemblages are changing (i.e. fast or gradual) and gives structure to the changes occurring between zonal boundaries providing information on different rates of change occurring throughout the zones. For example, the Loch of Sabiston diatom record suggests that ecological changes during the Lateglacial were more abrupt and with a higher degree of variability in FI scores than changes that occurred during the Holocene; which was more stable and with less variability. A comparison of the Shetland and Orcadian lochs suggest that the intensity of the ecological change was subtler at the Shetland sites. This could be due to Shetland being located at a higher latitude and lower / less severe temperature oscillations. For instance, shifts from cool to cold would be less extreme than from warmer to colder during the Lateglacial and early Holocene. Fisher Information has provided a regional comparison on the nature of changes that have occurred across the Northern Isles. The results here show that using the results of FI along with the zonation from a constrained cluster analysis provides more insight into how a system or assemblage changes through time. Detailed discussion about the individual FI results are discussed below.

5.3. Overview of the results from each of the study lochs

5.3.1. Loch of Sabiston

The results from Loch of Sabiston represent a multiproxy palaeoecological record that extends further into the Lateglacial than previous studies undertaken on Orkney. This is also the first high-resolution freshwater diatom record that extends from initial colonization of a newly deglaciated catchment to the mid-Holocene on Orkney. Based on the geochemical results, chronology, and the sediment stratigraphy, the sediment accumulation in Loch of Sabiston closely follows recorded major changes in temperature trends for the North Atlantic including the Bølling-Allerød, Younger Dryas and start of the Holocene (Boomer et al., 2012; Levesque et al.,

1993) (Fig.5.1). The results of the PCA analysis of diatom species show that species positively related to axis-1 are associated with clay and silt sediment, which corresponds to cold events in Loch of Sabiston, while negatively related species are associated with marl deposition, macrophytes, and warmer conditions. PCA axis-2 relates to the change from aquatic to a more terrestrial environment with a change in water levels and the beginning of peat deposition (Fig. 4.1.10).

Sabiston is a marl lake and its diatom assemblages have no strong analogues in the European Diatom Database (EDDI) for environmental reconstructions. Therefore, the diatom record is interpreted using the ecological preferences of known diatom species to reconstruct palaeolimnological conditions (Van Dam et al., 1994). This approach is coupled with the high-resolution geochemical analysis and the processes for the formation of marl sediments to support inferences regarding the environment and climatic conditions of the catchment. Marl lakes are highly alkaline, carbonate-rich, and precipitate calcite out of the water column due to photosynthetic activity including *Characeae*. Previous studies have not considered the role of marl deposition, and the circumstances surrounding its production in relation to environmental reconstructions on Orkney. The environment in which marl accumulates is quite unique and certain conditions must be met for it to precipitate. Temperature and pH are the main driving factors in marl precipitation along with the geology of the catchment (see section 2.4, Figs 2.5 - 2.7).

Marl precipitation tends to occur in the summer months during warmer temperatures, influenced by high productivity of aquatic macrophytes such as *Characeae* and *Potamogetonaceae* (Pentecost, 2009; Wiik et al., 2013), which dominate the macrophyte community in marl lakes. $CaCO_3$ and CO_2 is more soluble in colder water so the switching on and off of marl deposition on Orkney can be used as a proxy for temperature change as these systems are quite alkaline and a change in pH is not likely. In present marl lakes, the precipitation is initiated in the spring when the water temperature reaches above ~ 10ºC (Cole and Weihe, 2015). Using the results from the μ-XRF data and the Ca/Ti curve, which can indicate calcium derived from autogenic sources in the loch such as biogenic production from vegetation, the sediment sequence corroborates the change in temperature that occurred during the history of the loch. The onset of marl accumulation occurs when temperatures increased and the shutting down of the marl process occurred when temperatures returned to colder conditions. The main elements in this analysis that are highly correlated all correspond to in-lake processes and primary productivity (Table 4.1.3) and have very little relationship with catchment-based sediment input. There are peaks in Rb, Ti, K and Sr during the Lateglacial which correspond to detrital minerogenic input along with fine silt like particles which correspond to cooling events when marl production was shut down due to changes in the environment. These changes are also illustrated in changes occurring in the pollen and diatom record to an extent, which will be discussed below.

5.3.2.Loch of Clumlie

The record from Loch of Clumlie is complex, due to several drier phases which have likely resulted in hiatuses in the record at the top of the core. Inferences can still be made using the diatom, pollen, and elemental record to infer events such as the Bølling-Allerød, Younger Dryas and define the beginning of the Holocene. Due to the nature of each proxy, each responds at different temporal frequencies which show up in the palaeoenvironmental record, and how they responded will provide a better understating on the nature and timing of changes in the catchment at Loch of Clumlie. There are several instances where the geochemical profiles indicate changes that are not evidenced by the diatom or pollen records which illustrates the advantage of the high-resolution µ-XRF core scanning. The timing of changes in Loch of Clumlie recorded by the three proxies is summarised in Figure 5.2. Due to the number of uncountable diatom samples in the top portion of the core, the reporting of the data will be limited to the record below 78 cm as inferences made from above this point would be suspect (section 4.2.3).

The palaeoenvironmental history of Loch of Clumlie is dominated by fluctuations in effective moisture (precipitation – evaporation) leading to changes in the loch's water levels. During the Lateglacial, especially during the Bølling–Allerød, there is evidence of wet and dry periods along with evidence of water pulses into the loch based on the results of the diatom analysis evidenced by fluctuations aerophilous taxa such as *Achnanthes marginulata* and *Pinnularia biceps* (Fig. 4.2.8), and in the geochemical results evidenced by coarser grain size and increase in Si (Fig. 4.2.5). An increase in temperature and a sustained lowering of the water level marks the beginning of the Holocene which by ~ 60 cm core depth, led to oxidation of the sediments and a hiatus in the record while the spread of heathland across the site is indicated by the increase in *Calluna*. Loch of Clumlie presents a complex record that provides more insight into what has occurred during the Lateglacial on Shetland and provides evidence of how different areas in a small region can respond differently to larger regional influences.

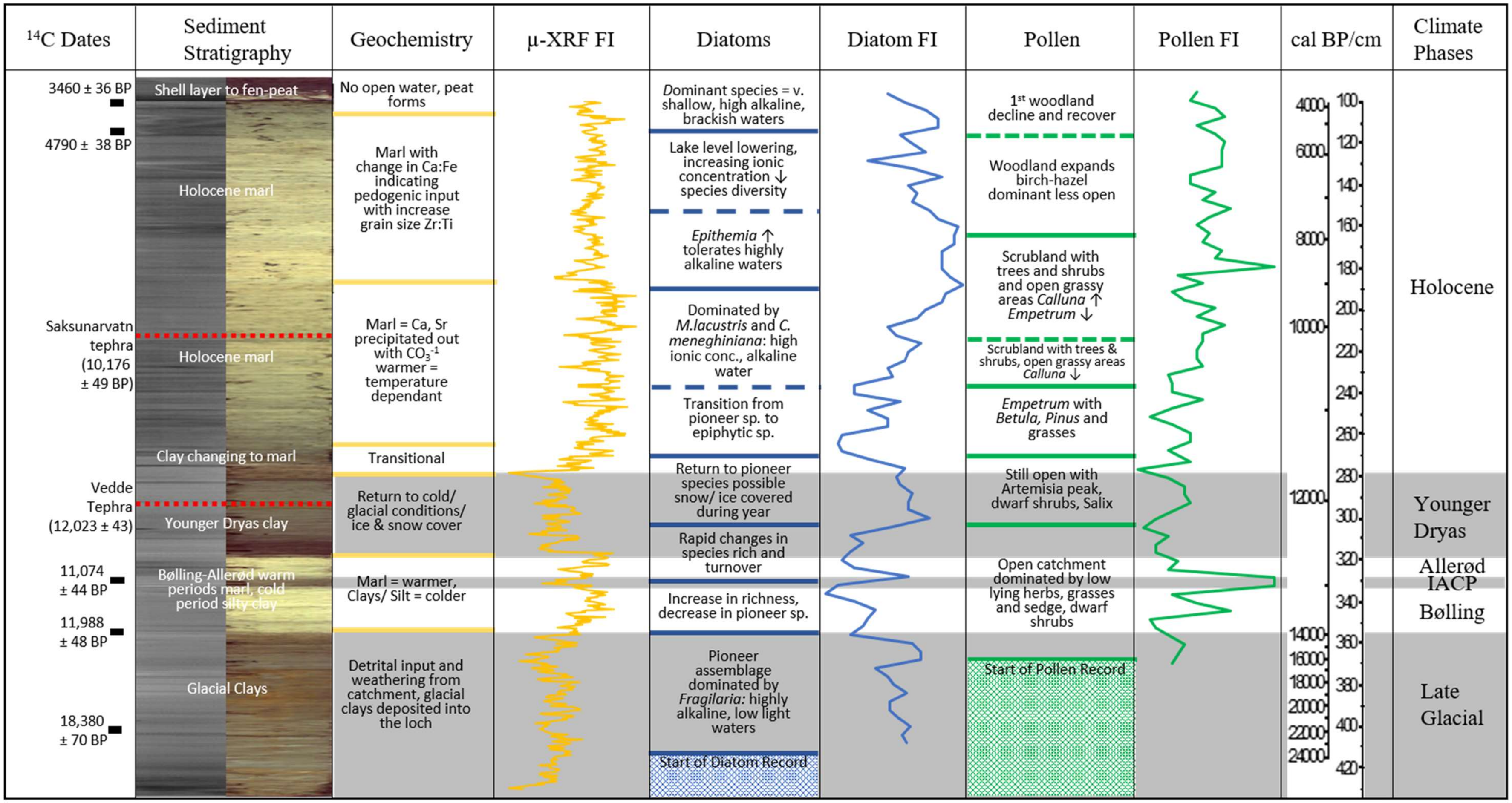

Figure 5.1 A schematic summary of the three proxies used from Loch of Sabiston illustrating the leads and lags that are evident throughout the record. The zones illustrated are based on the constrained cluster analysis of each record and the number of zones chosen were based on the results of Broken Stick analysis (Bennett, 1996).

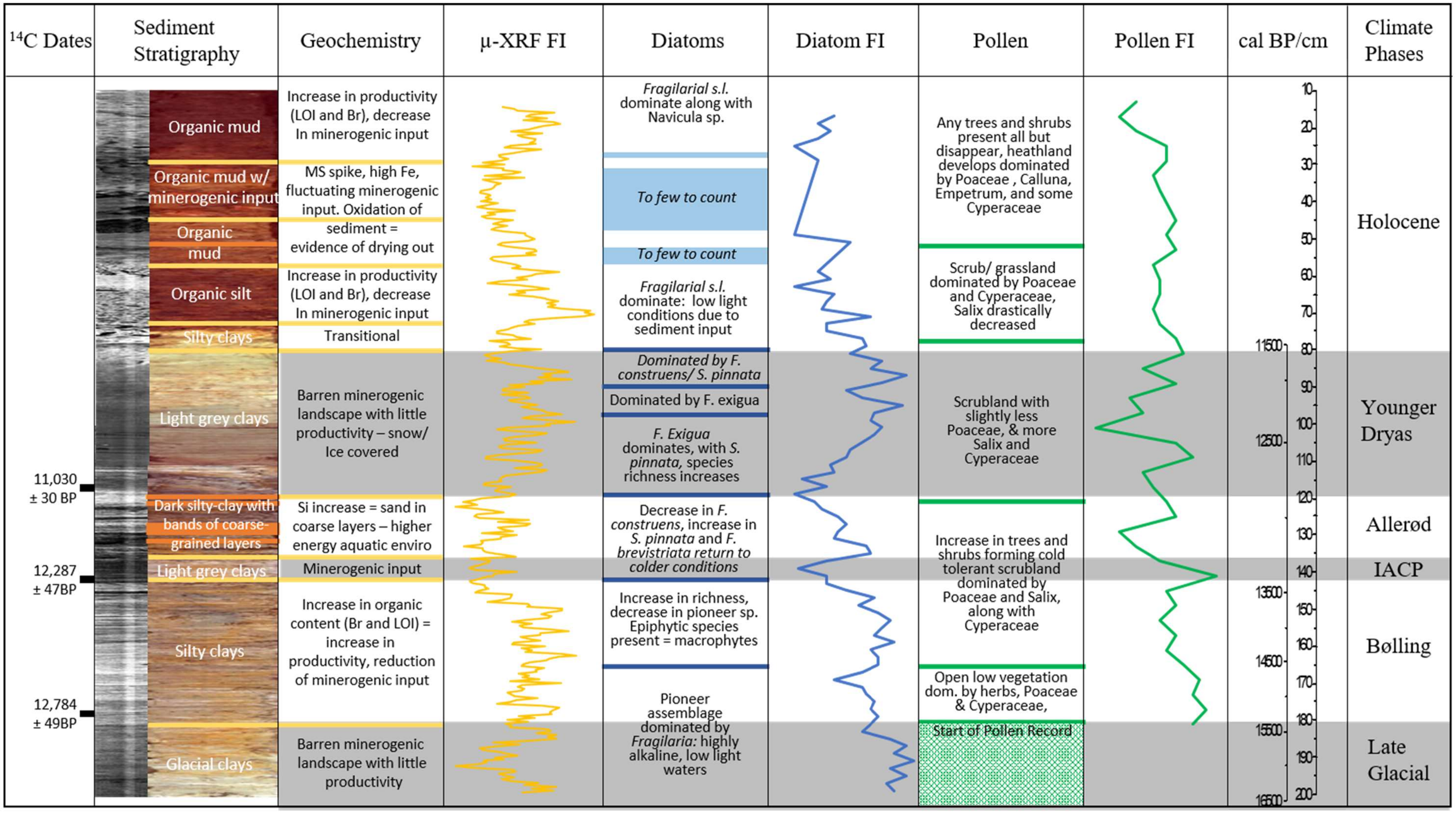

Figure 5.2 Summary diagram that consolidates the three proxies from Loch of Clumlie based on the results of the constrained cluster analysis (CONISS) illustrating the leads and lags that are evident throughout the record. The orange lines on the lithology define the locations of the coarse-grained/sand layers. To the left are the radiocarbon sample results and on the far right is the resulting age-depth model.

PCA axis-1 describes 36.7 % of the observed variation in species composition related to sample depth and PCA axis-2 explains 16.6% of species variation (Fig. 4.2.9). Axis-1 is negatively correlated to pioneer diatom species such as *S. construens*, *F, constricta*, and *A. pediculus* assemblage found in the base of the core containing clays and silty sediment. Axis-1 is positively correlated to species associated with small *Navicula* species where there is an increase in fresh water and the pH becomes more circumneutral (Fig. 4.2.9) and is associated with sites located at the top of the core (119 – 91 cm). PCA axis-2 is negatively correlated with diatoms species such as *P. elliptica, R. pusillum*, A. *minutissimum*. and *P. brevistriata*, which are predominately found in the organic component at the top of the core and are also associated with the diatom Zone CLMD-3 which correlates with the warmer Bølling – Allerød period (Fig. 4.2.9).

5.3.3.Loch of Grimsetter

The Holocene record from Loch of Grimsetter is characterized by a uniform record demonstrating subtle changes in the palaeoecological record. The records establish the progression of lake development highlighting the gradual changes in the aquatic and terrestrial environment. The climate on Bressay may be described as stable through the early Holocene with only subtle shifts in percentages occurring between plant groups in the terrestrial record (Fig. 4.3.10). Different influences in the environment define the zonation systems in both the pollen and the diatom records as there is no agreement in the timing of changes occurring (Fig. 5.3). The diatoms are consistently changing following shifts in the geochemistry while the pollen record seems to be changing prior to both with a distinct amount of time in between. This suggests, through most of the Holocene, the three proxies are responding to different influences in the environment over time.

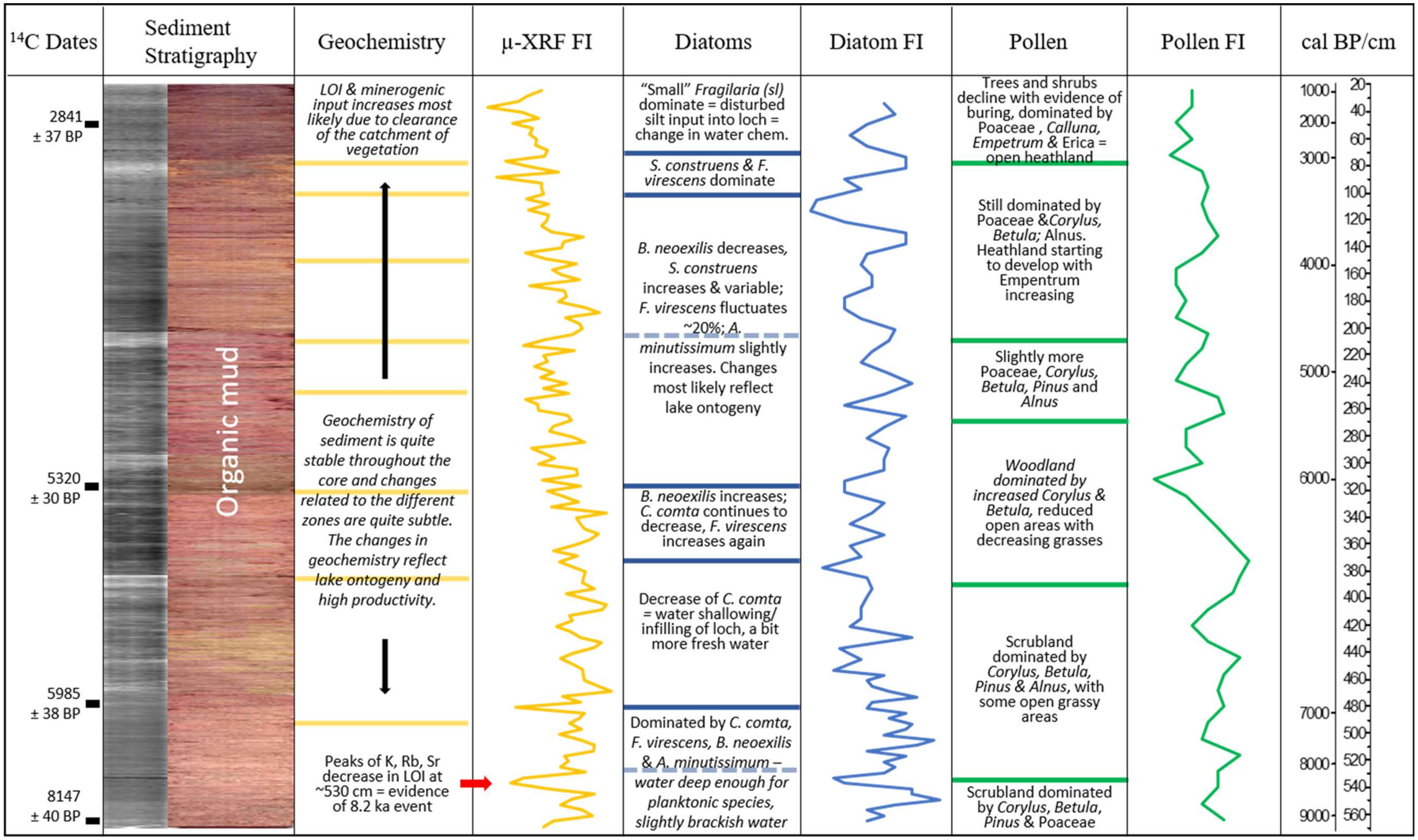

Figure 5.3 A compilation of the three proxies used from Loch of Grimsetter illustrating the leads and lags that are evident throughout the record. The zones illustrated are based on the constrained cluster analysis of each record and the number of zones chosen were based on the results of Broken Stick analysis (Bennett 1994).

5.4. Deglaciation after the LGM

5.4.1. Deglaciation of the North Atlantic

The retreat of the British-Irish Ice Sheet from the Northern Isles demonstrates a complex history in the rate of retreat and uncovering of land masses (Bradwell and Stoker, 2015). The warming at the end of the LGM caused rapid melting of the ice sheets covering the Nordic region. The way in which the pattern of sea ice retreat is still being debated and existing evidence suggests that Orkney was deglaciated *c.* 15,000 - 17,000 cal BP (Clark et al., 2012). Recently, older dates of deglaciation have been discovered such as on the Isle of Lewis (*c.* 25,000 cal BP) (Bradwell and Stoker, 2015). However, at Loch of Sabiston, the basal radiocarbon date of the core was determined to be *c*. 22,300 cal BP located ~ 30 cm from the base of the core, which extends the date of ice free conditions earlier than anticipated. This date is plausible in the light of recent reconstructions of the extent of the LGM (Bradwell and Stoker, 2015). These reconstructions suggest that Orkney might have been ice free at this earlier time if the pattern of deglaciation began to the west of Orkney, exposing land from west to east while the Fair Isle Channel to the north and the Pentland Firth to the South remained ice covered (personal comm. R. McCulloch). However, there is historically an issue with radiocarbon dating on Orkney and the influence of old carbon. The Lateglacial environment has not been well studied on Orkney and presently there are three records: Crudale Meadow, Quoyloo Meadow, and now, Loch of Sabiston, all located in western Mainland within 8 km of each other and only Loch of Sabiston has been independently dated.

Discussion on the nature and rate of deglaciation in Shetland after the LGM (Ballantyne, 2010; Bradwell et al., 2008; Bradwell and Stoker, 2015; Clark et al., 2012; Merritt et al., 2016) suggests that glacial retreat was complex in this region due to interactions between the British-Irish Ice Sheet and the Fennoscandian Ice Sheets (FIS). The oldest terrestrial dates for deglaciation after the LGM on Shetland are from Burn of Aith at *c.* 16,510 cal BP (13,680 ± 110 BP ^{14}C) (Birnie, 1993) and Lang Lochs at *c.* 16,181 ± 332 cal BP (Hulme and Durno, 1980), while sites further east in the North Atlantic, such as the west coast of Norway, do not provide ages older than the Bølling-Allerød (i.e. prior to *c.* 14,600 BP) due to the extent of the Fennoscandian Ice Sheet (Eldevik et al., 2014). At Loch of Clumlie, the basal age located ~ 25 cm from the base of the core is *c.*15,230 cal BP while at Lang Lochs, the basal age is *c.* 15,860 cal BP. However, a 30-cm bulk sample was used to obtain the age from Lang Loch and so the accuracy is less certain (Hulme and Shirriffs, 1994). The BRITICE project nevertheless concurs with these published dates for an ice-free Shetland (Golledge et al., 2008). Overall, there are few well dated Lateglacial records in the North Atlantic and subsequently the chronologies from Loch of Sabiston and Loch of Clumlie provide a significant advancement in the dating of deglaciation from the LGM in the North Atlantic.

5.4.2.Loch of Sabiston during the early Late Glacial Interstadial Transition

It is difficult to determine the basal age of the 432-cm core from Loch of Sabiston due to lack of organic matter in the basal sediment and therefore the earliest radiocarbon sample with adequate carbon is located at 403 cm which has an estimated date of *c.* 22,310 cal BP. The onset of the diatom record occurs at *c.* 23,830 cal BP (extrapolated) and the development of vegetation and the beginning of the pollen record occurs at *c.* 18,200 cal BP (Fig. 5.1). The geochemical record at the base of the core indicates an unstable, barren catchment denoted by the influx of minerogenic input caused by physical and chemical weathering of the catchment along with blueish grey clay/silt input which are most likely glaciogenic. This minerogenic input steadily decreases between the base of the core and *c.* 18,200 cal BP together with an increase in organic content which is typical of primary succession in recently deglaciated landscapes (Fritz and Anderson, 2013). The pollen record in Loch of Sabiston begins at *c.* 18,200 cal BP, just prior to the end of the LGM and contains species such as dwarf-shrubs, grasses, heath, and arctic-alpine species. The catchment would have been unstable arctic tundra, not very productive and probably had snow patches year round due to the presence of *Salix* (Wijk, 1986).

The diatom assemblage which begins at *c.* 23,830 cal BP has low diversity and is dominated by *Fragilaria* (*s.l.*) species which is typical of the first colonizers in recently deglaciated lakes (Haworth, 1976; Round, 1957). They are indicative of low light, low nutrient environments and persisted until *c.* 13,840 cal BP, the beginning of the Bølling-Allerød interstadial. These pioneer assemblages persisted for ~ 2500 years due to the colder temperatures and the minerogenic input into the loch. The low organic content and lack of macrophytes (observed during sampling for radiocarbon dating) during this time also indicates low productivity. Due to the periglacial environment, extended snow and ice cover during the year along with fine glacial clay particles in the water column would also have contributed to lower oxygen, together with extended low light conditions favoured by these pioneer species. The gradual change in species assemblage would be also due to changes in nutrient input into the loch from the stabilization of the catchment and soil formation rather than a response to changes in temperature or precipitation during this time (Fritz and Anderson, 2013).

The timing of the changes between the three proxies exhibit leads and lags in their occurrence (Fig. 5.1). The initial colonization of the catchment of both aquatic and terrestrial organisms was offset by ~ 5600 years with diatoms and macrophytes arriving first in the palaeoenvironmental record. The diatom assemblage appears to then change at *c.* 13,840 cal BP

(355 cm) while evidence in the geochemical record shows a change ~ 320 year earlier. Whereas, the pollen assemblage does not change significantly until after the beginning of the Younger Dryas at c. 12,270 cal BP.

Based on the results of the Fisher Information analysis, the stability of the ecosystem during deglaciation after the LGM is apparent in all three proxies studied from Loch of Sabiston. In the case of the pollen and diatom records, this is most likely due to the adaptation of cold tolerant species as even with warming it is still a periglacial environment. Gradual changes in stability occur towards the start of the Bølling – Allerød interstadial. The FI index shows change in the composition for the geochemistry starting at *c.* 14,500 cal BP, while the diatom FI begins to show slight instability in the assemblage at *c.* 15,400 cal BP with an increase in intensity of the shift from *c.* 14,700 cal BP, which is ~ 200 years prior to the change in geochemistry. Pollen FI index changes occur last with a subtle shift in FI at *c.* 14,900 cal BP, ~ 700 years prior to the onset of marl deposition at *c.*14,200 cal BP. This lag in the onset of marl precipitation is most likely due to the temperature not reaching the threshold to instigate marl precipitation. However, none of the changes occurring during the early phases of the LGIT and deglaciation could be considered to mark tipping points (which is determined by a sharp decrease in FI index) and only show gradual changes occurring prior to the full interstadial (Bølling - Allerød).

5.4.3. Orkney Islands during the early Late Glacial Interstadial Transition

At the two other Lateglacial sites on Orkney, Quoyloo Meadow and Crudale Meadow, the countable pollen record began just prior to the start of the interstadial (based on sediment stratigraphy) (Bunting, 1994; Whittington et al., 2015) while at Loch of Sabiston it began ~ 2280 years prior to the start of the interstadial (at *c.* 18,200 cal BP) however the chronologies at Crudale Meadow and Quoyloo Meadow are not as precise as the chronology from Loch of Sabiston. Crudale Meadow is correlated with the GICC05 dates (Whittington et al., 2015) due to problems associated with hard water error while previous attempts at dating Quoyloo Meadow (which also had problems with the calcareous nature of the underlying sediment) incorporated a radiocarbon sequence from Caithness along with the date of elm decline in Britain and the Saksunarvatn tephra layer for its chronology (Bunting, 1994). Loch of Sabiston shows a similar Lateglacial vegetation cover to Crudale and Quoyloo Meadow (Bunting, 1994; Whittington et al., 2015) indicating a sparsely vegetated and unstable landscape during the Lateglacial (from the beginning of the Bølling to the start of the Younger Dryas) with low-lying, open vegetation, such as grasses, and crowberry, sedges are typical with dwarf birch also thought to be present.

The diatom record from Loch of Sabiston is the first high resolution Lateglacial freshwater diatom record for the region and therefore, there are no other records on Orkney during the Lateglacial to compare to. Diatoms are recorded in sediments dating from *c.* 23,830 cal BP and consist of an assemblage dominated by pioneer species tolerant of low light, alkaline waters that are typical of a deglaciated environment that had prolonged ice cover throughout the year (Fritz and Anderson, 2013; Round, 1957).

5.4.4.Loch of Clumlie during the early Late Glacial Interstadial Transition

The basal age of the core from Loch of Clumlie has been estimated to be *c.* 16,400 cal BP based on a AMS calibrated radiocarbon date. The lithology and geochemistry of the basal sediment indicates an unstable, barren catchment denoted by the influx of minerogenic input caused by physical and chemical weathering along with fine silt and clay input. This input steadily decreases though the zone along with an increase of organic content which is typical of primary succession in recently deglaciated landscapes (Fritz and Anderson, 2013). Loch of Clumlie does not have an obvious visual organic layer that would reflect a warmer interstadial during the Lateglacial in the geochemical record.

The Lateglacial diatom assemblage at Loch of Clumlie is dominated by *S. construens* and *S. pinnata* and other *Fragilaria* (*s.l.*) species which is typical of the first colonizers in recently deglaciated lakes (Fritz and Anderson, 2013; Haworth, 1976; Round, 1957) and characteristic of an established but developing ecosystem after glacial cover in a low light, nutrient poor, alkaline environment. The high species diversity (compared to that of other Lateglacial sites with pioneer assemblages (Birnie, 2000; Haworth, 1976; Round, 1957; Smol, 1983)) and the increased presence of *Pediastrum* (Fig. 4.2.9), however, indicates a more developed successional stage which may suggest that the record does not include the earliest evidence for deglaciation. Due to the presence of a developed diatom community at the base of the core prior to the beginning of the interstadial (Fig. 4.2.8), Loch of Clumlie most likely experienced prolonged ice/ snow cover while the terrestrial environment was probably a less productive arctic tundra (Birnie, 2000).

The CONISS zonation for the three proxies shows variation in the timing of changes during the Lateglacial (Fig. 5.2). The geochemistry of the sediment changes at 180 cm (*c.* 15,340 cal BP) with an increase in organic content and silt signalling the beginning of the Bølling-Allerød interstadial, while the pollen record also begins at 180 cm. The diatoms do not show a significant shift in species assemblage until after the start of the Bølling-Allerød at 166 cm (*c.* 14,590 cal BP) along with an increase in diatom concentration. The time difference between the observed changes in the three proxies could be due to prolonged cold after deglaciation for longer periods of the year which may have resulted in a lag in response. Based on the ecological proxies (pollen and diatoms) the interstadial warming began at 166 cm (*c.* 14,590 cal BP).

The FI results of the geochemical analysis from Loch of Clumlie responded more abruptly than those observed at Loch of Sabiston. At *c*. 16,000 cal BP, the steep decrease in FI in the geochemical record (Fig. 4.4.3) marks a tipping point that corresponds to the change in sediment from being purely minerogenic to the beginning of increasing organic content along with a peak in magnetic susceptibility indicating an early transition in a landscape that is undergoing deglaciation (minerogenic sediments) and low lacustrine productivity and then with initial early LGIT warming an increase in lacustrine productivity and a rise in organic content.

During the remainder of the LGIT the FI index stabilized until ~ 100 years prior to the onset of the Bølling – Allerød interstadial (Fig 4.4.2) at which time FI increases indicating further changes in sediment composition. The results of the diatom FI analysis during the LGIT show a stable assemblage until *c.* 15,700 – 15,500 cal BP where there is a change in assemblage composition prior to the change in geochemistry and increase in organic content which could indicate increased productivity in the loch at this time. The pollen record from Loch of Clumlie does not begin until after the onset of the Bølling – Allerød interstadial and the FI index data sets are discussed in Section 5.5.4.

5.4.5. Shetland Islands during the early Late Glacial Interstadial Transition

Palaeoenvironmental evidence for the early Lateglacial on Shetland appear to be constrained to sites in the southern part of Mainland below Lerwick, with the exception of Grunna Water, which lies to the north (see Fig. 2.14, Chapter 2). The aquatic environment at Loch of Clumlie during the Lateglacial had a well-developed flora (unlike Loch of Sabiston where the base of the core was barren of diatoms) containing diatom species that are associated with macrophytes. These basal sediments also contained pioneer *Fragilaria* (*s.l.*) species that are tolerant of low light alkaline waters and *Pediastrum* was also quite abundant. The landscape was likely barren with very little organic content during the early Lateglacial based on the low concentration of pollen and the results of the LOI analysis (Figs 4.2.10 and 4.2.5). This juxtaposition between the productivity of the aquatic environment and the lack thereof in the terrestrial record is evident in the palaeorecords. Clettnadal has a similar pre-interstadial Lateglacial diatom record (*c.* 15,800 to *c.* 12,000 cal BP) but at a lower temporal resolution than Clumlie (Robinson, 2004), and is characteristic of other periglacial assemblages dominated by pioneer *Fragilaria* species (*s.l.*) (Birnie, 2000; Smol, 1983) which are typical of the first colonizers in recently deglaciated lakes (Haworth, 1976; Round, 1957) and are indicative of low light, low nutrient environments.

There is little detail regarding the vegetation during the early phases of the LGIT on Shetland as most pollen records begin just before the start of the Bølling – Allerød interstadial.

The terrestrial environment at Loch Clumlie is comparable to other records from Shetland such as Aith Voe and Clettnadal (Birnie, 2000; Robinson, 2004) which suggest low, herb-dominated vegetation tolerant of the unstable slopes, (glaciogenic regolith) and a cold arctic-alpine type environment.

5.4.6. Regional Synthesis for the early Late Glacial Interstadial Transition

In the North Atlantic, the literature suggests a similar successional pattern to that described from Loch of Sabiston and Loch of Clumlie. After deglaciation of barren unstable sediments were left, which were then colonized by open ground vegetation including dwarf arctic-alpine species (Walker et al., 1994). However, there are regional variations that have arisen that seem to be dependent on local conditions on Shetland. Similar to the findings from Loch of Sabiston and Loch of Clumlie, locations are affected by altitude, exposure, catchment characteristics, and geochemistry which can alter the timing of responses to change in the wider North Atlantic (e.g. Bradshaw et al. 2000; Brooks and Birks 2000a; Robinson 2004; Turner et al. 2015). The response of glacial retreat in Norway was later than that at the Northern Isles due to the continued presence of the FIS. At Kråkenes, Norway the pollen and diatom record showed similar successional patterns after the LGM, however this occurred during the timing of the Bølling – Allerød in the Northern Isles where the warmer temperatures had a greater influence on the newly deglaciated landscape. In southern sites early LGIT palaeoenvironmental records from Hawes Water (Brooks et al., 2012) and Thomastown Bog (Turner et al., 2015), suggest that sparse, open, herb rich terrestrial vegetation was established during the transition into the interstadial at *c.* 16,000 cal BP. Evidence for local conditions impacting on the timing of vegetation colonisation is seen at Whitrig Bog, where there was a delay in the response to warming due to persistent ice cover at the site during the early Lateglacial (based on stratigraphical evidence as there was a problem with radiocarbon dating) (Brooks and Birks, 2000).

The results from Loch of Sabiston and Loch of Clumlie provides insight into what these lacustrine environments were like just at the point of deglaciation which is what is underrepresented in much of the literature. This makes the present research new and the records incredibly valuable in terms of what can now be said about how these landscapes responded to the ice retreat and early colonisation.

5.5. Bølling – Allerød Interstadial

5.5.1. Characteristics of the North Atlantic Ocean

After glacial retreat, sea ice and increased fresh water entering the oceans from melting glaciers became strong influences on the North Atlantic climate (Ebbesen and Hald, 2004). During the Bølling-Allerød, there was an ice-free corridor between Shetland and Norway which allowed for warmer waters to influence the region, and therefore sea ice extent and density varied giving rise to cooling and warming events in the terrestrial climate records (Koç et al., 1993; Polyak et al., 2010). At *c.* 14,700 cal BP the oceans in the North Atlantic responded to the warming of the Northern Hemisphere, the AMOC resumed, driven by the release of fresh water from the melting glaciers, and warmer SSTs in the North Atlantic (Eldevik et al., 2014). However, this warming was not uniform across the Nordic Sea to Greenland and the North Atlantic (Rasmussen and Thomsen, 2008) with the maximum warming occurring at the beginning of the Bølling–Allerød (*c.* 14,700 cal BP) with cooling occurring during the final stages of the interstadial (*c.* 13,000 cal BP) (Eldevik et al., 2014).

5.5.2. Loch of Sabiston during the Bølling - Allerød Interstadial

A warming climate and the onset of marl precipitation signalled the commencement of the Bølling-Allerød interstadial at *c.* 14,160 cal BP at Loch of Sabiston. This interstadial is characterized by a series of warming and cooling events (Table 2.1) which are not always evident in Lateglacial records. The catchment of the loch is dominated by grasses, sedges and *Empetrum* species indicating open low-lying vegetation along with dwarf birch. The aquatic environment, represented by the diatom record, was very dynamic through the Bølling–Allerød at Loch of Sabiston and has corresponding subtle changes in sediment (Fig. 4.1.11). The pioneer diatom species that were present during the LGM, such as *Fragilaria* (*s.l.*) species decrease abruptly concurrent with the change in sediment from glacial clay to marl, while species diversity increased which included an increase in epiphytic species. The diatom community responded to a brief cooling event at *c.* 12,940 cal BP identified as the Intra Allerød Cold Period (IACP) evidenced by the change in sediment from marl back to a 5-cm layer of bluish-grey clays, likely due to the periglacial process of solifluction (Fig. 5.1), which suggests a return to a colder environment and increased ice cover. Diatom diversity slightly decreases and the return of post glacial colonizers occur during this colder period. However, a cooling event cannot be identified from the pollen assemblages due to the lack of countable pollen in this part of the core. There is however, a reduction in the pollen concentrations during this cold event. This could be due to prolonged ice cover during this cold period of the area along with the loch (Keatley et al., 2008; Whittington et al., 2015) or perhaps just reduced vegetation productivity due to the colder climate.

The switching on and off of the marl precipitation with the warming and cooling of the environment during the Bølling-Allerød is reflected in the elements associated with detrital input and physical weathering of fine grained sediment from the catchment (K, Si, Ti, Fe, Rb) and the elements associated with in lake productivity and marl deposition (Ca, Sr, Br) (Fig. 4.1.5). The high resolution geochemical record at Loch of Sabiston detects many of the cooling events during the Bølling - Allerød interstadial due to the sensitivity of marl precipitation to temperature changes (Kelts and Hsü, 1978; Pentecost, 2009; Wetzel, 2001). Changes in the minerogenic fractions (K, Fe, Rb, and Zr) along with Ca/Ti defined the cooling/warming events GI-1e (*c.* 14,700 – 14,000 cal BP), GI-1b (*c.* 13,900 – 13,100 cal BP), and GI-1a (*c.* 13,100 – 12,900 cal BP) (Fig. 5.1). GI-1d (*c.* 14,100 – 14,000 cal BP) is very subtle in the geochemical record from Loch of Sabiston at *c.* 13,700 - 13,590 cal BP which is slightly later than the Greenland ice core record of *c.* 14075 ± 169 cal BP. It is evident that the sediment composition at Loch of Sabiston is mediated by temperature changes during the Lateglacial.

During the Bølling-Allerød interstadial, the changes in FI were dynamic (Fig. 4.4.1). During the initial warming of the Bølling (GI-1e) the FI index indicates stability in the geochemistry of Loch of Sabiston. By *c.* 13,350 cal BP FI begins to decrease indicating instability in the geochemical composition ~ 450 years prior to the observed change in sediment during the IACP (GI-1b). There is some evidence in the FI diatom results that a drop in the FI index at *c.* 13,800 cal BP could indicate a change that would be associated with the cooling during GI-1d. However, there is no significant evidence for such cooling recorded in the three proxies and therefore this signal is ambiguous, but does highlight the fact that FI distinguishes the subtle nature of instabilities that can lead to changes in proxies by modelling the length of time a change takes based on the slope of the line in the graph. In both the diatom and pollen records, there are abrupt changes occurring within short times in between FI index changes, with some under ~100 years in length. At the onset of GI-1b cold event, the diatom assemblages rapidly changed indicated by sharp peaks, which is highlighted by the FI results and Bray-Curtis analysis. The diatom assemblages were very dynamic due to species turnover caused by the rapid changes in the environment at the transition between the Bølling-Allerød and the Younger Dryas.

5.5.3. Orkney Islands during the Bølling - Allerød Interstadial

The start of the interstadial at Loch of Sabiston is dated to *c.* 14,160 cal BP. Greenland ice core records were used to determine that the beginning of the Bølling at Crudale Meadow at *c.* 14,650 cal BP due to the problems and uncertainties of dating on Orkney (Whittington et al.,

2015) while at Quoyloo Meadow, the only reliable dates during the Lateglacial designate the beginning and end of the Younger Dryas (Bunting, 1994), leaving the start of the Bølling - Allerød interstadial undated. Quoyloo Meadow and Crudale Meadow show similar pollen assemblages and geochemical conditions during the transition out of the LGM into the Bølling-Allerød (Bunting, 1994; Whittington et al., 2015) as those found at Loch of Sabiston. The landscape on mainland Orkney during the Lateglacial consisted of open areas dominated by low lying herbs, grasses, and sedges along with dwarf shrub heath. The terrain consisted of glacial clays low in organic content and very minerogenic, which gradually increased in organic content progressing into the Bølling - Allerød. By the start of the interstadial, grasses, sedges, and willow colonized Orkney reflecting an arctic-alpine type environment with increasing organic content and stabilization of the catchment.

The pollen record at Loch of Sabiston displays an increase in grasses during the Bølling (at *c.* 13,500 cal BP), however, there is a gap in the record that extends from the end of GI-1e through to the IACP/ GI-1b (*c.* 13,000- 12,800 cal BP). This lack of countable pollen only occurs in the Sabiston record whereas at Crudale Meadow and Quoyloo Meadow the records are continuous, although it was noted that the preservation of the pollen was ‘pitted and thinned’ and concentration was very low during this period (Bunting, 1994; Whittington et al., 2015). This gap in the Sabiston record could mean that there was prolonged snow cover in this area which might have been the case as Loch of Sabiston is more inland than the other two sample sites (particularly since sea levels were ~ -40 m a.s.l. from present day).

Using the events outlined for this period from the Greenland ice core records (Rasmussen et al., 2014), the revised pollen record from Crudale Meadow is similar to that of Sabiston in that it showed no distinction between the Bølling (GI-1e) and following cold event GI-1d (Whittington et al., 2015), however, the IACP is present. There is also no differentiation between the IACP and the end of the Allerød in the pollen record. The low-resolution record at Quoyloo Meadow also finds no difference in the pollen assemblage through the interstadial until the start of the Younger Dryas (Bunting, 1994).

5.5.4. Loch of Clumlie during the Bølling – Allerød

The warming of the Bølling-Allerød period began *c.* 15,340 cal BP at the Loch of Clumlie as determined by the age-depth model produced with the Greenland ice core dates. Diatom species richness and evenness becomes stable while organic content keeps increasing which indicates that the aquatic environment is maturing and becoming more stable during this warm period. The geochemistry also indicates a more stable catchment with a levelling off the elements associated with minerogenic input (i.e. K, Rb, Fe) (Fig. 4.2.5). Bromine is higher during the early Bølling-

Allerød, which is an indicator of productivity as Br forms strong covalent bonds with organic molecules (Gilfedder et al., 2011). There is also an increase in diatom concentration that coincides with this increase in primary productivity, however the Si/Ti ratio does not reflect an increase in biogenic silica production (Brown, 2015). During this time grasses, dwarf willow, and sedges dominate the catchment vegetation along with herb species. This assemblage is characteristic of an arctic alpine environment with snow-loving *Salix* (Wijk, 1986) which continues until the start of the Younger Dryas. The results of the CONISS suggests that diatoms and geochemistry are changing together at the end of the Bølling-Allerød where the IACP begins at 142 cm (*c.* 13,310 cal BP) as well as the end of the Bølling-Allerød interstadial at *c.* 12,900 cal BP (Fig. 5.2). The pollen record shows no correlation with the end of the Bølling-Allerød but does start to change just prior to the diatoms and stratigraphy at the onset of the Younger Dryas. This slight offset could be due to the difference in the sampling resolution of the pollen.

There is a return to colder conditions at *c.* 13,310 cal BP which indicates the start of the IACP with the geochemistry responding first to the colder conditions with a return to minerogenic input along with a decreasing trend in organic content (Fig. 4.2.5). Bromine also decreases indicating a reduction in productivity in the loch. The diatom assemblage does not significantly change during the interstadial, however, the assemblages during the IACP are more comparable to the latter half of the Bølling-Allerød assemblages than those that were observed at the start of the interstadial. This indicates that the end of the interstadial was cooler than the start, which has been demonstrated elsewhere (Brooks et al., 2012). The pollen record does not show any significant change until 122 cm (*c.* 12,930 cal BP) around the beginning of the Younger Dryas (Fig. 4.2.10).

The geochemistry data suggests the end of the IACP began ~138 cm (*c.* 13,240 cal BP) (Figs 4.2.5 and 4.2.6) with a return to similar conditions to those found during the earlier half of the Bølling-Allerød, with reduced minerogenic input and an increase in organic content. The diatom assemblages have lower species richness and more variability between samples which indicate an unstable period. The diatom assemblage also indicates less brackish water was present (Fig. 4.2.8). The diatom species in this zone such as *Achnanthes marginulata* and *Achnanthidium subatomoides* are similar to those found in low to mid energy aquatic systems like pools in rivers, margins of streams and in backwaters (Patrick, 1996) which is consistent with the multiple layers of coarse sandy sediment bands that suggest a higher energy environment at this time causing coarse sediment to be washed into the loch from the catchment. The relationship of Zr/K has been shown to be a proxy for grain size (Fig. 5.6), with smaller numbers corresponding to finer sediment and larger numbers to coarser sediment which corresponds to the coarse sediment layers that are present in the Allerød. The Ca/Ti relationship (Fig. 4.2.6) can be used to determine hydrological variability where high values correspond with dry periods and low values with wetter conditions (Haberzettl et al., 2007) and also with autogenic and allogenic (detrital) lake productivity (Turner et al., 2015). Loch of Clumlie has variability in the Ca/Ti ratio through the Bølling-Allerød interstadial demonstrating periods of drier and wetter conditions. The Ca/Ti ratio corresponds to the coarse sediment inputs during the latter half of the Bølling-Allerød interstadial. This suggests that the catchment was at a hydrological tipping point between wetter and drier conditions and slight changes in the environment caused a detectable response.

The FI index response to environmental change in Loch of Clumlie during the interstadial are subtler than those found at Loch of Sabiston. Overall, all three proxies are quite stable through the interstadial despite the sediment geochemistry being highly variable. The occurrence of the IACP was characterized by smaller changes in the FI index (which reflects the intensity of the change occurring) than what was observed for Loch of Sabiston during this time. This response corroborates the fact that Shetland is more northerly and most likely did not experience as drastic temperature changes as more southern sites and therefore the response would be subtler.

5.5.5. Shetland Islands during the Bølling - Allerød

The evidence of the progression into the Bølling – Allerød at Loch of Clumlie was difficult to pinpoint due to the subtle responses of the proxies to warming and the uncertainties introduced with the age-depth model. The first location in the core that might correlate with the start of the interstadial has a date of *c.* 15,390 cal BP where the sediment consists of glacial clays and pioneer diatom species which is comparable to Clettnadal (*c.* 15,233 cal BP) (Robinson, 2004) and Aith Voe (*c.* 15,408 cal BP), although the record at Aith Voe has been adjusted to account for contamination due to old carbon which could make this date less reliable (Birnie, 2000). However, at *c.* 14,590 cal BP there are changes occurring in the record that more closely resemble those of a warming climate in both the diatom and geochemical records, which is more in line with the Greenland ice core records. It could be that this 800-year 'transition' evidenced in the data from Loch of Clumlie is due to the pattern of retreat of the BIIS and FIS, which could cause gradual warming on the eastern side of Shetland instead of a rapid response.

The pollen records from Shetland do not indicate a response to the warming and cooling events during the Bølling-Allerød (G1-1e to 1a). At Clettnadal, there is no evidence of change in the pollen record to mark the start of the interstadial while the record from Aith Voe suggests there is evidence for the beginning of the Bølling. However, there is no evidence of the subsequent cold and warm events that occurred during the interstadial. These observations also hold true for Loch of Clumlie, where significant changes in the pollen assemblage occur only after the beginning of the interstadial and at the start of the Younger Dryas. The pollen records at Grunna Water and Lang Lochs also do not seem to detect the IACP (GI-1b) but it is evident in the record from Clettnadal (Robinson, 2004; Whittington et al., 2003). This lack of sensitivity in the pollen record could be due to the position of the Shetland Islands in relation to the change in climate during this time. The species that dominate the assemblage are cold-tolerant and, therefore, a decrease in temperature would have not adversely affected them. It has been suggested that *Salix*, Poaceae, and *Rumex* would be the best indicators of stadial/ interstadial changes on the islands (Birnie, 2000). This is also observed in the record from Loch of Clumlie where arboreal taxa are not sensitive to stadial/ interstadial changes and changes in *Salix*, Poaceae, and *Rumex* were more indicative of the stadial-interstadial transition. The pollen records from Shetland show variation in their responses to climate change during the Lateglacial. It has been suggested that these changes could be due to a difference in altitude between the sites since the sea level was ~ 100 m lower than present sea levels (Birnie, 2000), and therefore a site such as Lang Lochs, which is presently 75 m a.s.l., is unique in that *Rumex* dominates, grasses were much lower in concentration than at other Shetland sites, and *Filipendula* was absent, suggesting that the high-altitude site was less stable and a harsher environment (Birnie, 2000).

Despite the lack of sensitivity in the pollen record during the Lateglacial on Shetland, the diatom records from Aith Voe, Clettnadal, and Loch of Clumlie appear to be more sensitive to climatic changes. However, the records at both Aith Voe and Clettnadal were completed at a very low resolution, therefore the possibility of observing subtle changes in the record may have been missed. At Aith Voe, the diatom record does not begin until after the beginning of the Bølling while at Loch of Clumlie and Clettnadal there are diatoms present in the glacial clays of the early Lateglacial. This suggests that at Aith Voe, there was significant snow and ice cover on the loch to inhibit diatom colonization during this time which was not the case at Loch of Clumlie. This caused the assemblage at Aith Voe to have a delayed response in colonization thus having an assemblage similar to that found in the LGM and not typical of an interstadial assemblage such as the one found at Loch of Clumlie. The diatom record at Loch of Clumlie responded to the changes occurring between the Bølling and the Allerød, however, the colder event of GI-1b (IACP) is not differentiated from the Allerød. This is similar to the results found at Kråkenes Norway. Both diatom records have persistent pioneer *Fragilaria* (*s.l.*) that extends into the Bølling – Allerød which is thought to be delayed lake development due to low summer temperatures and colder winters (Bradshaw et al., 2000).

Loch of Clumlie does not have a visible organic layer in the interstadial sediments similar to Lang Lochs (Hulme and Shirriffs, 1994). Hulme and Shirriffs (1994) thought that since Lang Lochs was further from the coast, at a higher altitude, and surrounded by hills, and that these conditions would allow organic content to increase only slightly without changing the appearance of the sediment due to the exposed nature of the loch. Loch of Clumlie has hills surrounding it however it is ~ 50 m lower a.s.l. and ~ 0.5 km closer to the coast than Lang Lochs. However, the lack of an obvious organic layer in Loch of Clumlie could also be due to persistent snow patches throughout the year limiting growth of vegetation in the catchment as opposed to exposure. The lack of countable pollen at the base of the core also supports this conclusion at Loch of Clumlie.

The geochemical record from Loch of Clumlie provides the most complete information regarding the changes occurring on Shetland during the LGM and Bølling – Allerød. Due to its high resolution, changes in the sediment content are quite apparent during the changes from warm to cold events of the interstadial. This is recorded in the results of the elements associated with minerogenic input, K, Rb and Ti where increases in the cps occurs during colder periods. The cold event associated with GI-1b is also evident (Figs 4.2.1 and 5.2). These results illustrate the usefulness of high resolution scanning of sediment cores in detecting short term climatic events where before they have been undetected.

5.5.6. Regional Synthesis for the Bølling–Allerød

The results from Loch of Clumlie and Loch of Sabiston demonstrate a sensitivity to temperature variability during the interstadial and there are subtle differences in the way that the two lochs responded to these changes. The geochemistry of Loch of Clumlie and Loch of Sabiston was the most sensitive of the three proxies in their response to the changing environment during the interstadial. In the case of Orkney, the islands seem to be at a location where the threshold for temperature driven marl precipitation occurs and, therefore, when there is a change in temperature during a cold event the marl formation ceases causing a change in sediment composition. The geology of Orkney has provided suitable conditions for the accumulation of marl sediments at Loch of Sabiston and therefore the switching on and off of marl deposition in response to climatic change provides a valuable temperature estimate for the North Atlantic region during the Lateglacial.

There is more variability observed between the sites on Shetland than between the sites on Orkney (this study and previous research) which all show a similar response to the changes occurring during the interstadial. On Shetland, there is more variation in the location of Lateglacial sites with Clettnadal located on the west coast, Lang Lochs is at an elevated altitude, and Spiggie Water which was the largest of the lochs and was formed from an inlet cut off from the sea. The site at Aith Voe is an infilled loch/ fen located in lowland Shetland while Loch of Clumlie adds to the heterogeneity of sites by being a small loch located also in the lowlands with a very dynamic past. On Orkney, the three Lateglacial sites are within 8 km of each other and respond similarly to the changes that underwent during the Bølling – Allerød in the pollen record. This regional variation has led to slight differences in the pollen assemblages between the two study sites on Shetland, and between Shetland and Orkney. For example, *Calluna* and *Empetrum*

(~ 30 %) is present in the Loch of Sabiston records as well as other records from Orkney during the interstadial (Bunting, 1994; Whittington et al., 2015) while they were a negligible component of the assemblage found at Loch of Clumlie and absent at other Shetland sites (Birnie, 2000). However, the Lateglacial pollen assemblage from Orkney is comparable to that found in Northern Scotland at Cross Lochs, Caithness (Charman 1994) which also did not respond to cold events during the Bølling – Allerød despite evidence in the lithology. The pollen record and to some extent, the diatom record did not respond to all three cooling events during the Bølling – Allerød as outlined in the Greenland ice core records (Rasmussen et al., 2014).

In the North Atlantic, the best evidence for temperature oscillations during the interstadial are provided by chironomid records and reconstructed summer temperatures along with the Greenland ice core records. In other Lateglacial records from southern Scotland, the demarcation of the IACP is usually not evident in the sediment stratigraphy. However, it does appear in some northern Scottish sites such as Shebster, Caithness (R McCulloch pers. comm.) and now on Orkney and Shetland. This could be due to the latitude in which these islands lie. This is in comparison to the records from Fiddaun, Co. Galway (Van Asch et al., 2012) where the marl is only interrupted during the Younger Dryas and did not completely shut off for lesser cooling events such as GI-1d and GI-1b. At Whitrig Bog, Loch Ashik, and Abernethy Forest GI-1d was colder than that of GI-1b while the reverse is found in Greenland (Brooks et al., 2012). Kråkenes, Norway chironomid inferred reconstructed mean July air temperatures were cooler than those found at the Scottish sites during the same period (Brooks and Birks, 2000). The results from Norway demonstrated a lag in the response to the warming of the interstadial with conditions resembling a pre-interstadial environment (Bradshaw et al., 2000). The diatom record from both Loch of Sabiston and Loch of Clumlie show that the assemblage for the cooling event GI-1d is not distinguishable from the Allerød. This suggests that the warming of the Allerød was not sufficient to significantly change the diatom community from what was present during GI-1d.

At Thomastown bog, another carbonaceous site, the relationship between Ca/Ti represents lake productivity and the biogenic production of calcium which is associated with marl production and is also meditated by temperature changes. Loch of Sabiston shows the same trends in the Ca/Ti ratio as Thomastown Bog with peaks during warm events, and it seems that marl lochs in the Northern Atlantic region are particularly sensitive to temperature changes regarding marl precipitation, and therefore demonstrate that chemo-stratigraphy can be used to track changes in climate variables that cause calcite to fluctuate with marl precipitation (Turner et al., 2015). The records from Thomastown Bog have been successfully correlated with $\delta^{18}O$ records from Greenland (Rasmussen et al., 2014). The $\delta^{18}O$ records from Crudale Meadow also show good agreement with the ice core sequence in the various warming and cooling events during the Bølling - Allerød interstadial including the IACP (GI-1b) (Whittington et al., 2015) and highlights a lag in the pollen response at Crudale Meadow to the cold events GI-1d and GI-1b compared to the ^{18}O records. The ^{18}O and μ-XRF investigations on Orkney provide crucial high-resolution information regarding the timing of changes occurring on the island archipelago. Most proxies do not distinguish GI-1d, but do detect the IACP on Orkney except for the δ^{18}O results from Crudale Meadow which detected both events (Whittington et al., 2015).

The results from chironomid inferred temperature reconstructions in marl lochs (Turner et al., 2015) has demonstrated that marl precipitation commences at ~ 10 - 12 °C. This consistent relationship between chironomid inferred temperature and marl production provides a benchmark to determine when temperature shifts above or below this threshold. This relationship is most evident in more northern lochs that are closer to the temperature transition than warmer, more southerly locations. Hawes Water (Marshall et al., 2002) shows this relationship, however, it does not detect the ICAP or any cold events during the interstadial. Irish carbonate sites (i.e. Van Asch et al. 2012) appear to be more sensitive to climatic changes but not to the same extent as Loch of Sabiston. The marl – temperature record from Loch of Sabiston provides a very useful (and high-resolution) temperature reconstruction for the region in determining temperatures above or below 10 °C.

5.6. Younger Dryas Stadial

5.6.1. Characteristics of the North Atlantic Ocean

In the North Atlantic the abrupt return to colder, glacial like conditions with extensive sea-ice cover and stable cold humid winters in Europe characterised the first part of the Younger Dryas, with the second half characterised by a less cold drier and a more unstable climate (Bakke et al., 2009; Baldini et al., 2015; Bradley and England, 2008; Coope et al., 1998; Ebbesen and Hald, 2004; Eldevik et al., 2014; Isarin et al., 1998). The differences observed between the first half and second half of the Younger Dryas are thought to have been due to changes in seasonality, sea-ice conditions, and atmospheric forcing (Brooks et al., 2016; Eldevik et al., 2014; Isarin et al., 1998). This return to glacial conditions triggered a re-advance of the Scottish ice sheet centred around Rannoch Moor (Benn et al., 1992; Bickerdike et al., 2016; Golledge, 2010; Hall, 2013) (The Loch Lomond Stadial). It is thought to have been caused by an influx of fresh water from the collapse of North American ice dammed pro-glacial lakes (Renssen et al., 2015; Teller et al., 2005) which caused a reorganization of the oceanic and atmospheric flow which in turn led to colder conditions across the Northern Hemisphere.

5.6.2. Loch of Sabiston during the Younger Dryas

The accumulation of bluish-grey clays and silts suggest a return to periglacial conditions at Loch of Sabiston and is further illustrated in the geochemical record where a return to increasing minerogenic sedimentation occurred (Fig. 4.1.5). This cooling signal is evidenced in the sediment and geochemical record at *c.* 12,580 cal BP, some ~ 300 years prior to changes in the pollen and diatom records. According to the ice core chronology (Rasmussen et al., 2014), the Younger Dryas began *c.* 12,900 cal BP.

The transition from the Younger Dryas into the Holocene has the diatoms responding to changes in the environment in relation to the warming occurring prior to the onset of marl deposition in the loch. This is also the case for the pollen record. This demonstrates that there is a threshold that had to be surpassed for marl precipitation to begin which is why there is a transitional zone of ~ 330 years detected in the geochemistry between these two periods. The lower boundary of this transitional zones has the highest sum of squares in the CONISS zones meaning that the most significant change occurred at *c.* 11,770 cal BP which would put this date as the start of the Younger Dryas, 260 years prior to the change in diatom and pollen assemblages.

The terrestrial vegetation record from Loch of Sabiston during the beginning of the Younger Dryas displays an open landscape dominated by grasses and sedges along with hazel and an increase in cold tolerant *Artemisia.* There is a reduction of pollen concentration with at least one sample containing insufficient pollen (296 cm; *c.* 12,120 cal BP) during the Younger Dryas. By the end of the stadial however, the vegetation composition changed with an increase of birch, grasses and *Ranunculaceae* and a decrease of low herb species such as *Artemisia* and *Asteraceae.* The aquatic environment of the loch also demonstrates a return to colder conditions in the diatom record where an assemblage resembling that of glacial conditions dominated by *S. pinnata* reappears which is indicative of high alkalinity and low light conditions caused by prolonged ice cover (Fritz and Anderson, 2013; Lotter and Bigler, 2000; Smol, 1988) (Fig. 5.1). By the middle of the stadial the aquatic environment has shifted and *P. brevistriata* and *S. venter* increase while *S. pinnata* decreases however still indicating a cold environment. The pollen and diatom record seem to be responding to the different phases that have been observed for the Younger Dryas with it being cold and dry during the first half and less cool and wetter during the end of the stadial while there is no evidence in the geochemical record for this shift in climate.

The types of changes occurring during the Younger Dryas at Loch of Sabiston as indicated by the FI demonstrates quite well in the diatom and pollen records the two phases that have been described during the Younger Dryas with two distinct decreases in FI along with distinct zones of change. Along with the CONISS zonation, there is good agreement regarding how the individual records are responding to changes in the environment. The geochemistry FI index is quite stable throughout the Younger Dryas after the initial (rapid) switch to a more minerogenic clay sediment at the commencement of the stadial. Whereas the diatom FI index shows initial instability at the beginning of the Younger Dryas and more stability in the latter half while the pollen also shows initial instability followed by a short plateau which occurred later then the start of the diatom stabilization in the second half. This discrepancy in the pollen response demonstrates the lag in the response to the change in climate during the Younger Dryas.

5.6.3. Orkney Islands during the Younger Dryas

The Younger Dryas stadial is well represented in the three Lateglacial records from Orkney. Based on the geochemical record, the Younger Dryas began at *c.* 13,080 cal BP at Loch of Sabiston, slightly earlier then the Greenland ice core records which record the beginning at *c.* 12,900 cal BP (Rasmussen et al., 2014). All three sites exhibit similar responses to the return to glacial conditions of the Younger Dryas with evidence of increased minerogenic sediment and decreased organic content in the catchment along with the cessation of marl deposition and the accumulation of bluish-grey clays and silts through solifluction. The condition of pollen during this event was found to be ‘pitted and thinned’ at Crudale Meadow and Quoyloo Meadow suggesting that the pollen was being redeposited from the catchment into the basin (Bunting 1994). At Loch of Sabiston, which is more inland, samples with insufficient pollen to count were from the Younger Dryas sediments (Appendix C1). During the Younger Dryas, the sea level would have been ~ -40 m from present levels making Loch of Sabiston more inland than Crudale Meadow and Quoyloo Meadow (Bunting, 1994; Whittington et al., 2015). The location of Sabiston could have been inland enough and more sheltered to have persistent ice and snow cover that would have supressed the growth of existing vegetation. Along with pollen and geochemical results, evidence for persistent snow and ice cover at Loch of Sabiston is also found in the diatom record with the return of a periglacial assemblage of pioneer taxa tolerant of low light conditions (Fritz and Anderson, 2013; Round, 1957). Other records have indicated that during the Younger Dryas, glaciers reformed on the northern part of Hoy (Gordon and Sutherland, 1993) and elsewhere on Orkney extended ice snow cover throughout the year was inferred (Bunting, 1994).

At Loch of Sabiston there seems to be a lag in the pollen response to the beginning of the Younger Dryas whereas, the pollen assemblage at Crudale and Quoyloo Meadows appear to change with the onset of the Younger Dryas (Bunting, 1994; Whittington et al., 2015). Again, this could be due to the location of Loch of Sabiston being further inland. The terrestrial vegetation from Loch of Sabiston suggests that the second part of the Younger Dryas was characterized by less colder and drier conditions than those during the first part of the Younger Dryas with birch and grasses increasing and *Artemisia* and sedges decreasing (Bakke et al., 2009). The pollen records at Crudale and Quoyloo Meadows do not reflect a vegetation response to this mid-Younger Dryas change in climate (Bunting, 1994; Whittington et al., 2015).

The diatom record from Loch of Sabiston is the only proxy that provides an indication of the aquatic environment during the Younger Dryas on Orkney. There is evidence that during the Younger Dryas the diatom record returned to an assemblage resembling that of glacial conditions dominated by *S. pinnata* which corresponds to high alkalinity and low light likely caused by prolonged ice cover as inferred for prior to the Bølling – Allerød interstadial (Fritz and Anderson, 2013; Lotter and Bigler, 2000; Smol, 1988). The mollusc record from both Crudale Meadow and Quoyloo Meadow was poorly preserved (O'Connor and Bunting, 2009; Whittington et al., 2015) and no individuals were recorded during the Younger Dryas. The lack of mollusc record at these sites could be explained by the diatom results from Loch of Sabiston suggesting prolonged snow and ice cover inhibiting mollusc persistence.

5.6.4.Loch of Clumlie during the Younger Dryas

The geochemical record indicates that the cooling during the Younger Dryas stadial began at ~ 120 cm (*c.* 12,900 cal BP). The loch sediment became more minerogenic with decreasing organic content. The pollen record indicates increases in *Salix* and Cyperaceae and to a lesser extent *Ranunculus* occurred due to the colder climatic conditions during the beginning of the Younger Dryas which favours dwarf willow and other arctic-alpine species such as *Ranunculus*. In the second half of the Younger Dryas, *Salix* begins to decrease and is replaced by grasses and at the top of the zone, *Salix* has virtually disappeared suggesting long lasting snow cover has been diminished (Wijk, 1986). The diatom record during the Younger Dryas encompasses three diatom zones (CLMD-4 to CLMD-6, Fig. 4.2.8) which illustrates the dynamic nature of the aquatic environment during this period. It has been suggested that the Younger Dryas began cold and wet and in the second half, became drier and slightly warmer (Alley, 2000; Benn et al., 1992; Isarin et al., 1998). This is reflected in the Clumlie record where in CLMD-4 diatoms such as *Pinnularia* spp. *N. molestiformis, S. anceps F. capucina* and *A. helivaticum* that are found in arctic and alpine waters with minimal ice cover (Smol, 1983). This is in step with CLMD-5 during the second half of the Younger Dryas which is said to have been drier and more unstable. Species richness decreases and *F. exigua*, which can survive in circumneutral moist to wet locations, dominates the diatom assemblage (Flower et al., 1996). Finally, by the beginning of CLMD-6 water depth deepens and a return of planktonic diatom species are present prior to the start of the Holocene occurring within ~ 126 years of the end of the Younger Dryas (Fig. 4.2.8). The results for Loch of Clumlie also demonstrates how hydrologically sensitive Clumlie is to a change in effective moisture that is reflected in the sediment composition and diatom assemblage.

At the end of the Younger Dryas, there appears to be a flickering climate signal prior to the shift into the Holocene recorded in the geochemical record (Fig. 4.2.5, 4.2.6). Most literature states that the transition from the Younger Dryas into the Holocene was abrupt (e.g. Alley et al. 1993; Tipping et al. 2012; Birks et al. 2012; Lane et al. 2013; Pearce et al. 2013; Rach et al. 2014), however, the records from Loch of Clumlie show over a period of ~185 years a decrease in the minerogenic elements occurred (i.e. K, Ti, Rb, Figs 4.2.5, 4.2.6) followed by a slight increase and then a final decrease associated with a colder environment. A similar signal in reconstructed temperature records has also been observed in chironomid data from Nova Scotia, Canada and other terrestrial and marine records surrounding the North Atlantic (Vincent and Cwynar, 2016). There are also corresponding increases and decreases in organic content, and in Si/Ti and Br, indicating variations in organic content, diatom concentration, and primary productivity which suggests a more gradual, step-wise transition between the Lateglacial and the Holocene at this site. The charcoal present at the end of the Younger Dryas (Fig. 4.2.10) has been recorded at other sites on Shetland (Edwards et al., 2000) and it has been suggested that it is due to the increase in aridity in the latter portion of the Younger Dryas due to reduced precipitation causing less breakdown of organic materials and conditions favourable to burning. It is also suggested that it could be due to long distance transport as is evident with *Pinus* pollen on the Northern Isles (Edwards et al., 2000).

5.6.5. Shetland Islands during the Younger Dryas

The palaeoenvironmental record from Loch of Clumlie provides a high-resolution analysis of the geochemical, terrestrial, and aquatic environment of an east coast, high elevation site and thus detected changes during the Younger Dryas that have not been recorded in the other Lateglacial records from Shetland. The pollen record responded to the start of the stadial and both the diatom and geochemical records have recorded changes during the Younger Dryas that are consistent with a shift from colder and wetter to less cold and drier conditions. However, the pollen record does not reflect this change.

The record of the Younger Dryas on Shetland reflects the individual nature of each of the Shetland Lateglacial sites demonstrating the regional variability in the response to the climatic cooling. Spiggie Water is 3 km West of Loch of Clumlie at ~ 1 m a.s.l while Aith Voe sits 13 km north of Loch of Clumlie at ~ 1 m a.s.l, both lower than Loch of Clumlie at 23 m a.s.l. This difference in altitude could be the cause of the differing records in this part of Shetland since during the Younger Dryas, sea levels were ~ -80 m from present day levels (Lambeck, 1991). The combination of being father inland (relative to sea levels) and slightly higher might have been enough to demonstrate these differences in climatic responses.

At Spiggie Water, Grunna Water, Clettnadal, and Loch of Clumlie (Birnie, 1981; Edwards et al., 2000, 1993; Robinson, 2004; Whittington et al., 2003), there is a distinct change in the sediment stratigraphy to clay marking the commencement of the Younger Dryas. However, at Aith Voe there is no real influx of minerogenic sediment into the basin which would suggest periglacial processes (Birnie, 2000). At Clettnadal the sediment during the Younger Dryas was comprised of silt and sand while the other sites consisted of mostly clay (Robinson, 2004; Whittington et al., 2003). This could be due to the location of Clettnadal on the west coast that is more exposed to the influence of the North Atlantic and high winds whereas the other sites are more sheltered on the east coast.

The Lateglacial records from both Lang Lochs and Aith Voe do not show a clear shift from the Bølling-Allerød interstadial to the Younger Dryas stadial in the pollen and diatom records (Birnie, 2000; Hulme and Shirriffs, 1994). The pollen and diatom evidence at Aith Voe (Birnie, 2000) demonstrates that terrestrial vegetation disappeared during the Younger Dryas while diatom palaeo-productivity was very low, however, epiphytic taxa were observed that would indicate some aquatic macrophytes were present. This may suggest that there might have been increased snow cover in the catchment (Birnie, 2000) but enough light entering the aquatic environment to support minimal productivity during the summer months. Loch of Clumlie also saw a reduction in productivity during the Younger Dryas but unlike Aith Voe, the diatom richness remained high during the stadial and there was no break in the pollen record such as at Spiggie Water and Aith Voe (Birnie, 2000, 1981). The diatom assemblage at Clettnadal is similar to Loch of Clumlie as it also divides the Younger Dryas up into several zones which could suggest a two phase Younger Dryas, however this evidence is based on 4 samples spanning the Younger Dryas (Robinson, 2004). Further evidence of a drier latter portion of the Younger Dryas is found in the charcoal record where there is increased charcoal where the diatom record indicates drying out of the basin which could make the catchment more vulnerable to fires (Edwards et al., 2000).

Though Spiggie Water is undated (Birnie, 1981), it displays similarities to the pollen record from Loch of Clumlie during the Younger Dryas. They both contain a pollen record dominated by grasses, sedges, willow, and *Artemisia*. However, at Aith Voe (Birnie, 2000) there is no evidence of terrestrial vegetation present during the Younger Dryas and it was concluded that the pollen present was long distance transport onto an ice-covered basin with a very low sedimentation rate (Birnie, 2000). Clettnadal, which is on the west coast also showed no distinct environmental change during the transition from the interstadial to the Younger Dryas with a pollen record dominated by willow and sedges and similar to the taxa found at Grunna Water (Edwards et al., 2000). On Shetland, Birnie (2000) determined that the relative changes in willow, grasses, and *Rumex* are the best indicators of changes between stadials and interstadials with *Rumex* either present or increasing during the stadial along with willow and less grasses contributing to the pollen taxa. This pattern also is evident at Loch of Clumlie with willow, grasses, and sedges dominating the assemblage during the Younger Dryas with *Rumex* present.

5.6.6. Regional Synthesis for the Younger Dryas

There are emerging subtle differences in the way that Orkney and Shetland have responded to the climatic changes that occurred during the Younger Dryas. Since the age-depth model developed for Loch of Clumlie is tied to the Greenland ice core chronology, specific timing of regional comparisons cannot be made. However, general comparisons can be made since most palaeoecological records from the Northern Isles that include the Younger Dryas show a change in sediment which can be used to constrain the event.

At Loch of Sabiston there was a return to a periglacial diatom assemblage and a decrease in species richness. This suggests that extended ice and snow cover resulted on Orkney. The shift from cold and humid to cool and dry is indicated with a change in dominant *Fragilaria* (*s.l.*) taxa, but the species present during the Younger Dryas remained constant (Fig. 4.1.9). In contrast, sites on Shetland with a Lateglacial diatom record (Birnie, 2000, 1993), provides evidence that the lochs did not return to a pioneer assemblage and that epiphytic species were present, however in low concentrations. This suggests that despite the return to a cold environment on Shetland there was still a sufficient amount of light entering the lochs to support the macrophyte and diatom communities. Loch of Clumlie, with its high-resolution diatom record, is the first on Shetland that confirms evidence for a two stage Younger Dryas event.

The transition from interstadial to stadial is apparent in the pollen record from Orkney, however some sites on Shetland do not show a difference in vegetation cover from the end of the Allerød to the Younger Drays. There is a slight difference in vegetation cover on the two island groups during the Younger Dryas. On Orkney pine, birch, and hazel dominate along with grasses and sedges while on Shetland willow, grasses and sedges are dominant. However, both island groups show a response to the two stage Younger Dryas climate. At Loch of Sabiston, Orkney, birch and grasses increase and *Artemisia* decreases in the second part, while Loch of Clumlie sees an increase in grasses with a decrease in snow-loving *Salix*. The FI index for Loch of Clumlie also demonstrates quite well in the diatom and pollen records for the two phases that have been described during the Younger Dryas (Fig. 5.2). However, the changes occurring at Loch of Clumlie are more subdued than those observed for Loch of Sabiston (Figs 5.1 and 5.2).

Both Loch of Clumlie and Loch of Sabiston show an increase in the minerogenic fraction of the sediment input into the lochs, along with a decrease in organic content and a return to clay deposition. As it was outlined in section 5.6.4, there is evidence at both Loch of Clumlie and Loch of Sabiston for a more transition-like progression out of the Younger Dryas into the Holocene in the geochemical record despite it being described as abrupt (Alley et al., 2003). With higher-resolution analysis being undertaken, especially with μ-XRF core scanning results, more evidence of a short-term transition period will become evident. This transition has also been observed in marine sediment records where a series of events might have occurred rather than one drastic event (Pearce et al., 2013).

Comparison of Greenland ice core records (Rasmussen et al., 2014) and northern European pollen records (Lane et al., 2013; Muschitiello and Wohlfarth, 2015) suggests that there is a time-transgressive shift during the beginning of the Younger Dryas across Northern Europe in the pollen records. It was demonstrated that there is a latitudinal delay of the onset of the Younger Dryas probably influenced by a gradual regional cooling from the south to the north (Lane et al., 2013; Muschitiello and Wohlfarth, 2015). The pollen record from Loch of Sabiston indicates that the beginning of the Younger Dryas commenced at *c.* 12,580 cal BP which is ~ 300 years earlier than that indicated by the pollen record from Kråkens, Norway (Lohne et al., 2014) and *c.* 320 years after the age of the beginning of the Younger Dryas determined by the Greenland ice core record (Rasmussen et al., 2014). Madtjärn, Sweden, which is latitudinally similar to Loch of Sabiston, has a mean age for the beginning of the Younger Dryas of *c.* 12,677 cal BP (Muschitiello and Wohlfarth, 2015) which is closer to the Orcadian age for the onset of the Younger Dryas. Subsequently, the chronology from Loch of Clumlie utilised the event chronological framework from Rasmussen et al. (2014), it cannot be confidently compared latitudinally to these sites.

Other Scottish sites that have a Younger Dryas pollen record show an east to west moisture gradient with sites in the east such as Abernethy Forest, having high incidences of *Artemisia* indicating aridity. There is a suggestion in the pollen record from Abernethy forest for two part Younger Dryas with *Pediastrum* and *Rumex* increasing in the latter part of the Younger Dryas and *Artemisia* decreasing (Birks and Mathewes, 1978). At Loch Ashik, there is a similar assemblage however a lower percentage of *Artemisia* suggests a wetter environment (Walker and Lowe 1991). However, the most comprehensive analysis of the temperature changes that occurred during the Younger Dryas in the North Atlantic can be found in chironomid records. Sites in Scotland and Northern England that have a Lateglacial chironomid record (Loch Ashik, Abernethy Forest, Hawes Water, and Muir Park Reservoir (Bedford et al., 2004; Brooks et al., 2016, 2012)) show variations in the way each location responded to changes in the Younger Dryas indicating regional differences in responses to the associated climate changes. Their results demonstrate that there is a definite division of the Younger Dryas into two distinct periods, however, each site responds differently based on regional variations. At Loch Ashik, the second half of the Younger Dryas becomes colder, whereas at Abernethy Forest there is evidence of warming which is similar to the results at Glen Roy which also showed warming at the end of the

Younger Dryas (Palmer et al. 2012), However, the later cooling recorded at Loch Ashik may have been due to its proximity to the Skye Ice Field (Brooks et al., 2012). At Hawes Water, comparable timing of temperature changes occurred with these other Lateglacial sites, however inferred slightly higher temperatures most likely due to the southerly nature of the site (Bedford et al., 2004).

5.7. Early Holocene

5.7.1. *Characteristics of the North Atlantic Ocean*

The transition from the stadial conditions of the Younger Dryas to the warmer interstadial of the Holocene was rapid, occurring *c.* 11,700 cal BP over a span of ~ 40 years (Oldfield, 2005; Taylor et al., 1997). This return to a warmer climate was due in part to maximum solar insolation occurring along with the stabilization of ocean circulation and the return of the AMOC bringing warmer waters from the south and driving sea ice extent northward allowing warmer Atlantic water to reach the region. However, this was an incomplete return from glacial conditions and the final deglaciation process (Mayewski et al., 2004) involved three cool events; the preboreal oscillation at *c.* 11,400 cal. BP, the 9300 cal BP event, and the 8200 cal BP event occurred during the early Holocene as defined from the Greenland ice cores (Rasmussen et al., 2014). The cool events at 9300 cal BP and 8200 cal BP are thought to be due to the still high and variable presence of sea ice and the polar front advancing and retreating southward (Mayewski et al., 2004). Post 8200 cal BP, the warming climate trend continues to the Holocene thermal maximum *c.* 7500 cal BP (Väliranta et al., 2015) after which temperatures began to decrease in the North Atlantic (Andersen et al., 2004).

5.7.2. *Loch of Sabiston during the early Holocene*

The start of the Holocene is marked by several rapid and high magnitude changes in the three proxies. The onset of the precipitation of marl suggests warmer temperatures are occurring while the geochemical record from Loch of Sabiston indicates that the sediment composition took ~ 350 years to change completely from the bluish-grey clay/silts of the Younger Dryas to the marl of the Holocene with the start of the change occurring at c. 11,950 cal BP (Fig. 5.1). This may have been due to a gradual increase in temperature or the result of continual lake development and infill through time (Fritz, 1989).

Macrophytes increased in the loch, especially charophytes, along with the presence of epiphytic diatoms and diatom species richness also increased. The vegetation record reflects the development of a more mature catchment with the increase of trees, shrubs, and herbs. This continued until c. 9120 cal BP when water levels and water chemistry changes affected the nature of the sediment and the diatom record. The decrease in the Ca/Fe, which indicates a change in sediment composition, is likely indicative of a decrease in authigenic carbonate precipitation in the catchment and therefore drier conditions (Mueller et al., 2009). During this time, the diatom assemblages decreased in diversity and evenness. Dominant species such as *G. angustatum and C. meneghiniana* are replaced by the brackish-freshwater, alkaline species *E. goeppertiana* along with *M. lacustis*.

The water levels in the loch at this point were probably lower and therefore became more concentrated in solutes. The lowering of the water levels can also be evidenced by the disappearance of planktonic *Cyclotella* species (Fig. 5.1) and the substantial increase in *Polypodiaceae* which may have flourished around the expanded margins of the loch. There is no equivalent change in the land pollen assemblage at this point in the core, however, between 179 and 175 cm (*c.* 8590 - 8400 cal BP) there was insufficient pollen to count in the samples which could be related to the conditions in between the 9300 cal BP event and the 8200 cal BP event when cooler and drier conditions occurred (Marshall et al., 2007). This lack of pollen could be due to lack of preservation, a reduction of pollen influx, or an increase in sediment deposition resulting in lower pollen concentration.

At *c.* 4000 cal BP the sediment stratigraphy and chemistry indicate a substantial lowering of the water level at the site and the switch to a terrestrial fen-peat. Manganese, which has been used to indicate oxidization of the sediment and to infer lower water levels indicating a decrease in water level. The diatom assemblage shifts to one that is dominated by few species while the pollen record sees an increase in charcoal, along with a slight decline in trees and shrubs (Fig. 4.1.9). The marl deposition switches off and a layer of *Lymnea sp*. shells formed indicating a near-shore environment at the coring location prior to the commencement of fen-peat deposition. (Fig. 5.1). The pollen record from Loch of Sabiston suggests that tree cover, comparable to the early Holocene, consisting of a birch-hazel woodland with a grass-fern understory persisted until after *c.* 3100 cal BP. The decline in tree cover has not been identified in the lacustrine part of the record and so likely happens within the top 100 cm of the fen-peat sediment. This evidence suggests a greater conservation of woodland resources in certain locations on the Orkney mainland than previously thought.

The correlation with the three proxies to each other seem to disconnect during the early Holocene with the diatom and pollen records appearing to respond to different environmental conditions. This could be due to the stabilization of the catchment and the overwhelming presence of marl precipitation in the loch. However, at the top of the record with the reduction of water levels, changes in the loch and in the catchment occurs prior to the final deposition of marl to be replaced by fen peat. The cessation of marl precipitation occurs *c.* 5500 cal BP is concomitant with the end of the Holocene thermal maximum and a shift to cooler summers (Geirsdóttir et al., 2013). Overall the general trend in the FI index is showing subtler changes occurring during the Holocene compared to the Lateglacial at Loch of Sabiston. There is slight variation occurring in each of the three proxies until a downward trend begins at *c.* 4900 cal BP which corresponds to the beginning in the change in sediment from marl to silt and then peat.

5.7.3. Orkney Islands

On land, the warming trends exhibited at the start of the Holocene translated into an increase in productivity and successional changes in vegetation across the North Atlantic. The transition from the Younger Dryas to the Holocene on Orkney was characterized by a return to marl deposition, a loss of the taxa associated with cold, open landscapes such as *Artemisia* and Asteraceae and an increase in diversity in both the aquatic and terrestrial taxa at Loch of Sabiston, Crudale Meadow and Quoyloo Meadow (Bunting, 1994; Whittington et al., 2015). These are the only three sites from Orkney that contain the transition from the Lateglacial into the Holocene. Thermophilous species increased such as *Betula* and *Corylus avellana* type while grasses continued to be an important part of the landscape.

Throughout the early Holocene the terrestrial succession of vegetation in the North Atlantic started with Ericaceous heath followed by replacement with tall herb taxa along with arrival of hazel and increasing birch. Vegetation succession on the islands is indicated by the development of woodlands and the corresponding decline in open landscapes. The process of woodland development on Orkney takes an east to west progression with sites in the east establishing birch - hazel woodlands at *c.* 9800 cal BP at Blows Moss (Farrell, 2009) while at Crudale Meadow and Quoyloo Meadow this occurred *c.* 8540 cal BP (Bunting, 1994), at Scapa Bay *c.* 9400 cal BP (de la Vega Leinert, 2007), and at Loch of Sabiston *c.* 7870 cal BP.

The decline of woodland across Orkney was not synchronous and probably occurred over several thousand years. On Hoy, this occurred from *c.* 7500 – 7000 cal BP, and at *c.* 6500 at Bay of Skaill while Glims Moss, Scapa Bay, and Loch of Knitchen showed woodland decline closer to *c.* 6000 cal BP. Quoyloo Meadow records a two stage decline with disturbance occurring at *c.* 7400 cal BP which reduced the tree cover and the later final decline at *c.* 5900 cal BP. While at Crudale Meadow, the final woodland decline occurred at *c.* 5400 cal BP. The region around Loch of Sabiston also showed variation with a reduction in tree cover occurring at Glims Moss at *c.* 5500 cal BP (3 km away) while at Burn of Rusht (6 km away) it was much later at *c.* 3800 cal BP. The record at Loch of Sabiston also appears to follow a two stage decline in arboreal pollen, however, the date of the initial disturbance is later at *c.* 5000 cal BP, some ~ 2400 years after the decline at Quoyloo Meadow. The date of the tree decline at Loch of Sabiston occurred sometime after *c*. 4000 cal BP. The latest dated tree disappearance occurs at Burn of Rusht *c*. 3590 cal BP which is near to both Loch of Sabiston and Glims Moss. However, Burn of Rusht lies at an altitude of 90 m a.s.l and is thought to have been a small relict stand (Keatinge and Dickson, 1979). At both Crudale and Quoyloo Meadow, there is evidence of a hiatus at the top of the marl deposition thus truncating the mid Holocene record. This is also plausible in Loch of Sabiston due to the shell layer indicating a 'near shore' environment or one that is wet/ moist that would be capable of drying out ceasing sedimentation (O'Connor and Bunting, 2009).

The change from lacustrine sediments to peat accumulation at the top of the record was during a time when climatic changes along with human influence impacted the region. There is evidence of a change to cooler and wetter conditions in Northern Europe *c*. 3000 cal BP (Bunting, 1996; Frenzel, 1966; Keatinge and Dickson, 1979). This may have led to waterlogging of soils and assisted in peat formation. The situation at Loch of Sabiston is similar to what occurred at Burn of Rusht, which still had trees present in the catchment prior to the commencement of peat formation (Keatinge and Dickson, 1979). However, the catchment at Loch of Sabiston during this time was in flux with changes in the geochemical record indicating an unstable catchment with an influx of minerogenic sediment and evidence of burning with charcoal increasing in the record.

At Loch of Sabiston there is evidence of a lowering of water levels and subsequent drying out at the transition from marl deposition to the development of fen peat accumulation. At Quoyloo Meadow there is evidence for a similar drying out at the transition from marl to peat formation with the recording of mollusc species that are associated with periodic desiccation (O'Connor and Bunting, 2009). At Crudale Meadow, the observed mollusc taxa are all generalist and associated with permanent water (Whittington et al., 2015) with no evidence of drying out.

The termination of marl deposition would have been triggered by the infilling of the loch of Sabiston and subsequent drying out as evidenced by the layer of mollusc shells on the top of the marl, just prior to the accumulation of fen peat. This caused an increase in nutrients and organic matter accumulation effectively changing the aquatic environment from one that stimulates marl production to one that is more acidic due to the fen peat. There is also more phosphorous available which would increase productivity based on the geochemical record which demonstrates the relationship between Ca (and marl production) and P and how during the process of marl formation P is sequestered in the sediments and is, therefore, not bioavailable (Pentecost, 2009). Following the hydroseral change from a lacustrine to a peat environment (which is also evidenced by the FI index indicating a regime shifty prior to the start of the CONISS zone SABP-5a), P became more available and along with evidence of grazing animals in the catchment, there were higher levels of P in the mid to Late Holocene compared to the LGM and early Holocene. However, the intensity of the P curve in the geochemical results (Fig. 4.1.5) indicates that phosphorous levels and eutrophication were not the cause of the marl deposition discontinuing

(Wiik et al., 2015a) and was most likely due to an increase in acidification along with the reduced water levels that occurred at the time.

5.7.4. Loch of Clumlie during the early Holocene

The warming that marks the start of the Holocene is identified at ~ 78 cm (*c.* 11,700 cal BP extrapolated) which is also very close to the end of the Younger Dryas (GS-1) (Rasmussen et al., 2014) at Loch of Clumlie, with a change in lithology to a more organic sediment and a reduction in minerogenic input. The top portion of the core is difficult to interpret due to the catchment drying out and the presence of one or more hiatuses in the stratigraphy. Therefore, the date of this transition is based on the Greenland ice core chronology (Rasmussen et al., 2014). There is compelling evidence in the diatom, pollen, and sediment stratigraphy that a hiatus might have occurred later (between ~ 50-25 cm), when the site may have dried out completely, marked by the development of an iron-pan (a terrestrial indicator) (Fig. 4.2.5, 4.2.6) and the lack of diatoms in that zone (Fig. 4.2.8). The condition of pollen preservation also deteriorates in this part of the core (Appendix C2). The ratio of Mn/Ti also reflects the development of oxic conditions and implies that oxygenation of the sediment has occurred by lowered lake levels (Kylander et al., 2011), along with an increase in Fe which results from detrital input and weathering (Fig. 4.2.6). The Ca/Ti relationship (Fig. 4.2.6) can also be used to determine hydrological variability where high values correspond with dry periods and low values with wetter conditions (Haberzettl et al., 2007), with extreme dry events evidenced during the beginning of the Holocene with maximum peaks occurring in the Ca/Ti record at Loch of Clumlie.

Despite the hiatus and the lack of an independent time line for this section of the core, some general inferences can be made regarding the history of the loch during this time. Between 78 – 52 cm, there is an increase in water level based on the number of green algae in the pollen samples along with the diatom record. Organic content also increases during this time as well as increased productivity and diatom concentration evidenced in the Br and Si/Ti signals respectively (Fig. 4.2.5, 4.2.6). Grasses and sedges are dominant in the catchment with the *Calluna* increasing in the record (Fig. 4.2.10). By 60 cm, peat formation has begun and any trees that were present have been drastically reduced. At this point, there is a clear hiatus due to the loch drying out completely and the loss of the diatom record. The pollen record also corroborates this in the increase in deteriorated pollen grains due to aerobic conditions (Fig. 4.2.10) and at least one sample with too few pollen to count at 40 cm.

5.7.5.Loch of Grimsetter during the early Holocene

The sediment record from Loch of Grimsetter, located on the island of Bressay, began in the early Holocene as a fresh to fresh-brackish, alkaline, oligotrophic loch (*c*. 9000 cal BP). Grasses, birch, and hazel along with some *Empetrum* and *Ranunculus* dominated the catchment, indicating a scrubby open woodland. This assemblage persisted until *c*. 5500 cal BP when grasses increased and *Plantago* spp. began to be more dominant. The increase in *Plantago* spp. has been associated with agriculture and land disturbance (Stolze et al., 2013) as it is photophilic and likes open spaces to which grazing contributes (Bonsall et al., 2002).

There is evidence in the pollen, diatom, and geochemical record for the 8.2 ka event with a return to a colder climate across Northern Europe (Rasmussen et al., 2014). The pollen record has a reduction in Poaceae and a slight increase in *Empetrum* while Polypodiaceae begins to decrease, which reflects the beginning of the event at *c*. 8300 cal BP. The geochemical record also indicates an increase in the elements associated with increased minerogenic input and a decrease in organic content probably in response to cooling and a reduction in productivity. At *c*. 8100 cal BP the Loch of Grimsetter diatom assemblage increases in species richness and evenness along with an increase in concentration which suggests a shift to a more temperate climate at the end of the 8.2 ka cooling event. The record from Loch of Grimsetter appears to be the first clear evidence for cooling lasting ~ 250 years, coeval with the 8.2 ka event on Shetland.

The most significant change occurred at *c*. 3040 cal BP in the pollen record at Loch of Grimsetter when trees and shrubs disappear to be replaced by *Calluna* and *Empetrum* heathland. At the same time, charcoal concentrations are at their highest in the record which suggests a likely human influence in the catchment. The burnt mound at Cruester on Bressay has been dated at *c*. 3000 cal BP which correlates with the Grimsetter charcoal record (Canmore, 2017). However, the most significant change occurred in the diatom record at *c*. 3330 cal BP with the increase in *Fragilaria s.l.* and small *Fragilaria*. Another major change also occurred at *c*. 2910 cal BP, ~ 150 years prior to the change in terrestrial vegetation, while the geochemistry recorded a slight increase in minerogenic elements at *c*. 3330 cal BP and a most distinctive change in sediment with a significant increase in minerogenic elements which was contemporary with changes in the pollen at *c*. 3050 cal BP. The influx of minerogenic material into the loch along with an increase in organics suggest increased disturbance in the catchment at *c*. 3050 cal BP.

Based on the diatom classification set out in Van Dam et al. (1994), the loch has been circumneutral to alkaliphilous and fresh to brackish during its existence. There does not seem to be any nutrient enrichment as oligotrophic diatom species dominate the loch record. The increase in undefined species at the top of the record is being driven by the small *Fragilaria* group which are prevalent in the assemblages. They can survive under a wide range of conditions but prefer higher nutrient levels and can survive lower dissolved oxygen levels in the water. Small *Fragilaria* (*s.l.*) are also known to tolerate low-light and fluctuating physical and chemical conditions (Griffiths et al., 2002; Punning and Puusepp, 2007).

The environment at Loch of Grimsetter has very little evidence of human influence in the catchment until *c*. 4500 cal BP and the increased presence of charcoal and *Plantago* spp. which indicates grazing animals (Stolze et al., 2013). The Loch of Grimsetter catchment remained a semi-open area with a combination of trees, shrubs, and herbs, while the loch remained a slightly alkaline, oligotrophic freshwater loch until *c*. 3000 cal BP when the catchment dramatically changed with a loss of trees and shrubs together with evidence of burning. Thereafter, the loch became more acidic and less oligotrophic and with a greater influx of minerogenic material into the loch reflecting the development of a less stable catchment. This may indicate greater population pressures on the isle of Bressay and an intensification of grazing pressures. The timing of changes between the three proxies are not necessarily in step with each other and there are considerable leads and lags between them which suggests that they are responding to different pressures which could be either human, climate or a combination of the two.

Loch of Grimsetter has a very stable FI index throughout the Holocene except for marked change in the sediment geochemistry (Fig 4.3.5) which is correlated to the 8.2 ka cooling event. This demonstrates the stable nature of the early Holocene climate on Shetland. By *c.* 6300 cal BP the FI index highlights a distinct change in the pollen record and did not stabilize until *c.* 5850 cal BP.

5.7.6.Shetland Islands

Due to the early to mid-Holocene hiatus in the accumulation of sediment at Loch of Clumlie the Holocene record from Shetland is provided by Loch of Grimsetter. The transition into the Holocene from the Younger Dryas was recorded in Loch of Clumlie *c.* 11,700 cal BP however at Grimsetter, the recovered record begins ~ 2600 years later at c. 9100 cal BP and does not include the Younger Dryas – Holocene transition. Other Shetland sites (Murraster, Lang Lochs, Clettnadal, Aith Voe (Bennett, 1993; Bennett et al., 1990; Birnie, 2000; Hulme and Shirriffs, 1994; Robinson, 2004; Whittington et al., 2003)) along with Loch of Clumlie have a distinct change in sediment occurring at the Younger Dryas – Holocene transition indicating an increase in organic content which is confirmed with the results of the LOI analysis. However, after the transition in the Holocene, there is evidence of dying out ~ 60 cm core depth. There is evidence that there was a hiatus a Grunna Water in the mid to late Holocene which could be related to the hiatus observed at Loch of Clumlie, however without confident dating in this part of the core it would be difficult to correlate the two.

After the initial increase in temperature and subsequent increase in bio-productivity at the start of the Holocene, Shetland palaeoenvironmental records start to deviate from each other and there tends to be more limited responses, likely driven by local conditions. As more records are obtained from Shetland, it is becoming clearer that the islands do not respond uniformly to changes in the environment during the Holocene and the record from Loch of Grimsetter is no exception. In the wider context of Shetland, the pattern of change at Loch of Grimsetter displays similarities and differences with other sites. Overall, herbs and shrubs dominated the vegetation of Shetland in the early Holocene alongside an open woodland (Hulme and Shirriffs, 1994). The dominant herbs tended to be grasses and sedges which is comparable to the vegetation that surrounded Loch of Grimsetter. However, as the Holocene progresses spatial differences developed between the various sites due to geographical factors along with the timing of changes that occurred. For example, while birch appears early at Dallican Water (Bennett et al., 1992), Lang Lochs (Hulme and Shirriffs, 1994), Murraster (Bennett, 1993) and Loch of Bunatwatt (Edwards and Whittington, 1998), at Clettnadal an Atlantic maritime grassland is formed which could be due to it being an extreme costal site on the west directly facing the North Atlantic (Whittington et al., 2003).

Several benchmark events that can be tracked across Shetland also demonstrates variation in the timing of these events. The rise of *Calluna* and the beginning of heath across Shetland ranges from early appearance at Lang Lochs *c.* 8500 cal BP (Hulme and Shirriffs, 1994) and the latest at Grimsetter *c.* 3000 BP (Table 5.1). *Plantago lancelota* has been used as evidence for the presence of grazing and disturbance (Bonsall et al., 2002; Stolze et al., 2013). At Grimsetter abundances of *Plantago lancelota* do not reach greater than 2 % in the pollen record, whereas at other sites it makes up a larger constituent such as at Clettnadal (~ 20%) which can indicate that Clettnadal had higher grazing pressures than other sites. Lang Lochs seems to have the latest arrival of *Plantago lancelota* while the earliest occurrence is at Loch of Brunatwatt (Table 5.1). Shetland has many early Holocene sites that are situated in areas with regional differences such as altitude, distance from the sea, vegetation, and bedrock, and the palaeoclimatic signal becomes more complex. Are these regional variations in the palaeoenvironmental record strong enough to mitigate the influences of changes in the marine environment or is it that the human influence is more pronounced with the addition of grazing animals and clearing of the land promoting increased wetness, burning, and podsolization instigating the increase in *Calluna* and peat formation during the early Holocene?

The diatom record at Clettnadal is the only other well dated Holocene record on Shetland and it is difficult to compare the catchments of Grimsetter and Clettnadal. The latter is located on the west coast with a strong marine influence in the diatom record (Robinson, 2004). Loch of Grimsetter is located on the east coast and appears to have received negligible impacts from the sea. The record at Clettnadal is also at a lower resolution which makes it difficult to pinpoint the precise timing of environmental changes recorded. There is a need to study more sites on Shetland to get a more complete picture of the timing of change that occurred on the Shetland Islands in both the diatom and pollen records. The environmental changes recorded in the diatom record at Grimsetter are quite subtle which suggests a stable environment though the early Holocene (Fig 4.3.8). The more substantial changes in the catchment are identified in the pollen and geochemical record at *c.* 3000 BP and were most likely human induced (Fig. 4.3.5, 4.3.10).

Table 5. 1 The differences in the timing of key events in the pollen records of the Shetland Islands. All dates have been calibrated using IntCal 7.1 (Stuiver et al. 2017) and are in cal BP from: [1]Hulme and Shirriffs (1994)**;** [2]Bennett et al. (1992)**;** [3]Bennett (1993)**;** [4]Whittington et al. (2003) and Robinson (2004)**;** [5]Edwards and Whittington (1998)**.**

Event	Loch of Grimsetter	Lang Lochs[1]	Dallican[2]	Murraster[3]	Clettnadal[4]	Loch of Brunatwatt[5]
Calluna increase	c. 3000	c. 8500	c. 5100	c. 5400	c. 3760	c. 5820
Plantago increase	c. 5500	c. 3180	c. 4400	c. 5400	c. 3760	c. 5740
Betula decrease	c. 3300	c. 5240	-	c. 5400	c. 10,370	c. 3770
Corylus avellana type decrease	c. 3000	c. 3300	-	c. 3360	-	c. 3770

5.7.7. Regional Synthesis for the Holocene

The warming and the stabilization the Atlantic Ocean underwent in terms of circulation and SST during the Holocene allows sites in the North Atlantic to respond more locally to any changes in the environment that may be occurring. During the Lateglacial, the terrestrial environment of the northern North Atlantic was most likely at the edge of tolerable ranges (*i.e.* temperature, nutrients) for many species and therefore succession would be constrained by severe climate and short summers which would reduce survival and slow advancement (Svoboda and Henry, 1987). Tipping (1994) outlines the general succession of woodland development and decline since deglaciation throughout Scotland and the Northern Isles and demonstrates the complex migration of trees from refugia from the south over several thousand years with Orkney and Shetland recording the latest changes in this regard. At this point in the Holocene, the human influence on the landscape is becoming more prevalent in the southern North Atlantic and can obstruct the climatic signal in palaeoenvironmental proxies. Therefore, this summary will be constrained to Shetland and Orkney only as they were more isolated and have a later date of human colonisation and an in-depth analysis of human influence on the landscape is outside the scope of this research.

In the Northern Isles, changes in the terrestrial and aquatic environments tend to be subtler during the Holocene compared to the Lateglacial, however no less significant on local scales, depending on the site. Loch of Grimsetter and Loch of Sabiston have different aquatic histories throughout the early Holocene. Grimsetter, on the Island of Bressay, is quite stable in all three proxies and is not responding to perturbations that have been recorded in other Shetland sites, but does show evidence of the 8200 cal BP event in the geochemical record which has not been recorded in either the pollen or the diatom record. Loch of Sabiston in comparison, has more variability recorded in its proxies during the early Holocene reflecting changes in the catchment including succession of terrestrial vegetation and lake ontogeny along with climatic changes. There is also evidence that terrestrial and aquatic proxies begin to respond independently of each other reflecting possible different forcings on the catchment such as lake ontogeny or succession in the catchment that would not necessarily react at the same time. Loch of Grimsetter and Loch of Sabiston both exhibited a stable FI index until *c.* 6500 cal BP at Sabiston and *c.* 6300 cal BP at Grimsetter where the FI index highlights a distinct change in the pollen record and did not stabilize until *c.* 5950 cal BP and *c.* 5850 cal BP respectively. However, neither of these time periods correspond to any known climatic events in the region, but are most likely related to human influences on each of the island groups.

The detection of cold events outlined in the Greenland ice cores during the early Holocene have been problematic in some proxies such as pollen in the North Atlantic. These cooling events such as the preboreal oscillation, the 8200 cal BP event and the 9300 cal BP event are thought to be caused by freshwater inputs into the North Atlantic during the final deterioration of the Laurentide Ice Sheet in North America (Nesje et al., 2004). However the evidence for these events are difficult to detect in more northern sites (Björck et al., 1997). This holds true for sites on Shetland where evidence for the 8200 cal BP event shows up in the geochemical record for Loch of Grimsetter. At Loch of Sabiston, there is no evidence in any of the proxies for the 8200 cal BP event.

Chapter 6

6. Conclusion

This study set out to determine the synchroneity of multiple proxies (diatoms, pollen, and geochemistry), in reconstructing past environments in the Northern Isles to determine the nature and timing of climatic and landscape changes. This study also sought to identify evidence of 'warning signals' in the records as precursors to regime shifts in the environment. Despite Orkney and Shetland being in the same phytogeographical climatic region, evidence has been shown that there are differences in the ecological responses of the study lochs and surrounding catchments on these two archipelagos to the nature and timing of Lateglacial and Holocene environmental changes.

The research undertaken here has demonstrated the relationship between terrestrial, aquatic and catchment based proxies and how they respond individually to changes in the environment in the Northern Isles. There are leads and lags that have been identified between the responses of the proxies that can differ by several hundred years, along with instances when proxies are synchronous. Therefore, constraining the timing of specific events based on one proxy can be problematic if it doesn't take into account the nature of the proxy as demonstrated by this study. At Loch of Sabiston and Loch of Clumlie, there is more synchroneity between the proxies during the Lateglacial than found during the Holocene, likely due to the higher magnitude of the climatic changes during the LGIT.

Applying Fisher Information to multiproxy palaeoenvironmental data demonstrates that it can be a useful and valuable tool in exploring the nature and timing of environmental changes that have occurred over time. It establishes when assemblages begin to shift and become unstable and not just the point at which the proxy becomes significantly different from an earlier time as is the case when using constrained cluster analysis and Broken Stick methods. It also provides information on when destabilization begins in a system, and when used together with constrained cluster analysis, FI provides a greater understanding of what is happening through time with individual proxies.

This research has provided new evidence for the nature of the palaeoenvironmental changes across the Northern Isles after deglaciation to the early Holocene. It has been shown that Orkney was free of ice at *c.* 23,830 cal BP; ~ 4000 - 6000 years earlier than what has previously been recorded. While this age appears anomalously older, there is no reason to reject it and therefore predicting a prolonged LGM into the Bølling – Allerød. Evidence of deglaciation from Loch of Clumlie corroborated other dates from Shetland at *c.* 16,000 cal BP. However, since Loch of Clumlie is not independently dated and is bound to the Greenland ice core chronology, further analysis to search for tephra to use to constrain the age-depth model rather than the Greenland ice core dates would be the logical next step in developing an independent chronology.

The ice retreat exposed a minerogenic landscape with little to no organic content at both sites. The dating of ice free conditions from Loch of Sabiston and Loch of Clumlie provides important insight for ongoing research on the timing and nature of the retreat of the BIIS. The palaeoecological record from Orkney begins at *c.* 23,830 cal BP with the initial colonization of the loch with pioneer diatom taxa followed by open, cold-tolerant vegetation species ~ 5600 years later. On Shetland, this process began at Loch of Clumlie prior to *c.* 16,000 cal BP in the aquatic record however, unlike Sabiston, the diatom assemblage was more mature with the presence of epiphytic species, along with taxa considered to be pioneer or arctic assemblages suggesting that the margins of the loch became ice free during the summer allowing for macrophytes to develop while maintaining a low-light cold environment for the remainder of the loch (Smol, 1988). The surrounding landscape at Loch of Clumlie remained mostly barren of vegetation during this period with evidence of terrestrial plants not present until the start of the interstadial. The diatom records from Loch of Sabiston and Loch of Clumlie provide greater insight into the lake environment at the point of deglaciation which hitherto have been neglected in the literature for the Northern Isles. This study provides a better understanding of the differences that are occurring in how the aquatic environment and the landscape responded to ice retreat and initial colonization. These differences are illustrated in Figure 6.1 which shows the timing of the climatic changes in the three study sites in comparison with $\delta^{18}O$ Greenland ice core records (Grootes et al., 1993).

The results of the palaeoenvironmental reconstructions during the Bølling - Allerød interstadial in the Northern Isles is quite complex in terms of disentangling the detection of the warm and cold events which have occurred (Fig. 6.1). The beginning of the interstadial was at *c.* 14,160 cal BP at Loch of Sabiston and on Shetland *c.* 15,300 cal BP based on Loch of Clumlie along with other Shetland sites (Birnie, 2000; Robinson, 2004). The initial warming of the interstadial resulted in grasses, sedges, and willow colonizing Orkney reflecting an arctic-type environment with increasing organic content and stabilization of the catchment while on Shetland there was a comparable assemblage, however lacking the *Calluna* and *Empetrum* which was present on Orkney. The Loch of Sabiston diatom community responds to the brief cooling event of the Intra Allerød Cold Period (IACP) which suggests a return to a colder environment and increased ice cover while at Loch of Clumlie the diatom assemblage does not significantly change during the interstadial providing evidence that the cooling occurring on Shetland was less influential on an already cold environment.

The Greenland ice core records indicate a series of cooling events during the Bølling - Allerød interstadial designated GI-1d, GI-1c2, and GI-1b. These events are not always evident in other palaeoecological records such as pollen in the North Atlantic region (Fig. 6.1), however chironomid and $\delta^{18}O$ records suggest periods of cooling contemporary with these events. In other Lateglacial records from southern Scotland, the demarcation of GI-1b (IACP) is usually not evident in the sediment stratigraphy. However, at Loch of Sabiston, Orkney, the IACP was identified from *c.* 13,900 – 13,100 cal BP, and is most distinct in the geochemical record which suggests that the sediment composition was strongly driven by temperature changes during the Lateglacial period. Loch of Clumlie also has evidence in the geochemical record for cooling during GI-1b from *c.* 13,310 - 13,240 cal BP. However, in the diatom records from both Loch of Sabiston and Loch of Clumlie the cooling event GI-1b is not distinguishable from GI-1a. This suggests that the warming of the Allerød was not sufficient to significantly change the diatom community from what was present during GI-1b.

The return to glacial-like conditions marks the Younger Dryas stadial. At both Loch of Sabiston and Loch of Clumlie there was a return to the accumulation of bluish-grey clays and silts and a decrease in bio-productivity. At Loch of Sabiston the precipitation of marl correspondingly ceased. At Loch of Sabiston arctic/alpine diatom assemblages returned, while the pollen record indicates a shift to more cold tolerant taxa and the loss of warm-temperate vegetation and increase in algae, most likely due to increased snow and ice cover in the catchment with periods of open water around the margins of the loch allowing for aquatic productivity. The terrestrial response at Loch of Clumlie was less extreme with an increase of willow along with grasses and sedges. This muted response could be due in part to the fact that warm-temperate vegetation never really established on Shetland during the interstadial and therefore the shift to stadial conditions would be subtler.

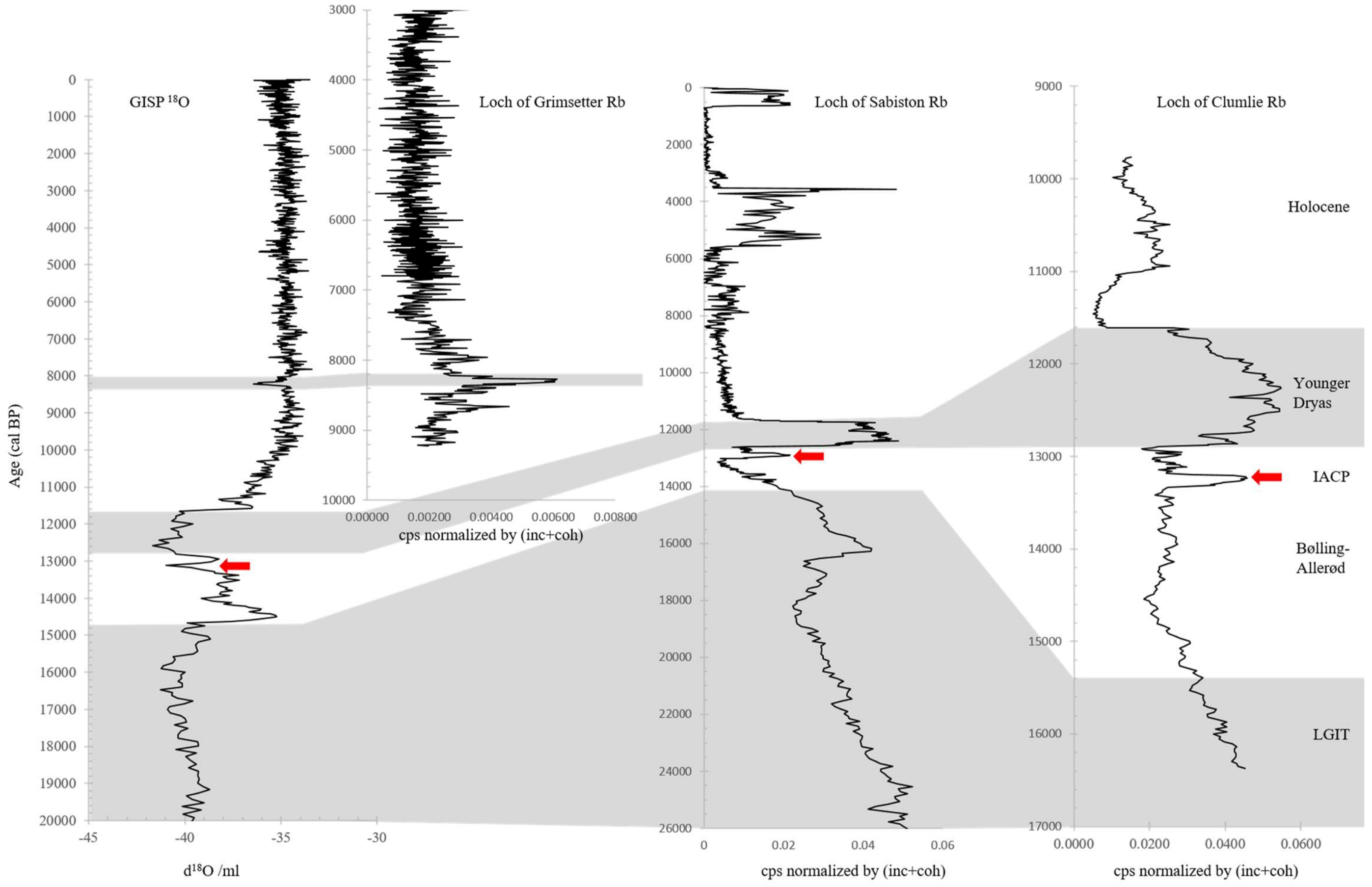

Figure 6.1 Summary of how the results of three lochs compare to the Greenland δ18O record. Rubidium has been chosen to represent the changes that have occurred at each of the three sites due to its presence in clay deposits and geochemically stable (Davies et al., 2015). The grey shaded areas designate the Younger Dryas (middle), and the LGIT (bottom) for Loch of Sabiston and Loch of Clumlie. The top grey area designates the 8.2 event which was evident in the record from Loch of Grimsetter. The red arrows are indicating the IACP for Greenland, Loch of Sabiston and Loch of Grimsetter.

The diatom and geochemical records from Loch of Clumlie demonstrate that the initial stage of the Younger Dryas (12,880 cal BP to 12,270 cal BP) was characterised by colder and wetter conditions. This was then followed by a later stage (c. 12, 270 cal BP and 12,020 cal BP) marked by a shift to a cooler and drier environment. At Loch of Sabiston there is an increase in bio-productivity and a subtle shift in the diatom record during the Younger Dryas which could be indicative of cooler and drier conditions at *c.* 11,900 cal BP. The results from the diatom records from Loch of Clumlie and Loch of Sabiston provide strong evidence for a two stage Younger Dryas in the North Atlantic.

Across the North Atlantic there is evidence of time transgression of environmental and climatic changes that occurred at the beginning of the Younger Dryas (Alley and Ágústsdóttir, 2005; Lane et al., 2013; Muschitiello and Wohlfarth, 2015) brought on by a gradual south to north regional cooling spanning from *c.* 12,550 cal BP to 13,150 cal BP from Sluggan Bog Ireland in the south (54°N) to Kråkenes Norway (62°N) in the north. Madtjärn, Sweden (58°N), has a mean age for the beginning of the Younger Dryas of *c.* 12,677 cal BP (Muschitiello and Wohlfarth, 2015) which is close to the age for the onset of the Younger Dryas at Loch of Sabiston (59°N). Loch of Sabiston provides an independently dated record further strengthening the concept of time transgressive cooling during the start of the Younger Dryas. The chronology from Loch of Clumlie (60°N) is based on the Greenland chronological framework from Rasmussen et al (2014). However, the Greenland ice core date for the start of the Younger Drays at *c.* 12,880 cal BP is comparable to that of Kråkenes (62°N) with a maximum presumed start of *c.* 12,800 cal BP. Therefore, it would be expected to be in that range.

The beginning of the Holocene was marked by a rapid warming at *c.* 11,700 cal BP over a span of ~ 40 years (Oldfield, 2005; Taylor et al., 1997) after the Younger Dryas. The warmer conditions during the early Holocene was punctuated by three cooler events identified in the Greenland ice cores; the preboreal oscillation at *c.* 11,400 cal. BP, the 9300 cal BP event, and the 8200 cal BP event (Rasmussen et al., 2014). These cooling events have not been clearly identified in the Northern Isles in previous studies. The geochemical record from Loch of Grimsetter suggest cooling contemporary with the 8200 cal BP event which is not reflected in either the pollen or the diatom record. However, the Loch of Grimsetter record suggests stable environmental conditions throughout the remainder of the early Holocene. Loch of Sabiston in comparison, demonstrates changes in the catchment during the early Holocene in its pollen, and diatom records reflecting succession of terrestrial vegetation, lake ontogeny and hydroseral succession along with climatic changes, and yet does not record the 8200 cal BP event. There is also evidence on both Shetland and Orkney that terrestrial and aquatic proxies begin to respond independently of each other reflecting possible different forcing on the catchment that would not necessarily react at the same time as well as localized human perturbations with the exception of Grimsetter, which does respond to the 8.2 ka event and shows little change through the mid Holocene.

The mid-Holocene pollen record from Loch of Grimsetter contrasts markedly with others from Shetland in the timing of deforestation. The catchment on Bressay retained its woodland cover until *c*. 3000 cal BP, ~ 2000 - 5000 years later in comparison to other sites on Shetland. The loss of woodland coincides with increased human activity on the island recorded after ~ 3000 years in Loch of Grimsetter. This indicates that the island of Bressay was left relatively undisturbed during much of the Holocene in contrast to other sites from Shetland. This suggests a re-evaluation in how people interacted with the landscape and questions the assumption of the scale of the mid-Holocene clearances of woodland on Shetland.

With the increase in the use of tephrochronology, more independent chronologies can allow for better comparisons of regional climate patterns without having to make assumptions about synchroneity with other dated records. The carbonate rich geology of the Orkney Mainland has thwarted previous efforts to constrain palaeoenvironmental data using radiocarbon dating. The age-depth model obtained from Loch of Sabiston, using a combination of radiocarbon and tephra ages, is the most robust chronology that has been developed from Orkney. An expanded tephrochronology has been developed for Quoyloo Meadow (Timms et al., 2016), but has not yet been applied to any palaeoenvironmental records to date. There is a potential offset in the chronology for Loch of Sabiston of 100 - 300 years, however, it is not clear if this represents a regional difference (or lag) in the response to paleoclimatic changes in Orkney when correlated to the distant Greenland ice core records or if it is an artefact of the dating. Dating Loch of Clumlie has been difficult due to root penetration and hiatuses caused by drying out of the basin. In order to develop a working chronology for Loch of Clumlie, dates from Greenland ice core records supplemented the radiocarbon dates from the core. However, this ties the palaeoenvironmental records of Loch of Clumlie to Greenland assuming synchroneity in the two regions. However, the results from this research suggest that Shetland has more in common with the Norwegian/ Greenland region than with Orkney and Northern Europe and that correlating Loch of Clumlie with the Greenland ice core records is a suitable alternative to an independent chronology. This study also highlights the potential of high-resolution sediment chemistry enabled by μ-XRF analysis to correlate regional climatic events, particularly during the Lateglacial and so may be used as chemo-chronostratigraphical information to aid the development of age-depth models.

This research has contributed to and expanded the knowledge of the nature and timing of palaeoenvironmental change in the Northern Isles with high-resolution diatom, pollen, and sediment chemistry profiles from three previously unstudied lochs on Shetland and Orkney to describe the changing aquatic environment and the surrounding landscape. The findings of this study in the context of the research questions posed in Chapter 1 are summarized as follows:

1) Comparison of the synchroneity of the changes occurring in the three proxies in each of the study lochs demonstrates that in some cases that they can differ by several hundred years and therefore stating the start or end of a climatic event on one proxy can be problematic. Also. the sensitivity of each proxy has been revealed in their ability to respond to the cold events that occurred during the Bølling – Allerød interstadial such as the IACP, GI-1d and GI-1c3 (determined by δ18O records). Only the IACP was recognized in the μ-XRF records, to a lesser extent in the diatom records, and not at all in the pollen records of Loch of Sabiston and Loch of Clumlie. The addition of diatoms and μ-XRF geochemical analysis along with pollen analysis provides a better understanding of the nature and timing each loch responded to climatic changes in their catchments.

2) Determining the sensitivity of the proxies and how synchronous they are with other North Atlantic records was problematic due to complications in radiocarbon dating for Loch of Clumlie whereby the age-depth model had to be supplemented with NGRIP ice core dates. Suitable chronologies were developed however, for Sabiston and Grimsetter. Therefore, conclusions regarding the timing during the Lateglacial will not be independent from Greenland when comparing Loch of Clumlie to other sites. The results demonstrate that the changes that Shetland experienced were less extreme compared to those experienced on Orkney due to its more northerly location and that despite being in the same phytogeographical climatic region, there is a difference in the way each archipelago responds to climatic changes. On each of the archipelagos regional variation was more pronounced on Shetland than Orkney with site characteristics having an increased influence on the observed response to a changing climate.

3) The results of Fisher Information provide information on when changes begin to occur in the three proxies examined, allowing for a better understanding of the nature and timing of these changes. FI compliments the results of the constrained cluster analysis, which delineates zones and determines when something significantly changes thus presenting a more comprehensive look at the nature and timing of climatic changes. The FI results also shows periods of rapid changes or "tipping points", however the results do not always coincide with the results of CONISS, therefore the breaks of the CONISS zonation cannot be assumed to be regime shifts in each proxy.

This research shows that the nature and timing of the palaeoenvironmental changes that occurred from the end of the LGM through the early Holocene are not straight forward and utilizing proxies such as diatom and μ-XRF scanning techniques more often in the future will provide greater understanding of the processes occurring during environmental change in this region. This research will also provide insight for modelling future climate scenarios by providing information on how various proxies have reacted in the past under different climatic conditions.

REFERENCES

Ahmad, N., Derrible, S., Eason, T., Cabezas, H., 2016. Using Fisher information to track stability in multivariate systems. Royal Society Open Science 3, 1–8.

Alley, R.B., 2000. The Younger Dryas cold interval as viewed from central Greenland. Quaternary Science Reviews 19, 213–226.

Alley, R.B., Ágústsdóttir, A.M., 2005. The 8k event: Cause and consequences of a major Holocene abrupt climate change. Quaternary Science Reviews 24, 1123–1149.

Alley, R.B., Marotzke, J., Nordhaus, W., Overpeck, J., Peteet, D., Pielke, R., Pierrehumbert, R., Rhines, P., Stocker, T., Talley, L., Wallace, J., 2003. Abrupt Climate Change. Science 299, 2005–2011.

Alley, R.B., Mayewski, P.A., Sowers, T., Stuiver, M., Taylor, K.C., Clark, P.U., 1997. Holocene climatic instability: a prominent, widespread event 8200 years ago. Geology 25, 483–486.

Alley, R.B., Meese, D.A., Shuman, C.A., Gow, A.J., Taylor, K.C., Grootes, P.M., White, J.W.C., Ram, M., Waddington, E.D., Mayewski, P.A., Zielinski, G.A., 1993. Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event. Nature 362, 527–529.

Andersen, C., Koç, N., Jennings, A., Andrews, J.T., 2004. Nonuniform response of the major surface currents in the Nordic Seas to insolation forcing: Implications for the Holocene climate variability. Paleoceanography 19, 1–16.

Andersen, T., Carstensen, J., Hernández-García, E., Duarte, C.M., 2009. Ecological thresholds and regime shifts: approaches to identification. Trends in Ecology and Evolution 24, 49–57.

Anderson, N.J., 2000. Miniview: Diatoms, temperature and climatic change. European Journal of Phycology 35, 307–314.

Andresen, C., Björck, S., Rundgren, M., Conley, D., Jessen, C., 2006. Rapid Holocene climate changes in the North Atlantic: evidence from lake sediments from the Faroe Islands. Boreas 35, 23–34.

Andresen, C.S., Björck, S., Jessen, C., Rundgren, M., 2007. Early Holocene terrestrial climatic variability along a North Atlantic Island transect: palaeoceanographic implications. Quaternary Science Reviews 26, 1989–1998.

Arnell, N.W., Halliday, S.J., Battarbee, R.W., Skeffington, R. a., Wade, a. J., 2015. The implications of climate change for the water environment in England. Progress in Physical Geography 39, 93–120.

Baillie, M., 1991. Suck in and smear : Two related chronological problems for the 90s. Journal of Theoretical Archaeology 2, 12–16.

Bakke, J., Lie, Ø., Heegaard, E., Dokken, T., Haug, G.H., Birks, H.H., Dulski, P., Nilsen, T., 2009. Rapid oceanic and atmospheric changes during the Younger Dryas cold period. Nature Geoscience 2, 202–205.

Baldini, L.M., McDermott, F., Baldini, J.U.L., Arias, P., Cueto, M., Fairchild, I.J., Hoffmann, D.L., Mattey, D.P., Müller, W., Nita, D.C., Ontañón, R., Garciá-Moncó, C., Richards, D.A., 2015. Regional temperature, atmospheric circulation, and sea-ice variability within the Younger Dryas Event constrained using a speleothem from northern Iberia. Earth and Planetary Science Letters 419, 101–110.

Ballantyne, C.K., 2010. Extent and deglacial chronology of the last British-Irish Ice Sheet: Implications of exposure dating using cosmogenic isotopes. Journal of Quaternary Science 25, 515–534.

Bates, C.R.R., Bates, M.R.R., Dawson, S., Huws, D., Whittaker, J.E.E., Wickham-Jones, C.R.R., 2016. The environmental context of the Neolithic monuments on the Brodgar Isthmus, Mainland, Orkney. Journal of Archaeological Science: Reports 7, 394–407.

Bates, M.R., Nayling, N., Bates, R., Dawson, S., Huws, D., Wickham-Jones, C., 2013. A Multi-disciplinary Approach to the Archaeological Investigation of a Bedrock-Dominated Shallow-Marine Landscape: an example from the Bay of Firth, Orkney, UK. International Journal of Nautical Archaeology 42, 24–43.

Battarbee, R., Kneen, M., 1982. The use of electronically counted microspheres in absolute diatom analysis. Limnology and Oceanography 27, 184–188.

Battarbee, R.W., Charles, D., Dixit, S., Renberg, I., 1999. Diatoms as indicators of surface water acidity, in: Stoermer, E., Smol, J. (Eds.), The Diatoms: Applications for the Environmental and Earth Sciences. Cambridge University Press, Cambridge, pp. 85–127.

Battarbee, R.W., Jones, V.J., Flower, R.J., Cameron, N.G., Bennion, H., Carvalho, L., Juggins, S., 2001. Diatoms, in: Smol, J.P., Birks, H.J.B., Last, W.M. (Eds.), Tracking Environmental Change Using Lake Sediments. KluwerAcademic Publishers, Dordrecht, The Netherlands, pp. 155–202.

Bedford, A., Jones, R.T., Lang, B., Brooks, S., Marshall, J.D., 2004. A Late-glacial chironomid record from Hawes Water, northwest England. Journal of Quaternary Science 19, 281–290.

Benn, D., Lowe, J.J., Walker, M.J.C., 1992. Glacier response to climatic change during the Loch Lomond Stadial and early Flandrian: geomorohological and palynological evidence from the Isle of Skye, Scotland. Journal of Quaternary Science 7, 125–144.

Bennett, K., 1996. Determination of the number of zones in a biostratigraphical sequence. New Phytologist 132, 155–170.

Bennett, K., Boreham, S., Sharp, M., Switsur, V., 1992. Holocene history of environment, vegetation and human settlement on Catta Ness, Lunnasting, Shetland. Journal of Ecology

80, 241–273.

Bennett, K., Willis, K., 2001. Pollen, in: Smol, J., Birks, H. (Eds.), Tracking Environmental Change Using Lake Sediments. Kluwer Academic Publishers, Dordrecht, The Netherlands, pp. 5–32.

Bennett, K.D., 1993. Murraster, in: Birnie, J.F., Gordon, J.E., Bennett, K.D., Hall, A.M. (Eds.), The Quaternary of Shetland: Field Guide. Quaternary Research Association, Cambridge, p. 140.

Bennett, K.D., Boreham, S., Hill, K., Packman, S., Sharp, M.J., Switsur, V.R., 1993. Holocene environmental history at Gunnister, north Mainland, Shetland, in: Birnie, J.., Gordon, J.E., Bennett, K.., Hall, A. (Eds.), The Quaternary of Shetland: Field Guide. Quaternary Research Association, Cambridge, pp. 83–98.

Bennett, K.D., Fossitt, J.A., Sharp, M.J., Switsur, V.R., 1990. Holocene vegetational and environmental history at Loch Lang, South Uist, Western Isles, Scotland. New Phytologist 114, 281–298.

Bennett, K.D., Whittington, G., Edwards, K.J., 1994. Recent plant nomenclatural changes and pollen morphology in the British Isles. Quaternary Newsletter 73, 6.

Bennion, H., Fluin, J., Simpson, G.L., 2004. Assessing eutrophication and reference conditions for Scottish freshwater lochs using subfossil diatoms. Journal of Applied Ecology 41, 124–138.

Bennion, H., Simpson, G.L., 2011. The use of diatom records to establish reference conditions for UK lakes subject to eutrophication. Journal of Paleolimnology 45, 469–488.

Bichet, V., Gauthier, E., Massa, C., Perren, B., Richard, H., Petit, C., Mathieu, O., 2013. The history and impacts of farming activities in south Greenland: an insight from lake deposits. Polar Record 49, 210–220.

Bickerdike, H.L., Evans, D.J.A., Ó Cofaigh, C., Stokes, C.R., 2016. The glacial geomorphology of the Loch Lomond Stadial in Britain: a map and geographic information system resource of published evidence. Journal of Maps 5647, 1–9.

Birks, C.J. a, Koç, N., 2002. A high-resolution diatom record of late-Quaternary sea-surface temperatures and oceanographic conditions from the eastern Norwegian Sea. Boreas 31, 323–344.

Birks, H., 1994. Late-glacial vegetational ecotones and climatic patterns in Western Norway. Vegetation History and Archaeobotany 3, 107–119.

Birks, H., Birks, H., 1980. Quaternary Palaeoecology. Edward Arnold, London.

Birks, H., Heiri, O., 2010. Strengths and weaknesses of quantitative climate reconstructions based on late-Quaternary biological proxies. Open Ecology Journal 3, 68–110.

Birks, H.H., Jones, V.J., Brooks, S.J., Birks, H.J.B., Telford, R.J., Juggins, S., Peglar, S.M., 2012. From cold to cool in northernmost Norway: Lateglacial and early Holocene multiproxy environmental and climate reconstructions from Jansvatnet, Hammerfest. Quaternary Science Reviews 33, 100–120.

Birks, H.H., Mathewes, R.W., 1978. Studies in the vegetational history of Scotland: v. late devensian and early flandrian pollen and macrofossil stratigraphy at Abernethy Forest, Inverness-Shire. New Phytologist.

Birks, H.J.B., Birks, H.H., 2008. Biological responses to rapid climate change at the Younger Dryas-Holocene transition at Krakenes, western Norway. The Holocene 18, 19–30.

Birks, H.J.B., Felde, V.A., Bjune, A.E., Grytnes, J.-A.A., Seppä, H., Giesecke, T., 2016. Does pollen-assemblage richness reflect floristic richness? A review of recent developments and future challenges. Review of Palaeobotany and Palynology 228, 1–25.

Birks, H.J.B., Line, J.M., Juggins, S., Stevenson, a. C., Braak, C.J.F.T., 1990. Diatoms and pH Reconstruction. Philosophical Transactions of the Royal Society B: Biological Sciences 327, 263–278.

Birnie, J., 1991. Burn of Aith, in: Quaternary of Scotland. pp. 1980–1982.

Birnie, J.F., 2000. Devensian Lateglacial palaeoecological changes in Shetland. Boreas 29, 205–218.

Birnie, J.F., 1993. Spiggie, in: Birnie, J., Gordon, J.E., Bennett, K., Hall, A.M. (Eds.), The Quaternary of Shetland: Field Guide. Cambridge, pp. 45–46.

Birnie, J.F., 1981. Environmental changes in Shetland since the end of the last glaciation. University of Aberdeen.

Björck, S., Rundgren, M., Ingólfsson, O., Funder, S., 1997. The Preboreal oscillation around the Nordic Seas: terrestrial and lacustrune responses. Journal of Quaternary Science 12, 455–465.

Blaauw, M., 2012. Out of tune: the dangers of aligning proxy archives. Quaternary Science Reviews 36, 38–49.

Blaauw, M., 2010. Methods and code for “classical” age modelling of radiocarbon sequences. Quaternary Geochronology 512–518.

Blaauw, M., Christen, J.A., Mauquoy, D., van der Plicht, J., Bennett, K.D., 2006. 21st century suck-in or smear: Testing the timing of events between archives. PAGES News 14, 15–16.

Blaauw, M., Christen, J.A., Mauquoy, D., Van Der Plicht, J., Bennett, K.D., 2007. Testing the timing of radiocarbon-dated events between proxy archives. The Holocene 2, 283–288.

Boettiger, C., Ross, N., Hastings, A., 2013. Early warning signals: the charted and uncharted territories. Theoretical Ecology 6, 255–264.

Bondevik, S., Mangerud, J., Dawson, S., Lohne, Ø., 2005. Evidence for three North Sea tsunamis at the Shetland Islands between 8000 and 1500 years ago. Quaternary Science Reviews 24, 1757–1775.

Bonsall, C., Macklin, M., Anderson, D., Payton, R., 2002. Climate change and the adoption of agriculture in north-west Europe. European Journal of Archaeology 5, 9–23.

Boomer, I., von Grafenstein, U., Moss, A., 2012. Lateglacial to early Holocene multiproxy record from Loch Assynt, NW Scotland. Proceedings of the Geologists' Association 123, 109–116.

Boyle, J.F., 2003. Inorganic geochemical methods in palaeolimnology, in: Last, W., Smol, J. (Eds.), Tracking Environmental Change Using Lake Sediments Volume 2: Physical and Geochemical Methods. KluwerAcademic Publishers, Dordrecht, The Netherlands, pp. 83–141.

Bradley, R.S., England, J.H., 2008. The Younger Dryas and the Sea of Ancient Ice. Quaternary Research 70, 1–10.

Bradshaw, E., Jones, V., Birks, H., Birks, H., 2000. Diatom responses to late-glacial and early-Holocene environmental changes at Kråkenes, western Norway. Journal of Paleolimnology 23, 21–34.

Bradshaw, E.G., Rasmussen, P., Odgaard, B.V., Nielsen, H., Anderson, N.J., 2005. Mid- to late-Holocene land-use change and lake development at Dallund Sø, Denmark: trends in lake primary production as reflected by algal and macrophyte remains. The Holocene 15, 1130–1142.

Bradwell, T., Stoker, M., 2015. Asymmetric ice-sheet retreat pattern around northern Scotland revealed by marine geophysical surveys. Earth and Environmental Science Transactions of the Royal Society of Edinburgh 1–26.

Bradwell, T., Stoker, M.S., Golledge, N.R., Wilson, C.K., Merritt, J.W., Long, D., Everest, J.D., Hestvik, O.B., Stevenson, A.G., Hubbard, A.L., Finlayson, A.G., Mathers, H.E., 2008. The northern sector of the last British Ice Sheet: Maximum extent and demise. Earth-Science Reviews 88, 207–226.

Bray, J., Curtis, J., 1957. An ordination of the upland forest communities of southern Wisconsin. Ecological Monographs 27, 325–349.

Broecker, W.S., Kennett, J.P., Flower, B.P., Teller, J.T., Trumbore, S., Bonani, G., Wolfli, W., 1989. Routing of meltwater from the Laurentide Ice Sheet during the Younger Dryas cold episode. Nature 341, 318–321.

Bromley, G.R., Putnam, A.E., Rademaker, K.M., Lowell, T. V, Schaefer, J.M., Hall, B., Winckler, G., Birkel, S.D., Borns, H.W., 2014. Younger Dryas deglaciation of Scotland driven by warming summers. Proceedings of the National Academy of Sciences USA 111, 6215–6219.

Bronk Ramsey, C., Albert, P.G., Blockley, S.P.E., Hardiman, M., Housley, R.A., Lane, C.S., Lee, S., Matthews, I.P., Smith, V.C., Lowe, J., 2014. Improved age estimates for key Late Quaternary European tephra horizons in the RESET lattice. Quaternary Science Reviews 118, 18–32.

Brooks, S., Birks, J.H.B., 2000. Chironomid-inferred late-glacial air temperatures at Whitrig Bog, southeast Scotland. Journal of Quaternary Science 15, 759–764.

Brooks, S.J., Davies, K.L., Mather, K.A., Matthews, I.P., Lowe, J.J., 2016. Chironomid-inferred summer temperatures for the Last Glacial-Interglacial Transition from a lake sediment sequence in Muir Park Reservoir, west-central Scotland. Journal of Quaternary Science 31, 214–224.

Brooks, S.J., Matthews, I.P., Birks, H.H., Birks, H.J.B., 2012. High resolution Lateglacial and early-Holocene summer air temperature records from Scotland inferred from chironomid assemblages. Quaternary Science Reviews 41, 67–82.

Brooks, Birks, H.J.B., 2000. Chironomid-inferred late-glacial and early-Holocene mean July air temperatures for Krakenes Lake, western Norway. Journal of Paleolimnology 23, 77–89.

Brown, E., 2015. Estimation of Biogenic Silica Concentrations Using Scanning XRF: Insights from Studies of Lake Malawi Sediments, in: Croudace, I., Rothwell, R.G. (Eds.), Development in Paleoenvironmental Research: Micro-XRF Studies of Sediment Cores. Springer, New York, pp. 267–277.

Bunting, M., 1994. Vegetation history of Orkney, Scotland; pollen records from two small basins in west Mainland. New Phytologist 128, 771–792.

Bunting, M.J., Tipping, R., Downes, J., 2001. "Anthropogenic" Pollen Assemblages from a Bronze Age Cemetery at Linga Fiold, West Mainland, Orkney. Journal of Archaeological Science 28, 487–500.

Bunting, M.J.M., 1996. The development of heathland in Orkney, Scotland: pollen records from Loch of Knitchen (Rousay) and Loch of Torness (Hoy). The Holocene 6, 193–212.

Cabezas, H., Campbell, D.E., Eason, T., Garmestani, A.S., Heberling, M.T., Hopton, M.E., Templeton, J., White, D., Zanowick, M., Sparks, R.T., 2010. San Luis Basin Sustainability Metrics Project: A Methodology for Evaluating Regional Sustainability.

Campbell, I.D., 1999. Quaternary pollen taphonomy: Examples of differential redeposition and differential preservation. Palaeogeography, Palaeoclimatology, Palaeoecology 149, 245–256.

Canmore, 2017. Bressay, Cruester [WWW Document]. https://canmore.org.uk/site/1100/bressay-cruester.

Carter, J., Bailey-Watts, A., 1981. A taxonomic study of diatoms from standing freshwaters in Shetland. Nova Hedwigia 33, 513–628.

Clark, C.D., Hughes, A.L.C., Greenwood, S.L., Jordan, C., Sejrup, H.P., 2012. Pattern and timing of retreat of the last British-Irish Ice Sheet. Quaternary Science Reviews 44, 112–146.

Clark, P.U., Pisias, N.G., Stocker, T.F., Weaver, A.J., 2002. The role of the thermohaline circulation in abrupt climate change. Nature 415, 863–869.

Clarke, K.R., Somerfield, P.J., Chapman, M.G., 2006. On resemblance measures for ecological studies, including taxonomic dissimilarities and a zero-adjusted Bray-Curtis coefficient for denuded assemblages. Journal of Experimental Marine Biology and Ecology 330, 55–80.

Cole, G., Weihe, P., 2015. Textbook of Limnology, Fifth Ed. ed. Waveland Press, Longs Grove, Illinois.

Coope, G.R., Lemdahl, G., Lowe, J.J., Walkling, A., 1998. Temperature gradients in northern Europe during the last glacial – Holocene transition (14 – 9 14 C kyr BP) interpreted from coleopteran assemblages. Journal of Quaternary Science 13, 419–433.

Coxon, P., 2016. Landscapes and environments of the last glacial- interglacial transition : a time of amazingly rapid change in Ireland. Irish Naturalists Journal 29, 45–61.

Curry, R., Mauritzen, C., 2005. Dilution of the northern North Atlantic Ocean in recent decades. Science 308, 1772–1774.

Cushing, E.J., 1967. Evidence for differential pollen preservation in late Quaternary sediments in Minnesota. Review of Palaeobotany and Palynology 4, 87–101.

Cwynar, L., Burden, E., Mcandrews, J., 1979. Inexpensive sieving method for concentrating pollen and spores from fine-grained sediments. Canadian Journal of Earth Sciences 16, 1115–1120.

Dakos, V., Carpenter, S.R., Brock, W. a, Ellison, A.M., Guttal, V., Ives, A.R., Kéfi, S., Livina, V., Seekell, D. a, van Nes, E.H., Scheffer, M., 2012. Methods for detecting early warnings of critical transitions in time series illustrated using simulated ecological data. PloS ONE 7, 1–20.

Dakos, V., Carpenter, S.R., van Nes, E.H., Scheffer, M., 2015. Resilience indicators: prospects and limitations for early warnings of regime shifts. Philosophical Transactions of the Royal Society B: Biological Sciences 370, 1–10.

Davidson, D., Jones, R., 1985. The Environment of Orkney, in: Renfrew, C. (Ed.), The Prehistory of Orkney. Edinburgh University Press, Edinburgh.

Davidson, D., Jones, R., Renfrew, C., 1976. Palaeoenvironmental reconstruction and evaluation: a case study from Orkney. Transactions of the Institute of British Geographers 1, 346–361.

Davies, S.J., Lamb, H.F., Roberts, S.J., 2015. Micro-XRF Core Scanning in Paleolimnology: Recent Developments, in: Croudace, I.W., Rothwell, R.G. (Eds.), Development in Paleoenvironmental Research: Micro-XRF Studies of Sediment Cores. Springer, pp. 189–226.

Davies, S.M., Abbott, P.M., Pearce, N.J.G., Wastegård, S., Blockley, S.P.E., 2012. Integrating the INTIMATE records using tephrochronology: rising to the challenge. Quaternary Science Reviews 36, 11–27.

de la Vega Leinert, A., 2007. Holocene coastal environmental changes on the periphery of an area of glacio-isostatic uplift: an example from Scapa Bay, Orkney, UK. Journal of

Quaternary Science 22, 755–772.

de la Vega Leinert, A., Smith, D.E., Jones, R.L., 2012. Holocene Coastal Barrier Development at Bay of Carness, Mainland Island, Orkney, Scotland. Scottish Geographical Journal 128, 119–147.

Dean, W.E., 1974. Determination of carbonate and organic matter in calcareous sediments and sedmentry rocks by loss on ignition: Comparision with other methods. Journal of Sedimentary Petrology 44, 242–248.

Douglas, M.S., Smol, J.P., Blake, W., 1994. Marked post-18th century environmental change in high-arctic ecosystems. Science 266, 416–419.

Douglas, M.V., Smol, J.P., 1999. Freshwater diatoms as indicators of environmental change in the High Arctic, in: Stoermer, E., Smol, J. (Eds.), The Diatoms: Applications for the Environmental and Earth Sciences. University Press, Cambridge, Cambridge, pp. 227–244.

Downes, J., 1994. Excavation of a Bronze Age burial at Mousland, Stromness, Orkney. Proceedings of the Society of Antiquaries of Scotland 124, 141–154.

Dugmore, A., 1989. Icelandic volcanic ash in Scotland. Scottish Geographical Magazine 105, 168–172.

Eason, T., Garmestani, A.S., Cabezas, H., 2014. Managing for resilience: Early detection of regime shifts in complex systems. Clean Technologies and Environmental Policy 16, 773–783.

Eason, T., Garmestani, A.S., Stow, C.A., Rojo, C., Alvarez-Cobelas, M., Cabezas, H., Allen, C., 2016. Managing for resilience: an information theory-based approach to assessing ecosystems. Journal of Applied Ecology 53, 656–665.

Ebbesen, H., Hald, M., 2004. Unstable Younger Dryas climate in the northeast North Atlantic. Geology 32, 673–676.

EDINA, 2016. OS Terrain 5 [WWW Document]. http://digimap.edina.ac.uk.

EDINA, 2007. Land Cover Map [WWW Document]. http://digimap.edina.ac.uk.

Edwards, K.J., Moss, A., Whittington, G., 1993. A late glacial pollen site from Grunna Water, Nesting, in: Birnie, J.., Gordon, J.E., Bennett, K.., Hall, A. (Eds.), The Quaternary of Shetland: Field Guide. Quaternary Research Association, Cambridge, pp. 99–102.

Edwards, K.J., Whittington, G., 1998. Landscape and environment in prehistoric west Mainland, Shetland. Landscape History 20, 5–17.

Edwards, K.J., Whittington, G., Robinson, M., Richter, D., 2005. Palaeoenvironments, the archaeological record and cereal pollen detection at Clickimin, Shetland, Scotland. Journal of Archaeological Science 32, 1741–1756.

Edwards, K.J., Whittington, G., Tipping, R., 2000. The incidence of microscopic charcoal in late glacial deposits. Palaeogeography, Palaeoclimatology, Palaeoecology 164, 247–262.

Eldevik, T., Risebrobakken, B., Bjune, A.E., Andersson, C., Birks, H.J.B., Dokken, T.M., Drange, H., Glessmer, M.S., Li, C., Nilsen, J.E.Ø., Otterå, O.H., Richter, K., Skagseth, Ø., 2014. A brief history of climate – the northern seas from the Last Glacial Maximum to global warming. Quaternary Science Reviews 106, 225–246.

Engstrom, D.R., Wright, H.E., 1984. Chemical stratigraphy of lake sediments as a record of environmental change, in: Haworth, E., Lund, J. (Eds.), Lake Sediments and Environmental History. Leicester University Press, Leicester, pp. 11–68.

Farrell, M., 2009. The environmental context of later prehistoric human activity in Orkney, Scotland. University of Hull.

Farrell, M., Bunting, M.J., Lee, D.H.J., Thomas, A., 2012. Neolithic settlement at the woodland's edge: palynological data and timber architecture in Orkney, Scotland. Journal of Archaeological Science.

Fath, B.D., Cabezas, H., Pawlowski, C.W., 2003. Regime changes in ecological systems: An information theory approach. Journal of Theoretical Biology 222, 517–530.

Flower, R.J., Jones, V.J., Round, F.E., 1996. The problematic Fragilaria (virescens v.) exigua grun./Fragilaria exiguiformis (grun.) Lange-bertalot: a new species or a new genus? Diatom Research 11, 41–57.

Frenzel, B., 1966. Climatic change in the Atlantic/sub-Boreal transition on the Northern Hemisphere: botanical evidence, in: World Climate from 8000-0 BC Proceedings of the International Symposium on World Climate. Royal Meteorological Society, London, pp. 99–123.

Fritz, S., 1993. Diatom assemblages and ionic characterization of lakes of the northern Great Plains, North America: a tool for reconstructing past salinity and climate fluctuations. Canadian Journal of Fisheries and Aquatic Sciences 50, 1844–1856.

Fritz, S., 1989. Lake development and limnological response to prehistoric and historic land-use in Diss, Norfolk, UK. The Journal of Ecology 77, 182–202.

Fritz, S.C., Anderson, N.J., 2013. The relative influences of climate and catchment processes on Holocene lake development in glaciated regions. Journal of Paleolimnology 49, 349–362.

Fyfe, R.M., Twiddle, C., Sugita, S., Gaillard, M.-J., Barratt, P., Caseldine, C.J., Dodson, J., Edwards, K.J., Farrell, M., Froyd, C., Grant, M.J., Huckerby, E., Innes, J.B., Shaw, H., Waller, M., 2013. The Holocene vegetation cover of Britain and Ireland: overcoming problems of scale and discerning patterns of openness. Quaternary Science Reviews 73, 132–148.

Gaiser, E., Hée, J.J.M. La, Tobias, F.A.C., Wachnicka, A.H., La Hée, J.M., Tobias, F.A.C., Wachnicka, A.H., 2010. Mastogloia smithii var lacustris Grun.: a structural engineer of calcareous mats in karstic subtropical wetlands. Proceedings of the Academy of Natural Sciences of Philadelphia 160, 99–112.

Geirsdóttir, Á., Miller, G.H., Larsen, D.J., Ólafsdóttir, S., 2013. Abrupt Holocene climate transitions in the northern North Atlantic region recorded by synchronized lacustrine records in Iceland. Quaternary Science Reviews 70, 48–62.

Gilfedder, B.S., Petri, M., Wessels, M., Biester, H., 2011. Bromine species fluxes from Lake Constance's catchment, and a preliminary lake mass balance. Geochimica et Cosmochimica Acta 75, 3385–3401.

Gillespies, 1998. A landscape assessment of The Shetland Isles. Scottish Natural Heritage Review No. 93.

Ginn, B.K., Cumming, B.F., Smol, J.P., 2007. Long-term lake acidification trends in high- and low-sulphate deposition regions from Nova Scotia, Canada. Hydrobiologia 586, 261–275.

Golledge, N., Finlayson, A., Bradwell, T., Everest, J.D., 2008. The last glaciation of Shetland, north Atlantic. Geografiska Annaler, Series A: Physical Geography 90A, 37–53.

Golledge, N.R., 2010. Glaciation of Scotland during the Younger Dryas stadial: A review. Journal of Quaternary Science 25, 550–566.

Gordon, J.E., Sutherland, D.G., 1993. Orkney Islands, in: Gordon, J.E., Sutherland, D.G. (Eds.), Quaternary of Scotland, Geological Conservation Review Series. Chapman and Hall, London, pp. 69–82.

Griffiths, H.I., Reed, J.M., Leng, M.J., Ryan, S., Petkovski, S., 2002. The recent palaeoecology and conservation status of Balkan Lake Dojran. Biological Conservation 104, 35–49.

Grimm, E.E.C., 1987. CONISS: a FORTRAN 77 program for stratigraphically constrained cluster analysis by the method of incremental sum of squares. Computers & Geosciences 13, 13–35.

Grootes, P.M., Stuiver, M., White, J.W.C., Johnsen, S.J., Jouzel, J., 1993. Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. Nature 366, 552–554.

Haberzettl, T., Corbella, H., Fey, M., Janssen, S., Lucke, A., Mayr, C., Ohlendorf, C., Schabitz, F., Schleser, G.H., Wille, M., Wulf, S., Zolitschka, B., 2007. Lateglacial and Holocene wet-dry cycles in southern Patagonia: chronology, sedimentology and geochemistry of a lacustrine record from Laguna Potrok Aike, Argentina. The Holocene 17, 297–310.

Hajdas, I., Ivy, O.D., 2006. Problems in the extension of the radiocarbon calibration curve (10-13 kyr BP). Radiocarbon 37, 75–79.

Hall, A., 2013. The last glaciation of Shetland: local ice cap or invasive ice sheet? Norwegian Journal of Geology/Norsk Geologisk Forening 93, 229–242.

Havinga, A.J., 1967. Palynology and pollen preservation. Review of Palaeobotany and Palynology 2, 81–98.

Haworth, E., 1976. Two late-glacial (Late Devensian) diatom assemblage profiles from northern Scotland. New Phytologist 77, 227–256.

Hayward, C., 2012. High spatial resolution electron probe microanalysis of tephras and melt inclusions without beam-induced chemical modification. The Holocene 22, 119–125.

Hedges, J., 1974. Excavation of two Orcadian burnt mounds at Liddle and Beaquoy. Proceedings of the Society of Antiquaries of Scotland 1974–1795, 39–100.

Hedges, M., 1976. The excavation of the Knowes of Quoyscottie, Orkney: a cemetery of the first millenium BC. Proceedings of the Society of Antiquaries of Scotland 130–156.

Heegaard, E., Lotter, a. F., Birks, H.J.B., 2006. Aquatic Biota and the Detection of Climate Change: Are there Consistent Aquatic Ecotones? Journal of Paleolimnology 35, 507–518.

Heiri, O., Lotter, A., Lemcke, G., 2001. Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. Journal of paleolimnology 101–110.

Hill, M., 2013. Diversity and Evenness : A Unifying Notation and Its Consequences. Ecology 54, 427–432.

Hughes, M., Hornby, D.D., Bennion, H., Kernan, M., Hilton, J., Phillips, G., Thomas, R., 2004. The development of a GIS-based inventory of standing waters in Great Britain together with a risk-based prioritisation protocol. Water, Air, and Soil Pollution: Focus 4, 73–84.

Hulme, P., Durno, S., 1980. A contribution to the phytogeography of Shetland. New Phytologist 84, 165–169.

Hulme, P., Shirriffs, J., 1994. The Late-glacial and Holocene vegetation of the Lang Lochs Mire area, Gulberwick, Shetland: a pollen and macrofossil investigation. New Phytologist 80, 793–806.

Isarin, R.F.B., Renssen, H., Vandenberghe, J., 1998. The impact of the North Atlantic Ocean on the Younger Dryas climate in northwestern and central Europe. Journal of Quaternary Science 13, 447–453.

James, F., Rathbun, S., 1981. Rarefaction, relative abundance, and diversity of avian communities. The Auk 98, 785–800.

Jansen, E., Overpeck, J., Briffa, K.R., Duplessy, J.-C., Joos, F., Masson-Delmotte, V., Olago, D., Otto-Bliesner, B., Peltier, W.R., Rahmstorf, S., Ramesh, R., Raynaud, D., Rind, D., Solomina, O., Villalba, R., Zhang, D., 2007. Paleoclimate. Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change 433–497.

Jennings, A.E., Walton, M.E., Ó Cofaigh, C., Kilfeather, A., Andrews, J.T., Ortiz, J.D., De Vernal, A., Dowdeswell, J. a., 2014. Paleoenvironments during Younger Dryas-Early Holocene retreat of the Greenland Ice Sheet from outer Disko Trough, central west Greenland. Journal of Quaternary Science 29, 27–40.

Johansen, J., 1975. Pollen Diagrams from the Shetland and Faroe Islands. New Phytologist 75, 369–387.

Jones, V.J., Stevenson, A.C., Battarbee, R.W., 1989. Acidification of lakes in Galloway, south west Scotland: a diatom and pollen study of the post-glacial history of the Round Loch of Glenhead. The Journal of Ecology 77, 1–23.

Jouve, G., Francus, P., Lamoureux, S., Provencher-Nolet, L., Hahn, A., Haberzettl, T., Fortin, D., Nuttin, L., 2013. Microsedimentological characterization using image analysis and µ-XRF as indicators of sedimentary processes and climate changes during Lateglacial at Laguna Potrok Aike, Santa Cruz, Argentina. Quaternary Science Reviews 71, 191–204.

Juggins, S., 2015. rioja: Analysis of Quaternary Science Data, R package version (0.9-5). (http://cran.r-project.org/package=rioja).

Kaplan, M.R., Wolfe, A.P., 2006. Spatial and temporal variability of Holocene temperature in the North Atlantic region. Quaternary Research 65, 223–231.

Karunanithi, A.T., Cabezas, H., Frieden, B.R., Pawlowski, C.W., 2008. Detection and assessment of ecosystem regime shifts from fisher information. Ecology and Society 13, 22 [online].

Keatinge, T., Dickson, J., 1979. Mid-Flandrian Changes in Vegetation on Mainland Orkney. New Phytologist 82, 585–612.

Keatley, B.E., Douglas, M.S. V, Smol, J.P., 2008. Prolonged Ice Cover Dampens Diatom Community Responses to Recent Climatic Change in High Arctic Lakes. Arctic, Antarctic, and Alpine Research 40, 364–372.

Kéfi, S., Guttal, V., Brock, W.A., Carpenter, S.R., Ellison, A.M., Livina, V.N., Seekell, D.A., Scheffer, M., van Nes, E.H., Dakos, V., 2014. Early warning signals of ecological transitions: methods for spatial patterns. PloS ONE 9, 1–13.

Kelly, T.J., Hardiman, M., Lovelady, M., Lowe, J.J., Matthews, I.P., Blockley, S.P.E., 2017. Scottish early Holocene vegetation dynamics based on pollen and tephra records from Inverlair and Loch Etteridge, Inverness-shire. Proceedings of the Geologists' Association 128, 125–135.

Kelts, K., Hsü, K., 1978. Freshwater Carbonate Sedimentation, in: Lerman, A. (Ed.), LAKES Chemisrty Geology Physics. Springer-Verlag, New York, p. 363.

Kingsbury, M. V., Laird, K.R., Cumming, B.F., 2012. Consistent patterns in diatom assemblages and diversity measures across water-depth gradients from eight Boreal lakes from north-western Ontario (Canada). Freshwater Biology 57, 1151–1165.

Koç, N., Jansen, E., Haflidason, H., 1993. Paleoceanographic reconstructions of surface ocean conditions in the Greenland, Iceland and Norwegian seas through the last 14 ka based on diatoms. Quaternary Science Reviews 12, 115–140.

Krammer, K., Lange-Bertalot, H., 1991a. Bacillariophyceae. 3: Teil: Centrales, Fragilariaceae, Eunotiaceae, in: Ettl, H., Gärtner, G., Gerloff, J., Heynig, H., Mollenhauer, D. (Eds.), Süßwasserflora von Mitteleuropa, Band 2/3. Gustav Fischer Verlag, Stuttgart/New York.

Krammer, K., Lange-Bertalot, H., 1991b. Bacillariophyceae. 4: Teil: Achnanthaceae, in: Ettl, H., Gärtner, G., Gerloff, J., Heynig, H., Mollenhauer, D. (Eds.), Süßwasserflora von Mitteleuropa, Band 2/4. Gustav Fischer Verlag, Stuttgart/New York.

Krammer, K., Lange-Bertalot, H., 1988. Bacillariophyceae. 2: Teil: Bacillariaceae, Epithmiaceae, Surirellaceae, in: Ettl, H., Gärtner, G., Gerloff, J., Heynig, H., Mollenhauer, D. (Eds.), Süßwasserflora von Mitteleuropa, Band 2/2. Gustav Fischer Verlag, Stuttgart/New York.

Krammer, K., Lange-Bertalot, H., 1986. Bacillariophyceae 1: Teil: Naviculaceae, in: Ettl, H., Gärtner, G., Gerloff, J., Heynig, H., Mollenhauer, D. (Eds.), Süßwasserflora von Mitteleuropa, Band 2/1. Gustav Fischer Verlag, Stuttgart/New York.

Kylander, M.E., Ampel, L., Wohlfarth, B., Veres, D., 2011. High-resolution X-ray fluorescence core scanning analysis of Les Echets (France) sedimentary sequence: new insights from chemical proxies. Journal of Quaternary Science 26, 109–117.

Laird, K., Fritz, S., Maasch, K., Cumming, B., 1996. Greater drought intensity and frequency before AD 1200 in the Northern Great Plains, USA. Nature 384, 552–554.

Lambeck, K., 1995. Late Devensian and Holocene shorelines of the British Isles and North Sea from models of glacio-hydro-isostatic rebound. Journal of the Geological Society 152, 437–448.

Lambeck, K., 1991. Glacial rebound and sea-level change in the British Isles. Terra Nova 3, 379–389.

Land Use Consultants, 1998. Orkney landscape character assessment. Scottish Natural Heritage Review No 100.

Lane, C.S., Blockley, S.P.E., Mangerud, J., Smith, V.C., Lohne, S., Tomlinson, E.L., Matthews, I.P., Lotter, A.F., 2012. Was the 12.1ka Icelandic Vedde Ash one of a kind? Quaternary Science Reviews 33, 87–99.

Lane, C.S., Brauer, A., Blockley, S.P.E., Dulski, P., 2013. Volcanic ash reveals time-transgressive abrupt climate change during the Younger Dryas. Geology 41, 1251–1254.

Last, W., Smol, J.P., 2001. Developments in Paleoenvironmental Research: Tracking Environmental Change Using Lake Sediments: Volume 1: Basin Analysis, Coring and Chronological Techniques. Springer, New York.

Leinert, A., Keen, D., 2000. Mid-Holocene environmental changes in the Bay of Skaill, Mainland Orkney, Scotland: an integrated geomorphological, sedimentological and stratigraphical study. Journal of Quaternary Science 15, 509–528.

Levesque, A., Mayle, F.E., Walker, I.R., Cwynar, L.C., 1993. The Amphi-Atlantic Oscillation: A proposed late-glacial climatic event. Quaternary Science Reviews 12, 629–643.

Lipa, B., Barrick, D., Alonso-Martirena, A., Fernandes, M., Ferrer, M.I., Nyden, B., 2014. Brahan project high frequency radar ocean measurements: Currents, winds, waves and their interactions. Remote Sensing 6, 12094–12117.

Lohne, Ø.S., Mangerud, J., Birks, H.H., 2014. IntCal13 calibrated ages of the Vedde and Saksunarvatn ashes and the Younger Dryas boundaries from Kråkenes, western Norway. Journal of Quaternary Science 29, 506–507.

Lotter, A.F., Bigler, C., 2000. Do diatoms in the Swiss Alps reflect the length of ice-cover? Aquatic Sciences 62, 125.

Lowe, D., 2015. Connecting and dating with tephras: principles, functioning, and application of tephrochronology in Quaternary research. 12th Quaternary Techniques Short Course - Techniques of Palaeoclimatic and Palaeoenvironmental Reconstruction 1–30.

Lowe, J., Walker, M., 2015. Measuring Quaternary time: A 50-year perspective. Journal of Quaternary Science 30, 104–113.

Mackereth, F.J.H., 1966. Some chemical observations on Post-glacial lake sediments. Philosophical Transactions of the Royal Society B: Biological Sciences 250, 165–213.

Maitland, P.S., Boon, P.J., McLusky, D.S., 1994. The fresh waters of Scotland: a national resource of international significance. John Wiley & Sons Ltd., Chichester.

Mann, D.D.G., 1999. The species concept in diatoms. Phycologia 38, 437–495.

Mansilla, C.A., 2015. Palaeoenvironmental changes in southern Patagonia during the Late-glacial and the Holocene : implications for forest establishment and climate reconstructions . Submitted for the degree of Doctor of Philosophy Biological and Environmental Sciences School. University of Stirling.

Marquer, L., Gaillard, M.-J., Sugita, S., Trondman, A.-K., Mazier, F., Nielsen, A.B., Fyfe, R.M., Odgaard, B.V., Alenius, T., Birks, H.J.B., Bjune, A.E., Christiansen, J., Dodson, J., Edwards, K.J., Giesecke, T., Herzschuh, U., Kangur, M., Lorenz, S., Poska, A., Schult, M., Seppä, H., 2014. Holocene changes in vegetation composition in northern Europe: why quantitative pollen-based vegetation reconstructions matter. Quaternary Science Reviews 90, 199–216.

Marshall, J.D., Jones, R.T., Crowley, S.F., Oldfield, F., Nash, S., Bedford, A., 2002. A high resolution Late-Glacial isotopic record from Hawes Water, Northwest England. Climatic oscillations: Calibration and comparison of palaeotemperature proxies. Palaeogeography, Palaeoclimatology, Palaeoecology 185, 25–40.

Marshall, J.D., Lang, B., Crowley, S.F., Weedon, G.P., van Calsteren, P., Fisher, E.H., Holme, R., Holmes, J. a., Jones, R.T., Bedford, A., Brooks, S.J., Bloemendal, J., Kiriakoulakis, K., Ball, J.D., 2007. Terrestrial impact of abrupt changes in the North Atlantic thermohaline circulation: Early Holocene, UK. Geology 35, 639–642.

Marty, J., Myrbo, A., 2014. Radiocarbon dating suitability of aquatic plant macrofossils. Journal of Paleolimnology 52, 435–443.

Massa, C., Perren, B.B., Gauthier, É., Bichet, V., Petit, C., Richard, H., 2012. A multiproxy evaluation of Holocene environmental change from Lake Igaliku, South Greenland. Journal of Paleolimnology 48, 241–258.

Mayer, A.L., Pawlowski, C.W., Cabezas, H., 2006. Fisher Information and dynamic regime changes in ecological systems. Ecological Modelling 195, 72–82.

Mayewski, P. a., Rohling, E.E., Curt Stager, J., Karlén, W., Maasch, K. a., David Meeker, L., Meyerson, E. a., Gasse, F., van Kreveld, S., Holmgren, K., Lee-Thorp, J., Rosqvist, G., Rack, F., Staubwasser, M., Schneider, R.R., Steig, E.J., 2004. Holocene climate variability. Quaternary Research 62, 243–255.

Merritt, J.W., Connell, E.R., Hall, A.M., 2016. Middle to Late Devensian glaciation of north-east Scotland : implications for the north-eastern quadrant of the last British – Irish ice sheet. Journal of Quaternary Science 32, 276–294.

Michelutti, N., McCleary, K.M., Antoniades, D., Sutherland, P., Blais, J.M., Douglas, M.S. V, Smol, J.P., 2013. Using paleolimnology to track the impacts of early Arctic peoples on freshwater ecosystems from southern Baffin Island, Nunavut. Quaternary Science Reviews 76, 82–95.

Millard, A., 2014. Conventions for Reporting Radiocarbon Determinations. Radiocarbon 56, 555–559.

Mithen, S., Wicks, K., Pirie, A., Riede, F., Lane, C., Banerjea, R., Cullen, V., Gittins, M., Pankhurst, N., 2015. A Lateglacial archaeological site in the far north-west of Europe at Rubha Port an t-Seilich, Isle of Islay, western Scotland: Ahrensburgian-style artefacts, absolute dating and geoarchaeology. Journal of Quaternary Science 30, 396–416.

Moar, N., 1969. Two pollen diagrams from the Mainland, Orkney Islands. New Phytologist 68, 201–208.

Moore, P., Webb, J., Collinson, M., 1991. Pollen Analysis, Second. ed. Blackwell Scientific Publications, Oxford, UK.

Moos, M.T., Laird, K.R., Cumming, B.F., 2009. Climate-related eutrophication of a small boreal lake in northwestern Ontario: a palaeolimnological perspective. The Holocene 19, 359–367.

Moritz, R.E., Bitz, C.M., Steig, E.J., 2002. Dynamics of recent climate change in the Arctic. Science 297, 1497–1502.

Morris, C.D., Emery, N., 1986. The setting for the Brough of Deerness, Orkney. Northern Studies 23, 1–30.

Mortensen, A.K., Bigler, M., Grönvold, K., Steffensen, J.P., Johnsen, S.J., 2005. Volcanic ash layers from the last glacial termination in the NGRIP ice core. Journal of Quaternary Science 20, 209–219.

Mueller, A.D., Islebe, G.A., Hillesheim, M.B., Grzesik, D.A., Anselmetti, F.S., Ariztegui, D., Brenner, M., Curtis, J.H., Hodell, D.A., Venz, K.A., 2009. Climate drying and associated forest decline in the lowlands of northern Guatemala during the late Holocene. Quaternary Research 71, 133–141.

Muschitiello, F., Wohlfarth, B., 2015. Time-transgressive environmental shifts across Northern Europe at the onset of the Younger Dryas. Quaternary Science Reviews 109, 49–56.

National Records of Scotland, 2012. 2011 Census : First Results on Population Estimates for Scotland - Release 1A.

Nesje, A., Dahl, S.O., 1993. Lateglacial and Holocene glacier fluctuations and climate variations in western Norway: A review. Quaternary Science Reviews 12, 255–261.

Nesje, A., Dahl, S.O., Bakke, J., 2004. Were abrupt Lateglacial and early-Holocene climatic changes in northwest Europe linked to freshwater outbursts to the North Atlantic and Arctic Oceans? The Holocene 14, 299–310.

Newton, A.A., Dugmore, A.A., Gittings, B., 2007. Tephrabase: tephrochronology and the development of a centralised European database. Journal of Quaternary Science 22, 737–743.

O'Connor, T., Bunting, M., 2009. Environmental change in an Orkney wetland: plant and molluscan evidence from Quoyloo Meadow, in: Allen, M., Sharples, N., O'Connor, T. (Eds.), Land and People: Papers in Memory of John G. Evans. Oxbox Books, Oxford, UK.

Oksanen, J., Blanchet, G., Kindt, R., Legendre, P., Minchin, P.R., O'Hara, R.., Simpson, G.L., Solymos, P., Stevens, M.H.H., Wagner, H., 2015. vegan: Community Ecology Package. R package version 2. 3-0.

Oldfield, F., 2005. Environmental changes Key issues and alternative perspectives. Cambridge University Press.

Ottersen, G., Planque, B., Belgrano, A., Post, E., Reid, P., Stenseth, N., 2001. Ecological effects of the North Atlantic Oscillation. Oecologia 128, 1–14.

Palmer, a. P., Rose, J., Rasmussen, S.O., 2012. Evidence for phase-locked changes in climate between Scotland and Greenland during GS-1 (Younger Dryas) using micromorphology of glaciolacustrine varves from Glen Roy. Quaternary Science Reviews 36, 114–123.

Parnell, A.C., Buck, C.E., Doan, T.K., 2011. A review of statistical chronology models for high-resolution, proxy-based Holocene palaeoenvironmental reconstruction. Quaternary Science Reviews 30, 2948–2960.

Patrick, R., 1996. Rivers of the United States, Volume III: The Eastern and Southeastern States. John Wiley & Sons Ltd., New York.

Pearce, C., Seidenkrantz, M.-S., Kuijpers, A., Massé, G., Reynisson, N.F., Kristiansen, S.M., 2013. Ocean lead at the termination of the Younger Dryas cold spell. Nature Communications 4, 1–6.

Pentecost, A., 2009. The Marl Lakes of the British Isles. Freshwater Reviews 2, 167–197.

Pentecost, A., Andrews, J.E., Dennis, P.F., Marca-Bell, A., Dennis, S., 2006. Charophyte growth in small temperate water bodies: Extreme isotopic disequilibrium and implications for the palaeoecology of shallow marl lakes. Palaeogeography, Palaeoclimatology, Palaeoecology 240, 389–404.

Perren, B.B., Massa, C., Bichet, V., Gauthier, E., Mathieu, O., Petit, C., Richard, H., 2012. A paleoecological perspective on 1450 years of human impacts from a lake in southern Greenland. The Holocene 22, 1025–1034.

Phillips, W., Hall, A., Ballantyne, C., Binnie, S., Kubik, P., Freeman, S., 2008. Extent of the last ice sheet in northern Scotland tested with cosmogenic 10Be exposure ages. Journal of Quaternary Science 23, 101–107.

Plagányi, É., Ellis, N., Blamey, L., Morello, E., Norman-Lopez, A., Robinson, W., Sporcic, M., Sweatman, H., 2014. Ecosystem modelling provides clues to understanding ecological tipping points. Marine Ecology Progress Series 512, 99–113.

Polyak, L., Alley, R.B., Andrews, J.T., Brigham-Grette, J., Cronin, T.M., Darby, D. a., Dyke, A.S., Fitzpatrick, J.J., Funder, S., Holland, M., Jennings, A.E., Miller, G.H., O'Regan, M., Savelle, J., Serreze, M., St. John, K., White, J.W.C., Wolff, E., 2010. History of sea ice in the Arctic. Quaternary Science Reviews 29, 1757–1778.

Punning, J.-M., Puusepp, L., 2007. Diatom assemblages in sediments of Lake Juusa, Southern Estonia with an assessment of their habitat. Hydrobiologia 586, 27–41.

Rach, O., Brauer, A., Wilkes, H., Sachse, D., 2014. Delayed hydrological response to Greenland cooling at the onset of the Younger Dryas in western Europe. Nature Geoscience 7, 109–112.

R Development Core Team, 2016. R: A Language and Environment for Statistical Computing.

Rahmstorf, S., 2007. Thermohaline Ocean Circulation. Encyclopedia of Quaternary Science 739–750.

Randsalu-Wendrup, L., Conley, D.J., Carstensen, J., Fritz, S.C., 2016. Paleolimnological records of regime shifts in lakes in response to climate change and anthropogenic activities. Journal of Paleolimnology 56, 1–14.

Rasmussen, S.O., Andersen, K.K., Svensson, a. M., Steffensen, J.P., Vinther, B.M., Clausen, H.B., Siggaard-Andersen, M.-L.L., Johnsen, S.J., Larsen, L.B., Dahl-Jensen, D., Bigler, M., Röthlisberger, R., Fischer, H., Goto-Azuma, K., Hansson, M.E., Ruth, U., 2006. A new Greenland ice core chronology for the last glacial termination. Journal of Geophysical Research 111, 1–16.

Rasmussen, S.O., Bigler, M., Blockley, S.P., Blunier, T., Buchardt, S.L., Clausen, H.B., Cvijanovic, I., Dahl-Jensen, D., Johnsen, S.J., Fischer, H., Gkinis, V., Guillevic, M., Hoek, W.Z., Lowe, J.J., Pedro, J.B., Popp, T., Seierstad, I.K., Steffensen, J.P., Svensson, A.M., Vallelonga, P., Vinther, B.M., Walker, M.J.C., Wheatley, J.J., Winstrup, M., 2014. A stratigraphic framework for abrupt climatic changes during the Last Glacial period based on three synchronized Greenland ice-core records: refining and extending the INTIMATE event stratigraphy. Quaternary Science Reviews 106, 14–28.

Rasmussen, S.O., Vinther, B.M., Clausen, H.B., Andersen, K.K., 2007. Early Holocene climate oscillations recorded in three Greenland ice cores. Quaternary Science Reviews 26, 1907–1914.

Rasmussen, T.L., Thomsen, E., 2008. Warm Atlantic surface water inflow to the Nordic seas 34-10 calibrated ka B.P. Paleoceanography 23, 1–13.

RCAHMS, 1946. The Royal Commission on the Ancient and Historical Monuments of Scotland. Twelfth report with an inventory of the ancient monuments of Orkney and Shetland. 3v. Edinburgh.

Reimer, P., 2013. IntCal13 and Marine13 Radiocarbon Age Calibration Curves 0–50,000 Years cal BP. Radiocarbon 55, 1869–1887.

Renssen, H., Mairesse, A., Goosse, H., Mathiot, P., Heiri, O., Roche, D.M., Nisancioglu, K.H., Valdes, P.J., 2015. Multiple causes of the Younger Dryas cold period. Nature Geoscience.

Robinson, M., 2004. A Late glacial and Holocene diatom record from Clettnadal, Shetland Islands, northern Scotland. Journal of Paleolimnology 31, 295–319.

Round, F.E., 1957. The late-glacial and post-glacial diatom succession in the Kentmere Valley deposit: I. Introduction, methods and flora. New Phytologist 56, 98–126.

Rühland, K., Priesnitz, A., Smol, J., 2003. Paleolimnological evidence from diatoms for recent environmental changes in 50 lakes across Canadian Arctic treeline. Arctic, Antarctic, and Alpine Research 35, 110–123.

Rühland, K., Smol, J., 2002. Freshwater diatoms from the Canadian arctic treeline and development of paleolimnological inference models. Journal of Phycology 264, 249–264.

Rühland, K.M., Karst, T., Paterson, A., Gregory-Eaves, R., Smol, J.P., Cumming, B.F., 1999. Standard Sediment Sample Preparation Methods for Siliceous Microfossils (Diatoms and Chrysophyte Scales and Cysts). Lab Manual 25.

Rühland, K.M., Smol, J.P., Pienitz, R., 2003. Ecology and spatial distributions of surface-sediment diatoms from 77 lakes in the subarctic Canadian treeline region. Canadian Journal of Botany 81, 57–73.

Scheffer, M., Bascompte, J., Brock, W.A., Brovkin, V., Carpenter, S.R., Dakos, V., Held, H., van Nes, E.H., Rietkerk, M., Sugihara, G., 2009. Early-warning signals for critical transitions. Nature 461, 53–9.

Scheffer, M., Carpenter, S.R., Lenton, T.M., Bascompte, J., Brock, W., Dakos, V., van de Koppel, J., van de Leemput, I.A., Levin, S.A., van Nes, E.H., Pascual, M., Vandermeer, J., 2012. Anticipating Critical Transitions. Science 338, 344–348.

Schleussner, C., Divine, D., Donges, J.F., Miettinen, A., Donner, R. V, 2015. Evidence for a non-linear regime shift in the North Atlantic ocean circulation at the onset of the Little Ice

Age. Climate Dynamics 16, 4268.

Scottish Natural Heritage, 2001. Natural Heritage Zones: A national assessment of Scotland's fresh waters.

Shotton, F., 1972. Example of hard-water error in radiocarbon dating of vegetable matter. Nature 240, 460–461.

Shuman, B., Bartlein, P.J., Webb, T., 2005. The magnitudes of millennial- and orbital-scale climatic change in eastern North America during the Late Quaternary. Quaternary Science Reviews 24, 2194–2206.

Sissons, J., 1967. The evolution of Scotland's scenery. Oliver and Boyd, Edinburgh.

Smith, D., 1993. Norwick and Burragarth, in: Birnie, J.F., Gordon, J.E., Bennett, K.D., Hall, A.M. (Eds.), The Quaternary of Shetland: Field Guide. Quaternary Research Association, Cambridge.

Smith, D.E., Hunt, N., Firth, C.R., Jordan, J.T., Fretwell, P.T., Harman, M., Murdy, J., Orford, J.D., Burnside, N.G., 2012. Patterns of Holocene relative sea level change in the North of Britain and Ireland. Quaternary Science Reviews 54, 58–76.

Smol, J., 1983. Paleophycology of a high arctic lake near Cape Herschel, Ellesmere Island. Canadian Journal of Botany 61, 2195–2204.

Smol, J., 1981. Problems associated with the use of "species diversity" in paleolimnological studies. Quaternary Research 15, 209–212.

Smol, J.P., 1988. Paleoclimate proxy data from freshwater Arctic diatoms. Verhandlungen Internationale Vereingung Limnologen 23, 837–844.

Smol, J.P., Birks, H.J.B., Last, W.M., 2001. Developments in Paleoenvironmental Research: Tracking Environmental Change Using Lake Sediments: Volume 3:Terrestrial, Algal, and Siliceous Indicators. KluwerAcademic Publishers, Dordrecht, The Netherlands.

Smol, J.P., Stoermer, E.F., 2010. The Diatoms Applications for the Environmental and Earth Sciences, Second Edi. ed. Cambridge University Press, Cambridge.

Solignac, S., de Vernal, A., Hillaire-Marcel, C., 2004. Holocene sea-surface conditions in the North Atlantic - contrasted trends and regimes in the western and eastern sectors (Labrador Sea vs. Iceland Basin). Quaternary Science Reviews 23, 319–334.

Spanbauer, T.L., Allen, C.R., Angeler, D.G., Eason, T., Fritz, S.C., Garmestani, A.S., Nash, K.L., Stone, J.R., 2014. Prolonged instability prior to a regime shift. PloS ONE 9, 1–7.

Stockmarr, J., 1971. Calcium tablets with spores used in absolute pollen analysis. Pollen et Spores 13, 615–621.

Stolze, S., Muscheler, R., Dörfler, W., Nelle, O., 2013. Solar influence on climate variability and human development during the Neolithic: evidence from a high-resolution multi-proxy record from Templevanny Lough, County Sligo, Ireland. Quaternary Science Reviews 67, 138–159.

Streeter, R., Dugmore, A.J., 2013. Anticipating land surface change. Proceedings of the National Academy of Sciences of the United States of America 110, 5779–84.

Stuiver, M., Reimer, P., RW, R., 2013. CALIB 7.0 (program and documentation).

Sugita, S., 1994. Pollen Representation of Vegetation in Quaternary Sediments : Theory and Method in Patchy Vegetation. Journal of Ecology 82, 881–897.

Svoboda, J., Henry, G.H.R., 1987. Succession in Marginal Arctic Environments. Arctic and Alpine Research 19, 373–384.

Taylor, K.C., Mayewski, P.A., Alley, R.B., Brook, E.J., Gow, A.J., Grootes, P.M., Meese, D.A., Saltzman, E.S., Severinghaus, J.P., Twickler, M.S., White, J.W.C., Whitlow, S., Zielinski, G.A., 1997. The Holocene Younger Dryas transition recorded at Summit, Greenland. Science 278, 825–827.

Teasdale, P.A., Collins, P.E.F., Firth, C.R., Cundy, A.B., 2011. Recent estuarine sedimentation rates from shallow inter-tidal environments in western Scotland: implications for future sea-level trends and coastal wetland development. Quaternary Science Reviews 30, 109–129.

Teller, J.T., Boyd, M., Yang, Z., Kor, P.S.G., Mokhtari Fard, A., 2005. Alternative routing of Lake Agassiz overflow during the Younger Dryas: new dates, paleotopography, and a re-evaluation. Quaternary Science Reviews 24, 1890–1905.

Thirumalai, K., Richey, J., 2016. Potential for paleosalinity reconstructions to provide information about AMOC variability. US Clivar Variations 14, 8–12.

Timms, R.G.O., Matthews, I.P., Palmer, A.P., Candy, I., Abel, L., 2016. A high-resolution tephrostratigraphy from Quoyloo Meadow, Orkney, Scotland: Implications for the tephrostratigraphy of NW Europe during the Last Glacial-Interglacial Transition. Quaternary Geochronology 1–15.

Tipping, 1991. Climatic change in Scotland during the Devensian Late Glacial: The palynological record, The Late Glacial in north-west Europe: human adaptation and environmental change at the end of the Pleistocene. Council for British Archaeology, Oxford.

Tipping, R., 2010. The case for climatic stress forcing choice in the adoption of agriculture in the British Isles, in: Finlayson, B., Warren, G. (Eds.), Landscapes in Transition, Levant Supplementary Series 8. Oxbow Books, Oxford, pp. 66–76.

Tipping, R., 1994. The form and fate of Scotland’s woodlands. Proceedings of the Society of Antiquaries of Scotland 124, 1–54.

Tipping, R., 1987. The origins of corroded pollen grains at five early postglacial pollen sites in western Scotland. Review of Palaeobotany and Palynology 53, 151–161.

Tipping, R., Bradley, R., Sanders, J., 2012. Moments of crisis: climate change in Scottish prehistory. Proceedings of the Society of Antiquaries of Scotland 142, 9–25.

Tisdall, E., McCulloch, R., 2013. Living with sand: A record of landscape change and storminess during the Bronze and Iron Ages Orkney, Scotland. Quaternary International 308–309, 205–215.

Trumbore, S.E., Sierra, C.A., Pries, C.E.H., 2016. Radiocarbon Nomenclature, Theory, Models, and Interpretation: Measuring Age, Determining Cycling Rates, and Tracing Source Pools, in: Schuur, E.A.G., Druffel, E.R.M., Trumbore, S.E. (Eds.), Radiocarbon and Climate Change. Springer Nature, Switzerland, p. 314.

Turner, J.N., Holmes, N., Davis, S.R., Leng, M.J., Langdon, C., Scaife, R.G., 2015. A multiproxy (micro-XRF, pollen, chironomid and stable isotope) lake sediment record for the Lateglacial to Holocene transition from Thomastown Bog, Ireland. Journal of Quaternary Science 30, 514–528.

Turney, C.S.M., Lowe, J.J., 2001. Tephrochronology, in: Last, W., Smol, J. (Eds.), Tracking Environmental Change Using Lake Sediments. KluwerAcademic Publishers, Dordrecht, The Netherlands, pp. 451–471.

Väliranta, M., Salonen, J.S., Heikkilä, M., Amon, L., Helmens, K., Klimaschewski, A., Kuhry, P., Kultti, S., Poska, A., Shala, S., Veski, S., Birks, H.H., 2015. Plant macrofossil evidence for an early onset of the Holocene summer thermal maximum in northernmost Europe. Nature Communications 10 Apr, 1–8.

Van Asch, N., Lutz, A.F., Duijkers, M.C.H., Heiri, O., Brooks, S.J., Hoek, W.Z., 2012. Rapid climate change during the Weichselian Lateglacial in Ireland: Chironomid-inferred summer temperatures from Fiddaun, Co. Galway. Palaeogeography, Palaeoclimatology, Palaeoecology 315–316, 1–11.

Van Dam, H., Mertens, A., Sinkeldam, J., 1994. A coded checklist and ecological indicator values of freshwater diatoms from The Netherlands. Netherlands Journal of Aquatic Ecology 28, 117–133.

Vincent, J.H., Cwynar, L.C., 2016. A temperature reversal within the rapid Younger Dryas-Holocene warming in the North Atlantic? Quaternary Science Reviews 153, 199–207.

Walker, M.J.C., Bohncke, S.J.P., Coope, G.R., O'Connell, M., Usinger, H., Verbruggen, C., 1994. The Devensian/Weichselian Late-glacial in northwest Europe (Ireland, Britain, north Belgium, The Netherlands, northwest Germany). Journal of Quaternary Science 9, 109–118.

Wanner, H., Solomina, O., Grosjean, M., Ritz, S.P., Jetel, M., 2011. Structure and origin of Holocene cold events. Quaternary Science Reviews 30, 3109–3123.

Weltje, G.J., Tjallingii, R., 2008. Calibration of XRF core scanners for quantitative geochemical logging of sediment cores: Theory and application. Earth and Planetary Science Letters 274, 423–438.

Wetzel, R.G., 2001. Limnology, Third Ed. ed. Academic Press, San Diego, California.

Whittington, G., Buckland, P., Edwards, K.J., Greenwood, M., Hall, A.M., Robinson, M., 2003. Multiproxy Devensian Late-glacial and Holocene environmental records at an Atlantic coastal site in Shetland. Journal of Quaternary Science 18, 151–168.

Whittington, G., Edwards, K.J., Zanchetta, G., Keen, D.H., Bunting, M.J., Fallick, A.E., Bryant, C.L., 2015. Lateglacial and early Holocene climates of the Atlantic margins of Europe: Stable isotope, mollusc and pollen records from Orkney, Scotland. Quaternary Science Reviews 122, 112–130.

Whittington, G., Fallick, A.E., Edwards, K.J., 1996. Stable oxygen isotope and pollen records from eastern Scotland and a consideration of Late-glacial and early Holocene climate change for Europe. Journal of Quaternary Science 11, 327–340.

Wickham-Jones, C.R., Dawson, S., Bates, R., 2009. The Submerged Landscape of Orkney. Archaeological Journal 166, 26–31.

Wiik, E., Bennion, H., Sayer, C.D., Davidson, T.A., Clarke, S.J., McGowan, S., Simpson, G.L., Stone, L., Prentice, S., Simpson, G.L., Stone, L., 2015a. The coming and going of a marl lake : multi-indicator palaeolimnology reveals abrupt ecological change and alternative views of reference conditions. Frontiers in Ecology and Evolution 3, 1–20.

Wiik, E., Bennion, H., Sayer, C.D., Davidson, T. a., McGowan, S., Patmore, I.R., Clarke, S.J., 2015b. Ecological sensitivity of marl lakes to nutrient enrichment: evidence from Hawes Water, UK. Freshwater Biology 60, 2226–2247.

Wiik, E., Bennion, H., Sayer, C.D., Willby, N.J., 2013. Chemical and biological responses of marl lakes to eutrophication. Freshwater Reviews 6, 35–62.

Wijk, S., 1986. Performance of Salix Herbacea in an Alpine Snow-Bed Gradient. Journal of Ecology 74, 675–684.

Williams, J.W., Blois, J.L., Gill, J.L., Gonzales, L.M., Grimm, E.C., Ordonez, A., Shuman, B., Veloz, S.D., 2013. Model systems for a no-analog future: species associations and climates during the last deglaciation. Annals of the New York Academy of Sciences 1297, 29–43.

Williams, J.W., Blois, J.L., Shuman, B.N., 2011. Extrinsic and intrinsic forcing of abrupt ecological change: case studies from the late Quaternary. Journal of Ecology 99, 664–677.

Willis, K.J., MacDonald, G.M., 2011. Long-Term Ecological Records and Their Relevance to Climate Change Predictions for a Warmer World. Annual Review of Ecology, Evolution, and Systematics 42, 267–287.

Yu, S.-Y., Shen, J., Colman, S.M., 2007. Modeling the Radiocarbon Reservoir Effect in Lacustrine Systems. Radiocarbon 49, 1241–1254.

APPENDIX

Appendix A - Diatom species names and authorities

Names and naming authority of species reaching a relative abundance of greater than three percent in the diatom assemblage of the three study lochs.

<u>Species Name</u>	<u>Naming Authority</u>
Achnanthes childanos	Hohn & Hellerman
Achnanthes lacusvulcani	Lange-Bertalot & Krammer
Achnanthes lanceolata	(Brébisson ex Kützing) Grunow
Achnanthes marginulata	Grunow
Achnanthes suchlandtii	Hustedt
Achnanthes ventralis	(Krasske) Lange-Bertalot
Achnanthidium helveticum	(Hustedt) Monnier, Lange-Bertalot & Ector
Achnanthidium minutissimum	(Kützing) Czarnecki
Achnanthidium subatomoides	(Hustedt) Monnier, Lange-Bertalot & Ector
Amphora libyca	Ehrenberg
Amphora ovalis	(Kützing) Kützing
Amphora pediculus	(Kützing) Grunow ex A.Schmidt
Amphora thumensis	(Mayer) Krieger
Aulacoseira granulata	(Ehrenberg) Simonsen
Brachysira aponina	Kützing
Brachysira neoexilis	Lange-Bertalot
Brachysira styriaca	(Grunow) R.Ross
Brachysira vitrea	(Grunow) R.Ross
Brachysira zellensis	(Grunow) Round & D.G.Mann
Caloneis latiuscula	(Kützing) Cleve
Cocconeis placentula var. lineata	(Ehrenberg) van Heurck
Cyclotella bodanica var. lemanica	(Otto Müller ex Schroter) Bachmann
Cyclotella comta	(Ehrenb.) Kütz.
Cyclotella distinguenda	Hustedt
Cyclotella meneghiniana	Kützing
Cyclotella stelligeroides	Hustedt
Cymbella cistula	(Ehrenberg) O.Kirchner
Cymbella cymbiformis	C.Agardh

Cymbella descripta	(Hustedt) Krammer & Lange-Bertalot
Cymbellafalsa diluviana	(Krasske) Lange-Bertalot & Metzeltin
Denticula kuetzingii	Grunow
Denticula tenuis	Kützing
Diatoma tenuis	C.Agardh
Encyonema silesiacum	(Bleisch) D.G.Mann
Encyonopsis microcephala	(Grunow) Krammer
Eolimna rotunda	(Hustedt) Lange-Bertalot, Kulikovskiy & Witkowski
Epithemia goeppertiana	Hilse
Epithemia smithii	Carruthers
Eucocconeis flexella	(Kützing) Meister
Eunotia arcus	Ehrenberg
Fragilaria capucina	Desmazières
Fragilaria constricta	Ehrenberg
Fragilaria nanana	Lange-Bertalot
Fragilariforma exigua	(Grunow) M.G.Kelly
Fragilariforma virescens	(Ralfs) D.M.Williams & Round
Gliwiczia calcar	(Cleve) M.Kulikovskiy, Lange-Bertalot & A.Witkowski
Gomphonema angustum	C.Agardh
Halamphora veneta	(Kützing) Levkov
Hippodonta capitata	(Ehrenberg) Lange-Bertalot, Metzeltin & Witkowski
Mastogloia smithii	Thwaites ex W.Smith
Navicula minima	Grunow
Navicula molestiformis	Hustedt
Navicula rhynchocephala	Kützing
Naviculadicta digitulus	(Hustedt)
Neidiomorpha binodis	(Ehrenberg) M.Cantonati, H.Lange-Bertalot & N.Angeli
Nitzschia bacillum	Hustedt
Nitzschia palea	(Kütz.) W. Sm.
Nitzschia perminuta	(Grunow) M. Perag.
Nitzschia linearis	(C. Agardh) W. Sm.
Nupela vitiosa	(Schimanski) P.Siver & P.B.Hamilton
Pinnularia biceps	W.Gregory
Pinnularia microstauron	(Ehrenberg) Cleve
Planothidium conspicuum	(A.Mayer) M.Aboal
Psammothidium levanderi	(Hustedt) L.Bukhtiyarova & Round

Pseudostaurosira brevistriata	(Grunow) Williams and Round 1987
Pseudostaurosira elliptica	(Schumann) Edlund, Morales & Spaulding
Rossithidium pusillum	(Grunow) Round & Bukhtiyarova
Sellaphora laevissima	(Kützing) D.G.Mann
Sellaphora vitabunda	(Hustedt) D.G.Mann
Stauroneis anceps	Ehrenberg
Staurosira construens	(Ehrenberg) Grunow
Staurosira venter	(Ehrenberg) Cleve & Moeller
Staurosirella leptostauron	(Ehrenberg) D.M.Williams & Round
Staurosirella pinnata	(Ehrenberg) D.M.Williams & Round
Tabellaria flocculosa	(Roth) Kützing

Appendix B – Standardization of published radiocarbon dates used in text to IntCal13 calibration curve (Reimer, 2013)**.**

Site	Sample ID	Depth (cm)	^{14}C Date	Error	Calibrated Age (median)	Reference
Shetland						
Dallican Water	Q-2755	698-710	9350	90	10562	Bennett et al. 1992
	Q-2756	652-664	7775	90	8565	
	Q-2757	578-590	5670	95	6466	
	Q-2758	552-564	5085	100	5821	
	Q-2759	478-486	3350	70	3590	
	Q-2760	433-439	1565	65	1461	
Gunnister Water	Not cited	Not cited	9785	80	11205	Bennett et al. 1993
	Not cited	Not cited	9405	70	10638	
	Not cited	Not cited	9300	70	10491	
	Not cited	Not cited	8605	100	9608	
	Not cited	Not cited	7110	70	7936	
	Not cited	Not cited	5660	125	6463	
	Not cited	Not cited	2685	60	2804	
	Not cited	Not cited	1870	50	1807	
Lang Lochs Mire	SRR-1552	725-755	13200	100	15859	Hulme and Shirriffs 1994
	SRR-1551	645-650	10450	70	12346	
	SRR-1550	515-520	9000	70	10156	
	SRR-1549	370-375	7720	70	8503	
	SRR-1548	295-300	7280	60	8095	
	SRR-1648	280-285	6445	80	7361	
	SRR-1547	195-200	5250	50	6023	
	SRR-1647	180-185	4575	95	5237	
	SRR-1646	65-70	1030	45	947	
Murraster	Not cited	Not cited	10400	160	12237	Bennett 1993
	Not cited	Not cited	10110	160	11721	
	Not cited	Not cited	7850	120	8695	
	Not cited	Not cited	4650	80	5391	
	Not cited	Not cited	520	70	548	
Crudale Meadow	* Too old - Hard water effect					Whittington et al. 2015

Clettnadal	AA-33292	210.5-211.5	3545	55	3833	Whittington et al. 2003
	AA-33291	218.5-219.5	3755	55	4119	
	CAMS-63296	225-226	7070	30	7896	
	CAMS-63297	235-236	8560	50	9532	
	AA-33290	245-246	9190	75	10366	
	AA-33289	254.5-255.5	9530	100	10871	
	AA-33288	265-266	10125	80	11741	
	AA-33287	288-289	11180	80	13041	
	AA-33286	296-297	12070	85	13923	
	AA-33285	305-306	11575	80	13404	
	AA-33284	309-310	12780	90	15233	
Orkney						
Loch of Knitchen	Q-2892	232.5-243.5	2780	50	2880	Bunting 1996
	Q-2891	377.5-386.5	5345	55	6126	
	Q-2890	407-414	5975	70	6816	
Loch of Torness	Q-2897	257.5-266.5	5500	50	6302	
	Q-2896	289.5-298.5	5930	50	6757	
	Q-2895	369.5-378.5	7035	60	7869	
	Q-2894	407-415	7350	65	8164	
	Q-2893	428-434	7660	60	8460	
Glims Moss	SRR-973	225-232	2145	65	2140	Keating and Dickenson 1979
	SRR-974	288-295	2090	60	2066	
	SRR-975	369-376	2886	65	3022	
	Birm-634a	395-400	2690	500	2808	
	Birm-634b	395-400	2960	110	3122	
	Birm-635	454-459	4200	160	4727	
	SRR-976	501-508	5681	55	6468	

Appendix C – Condition of Pollen from Loch of Sabiston and Loch of Clumlie

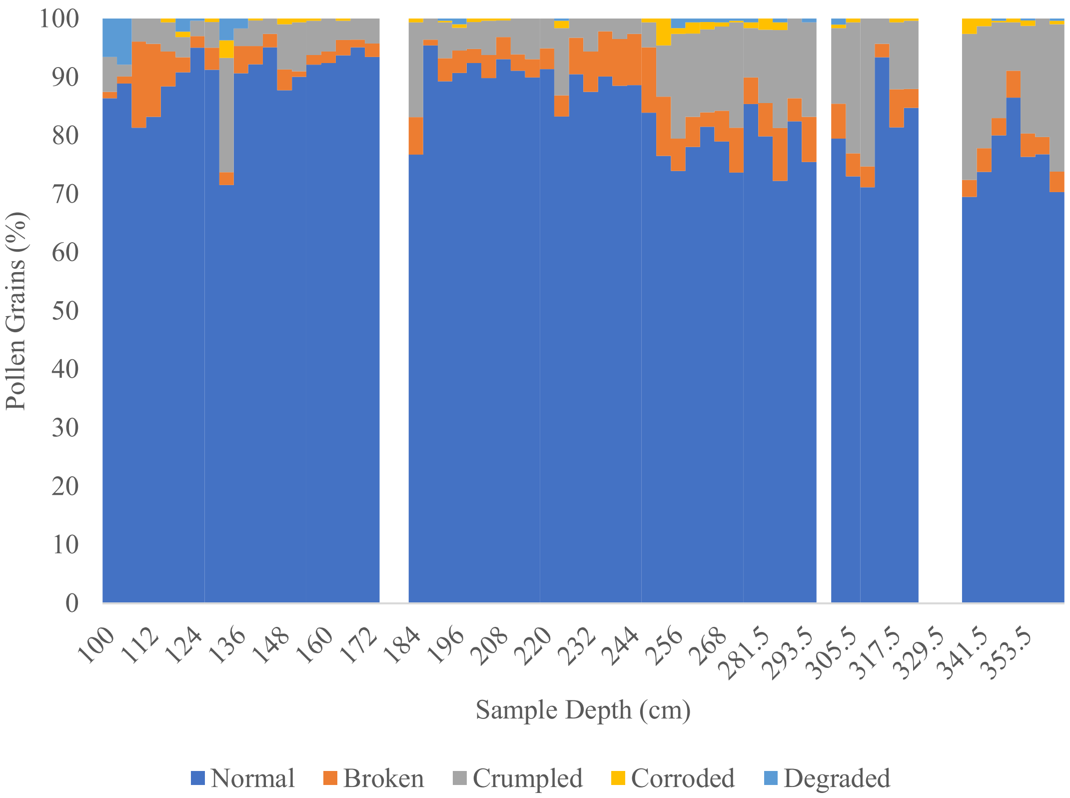

Figure C1. Condition of pollen from Loch of Sabiston. Spaces in the graph represent not enough pollen to count.

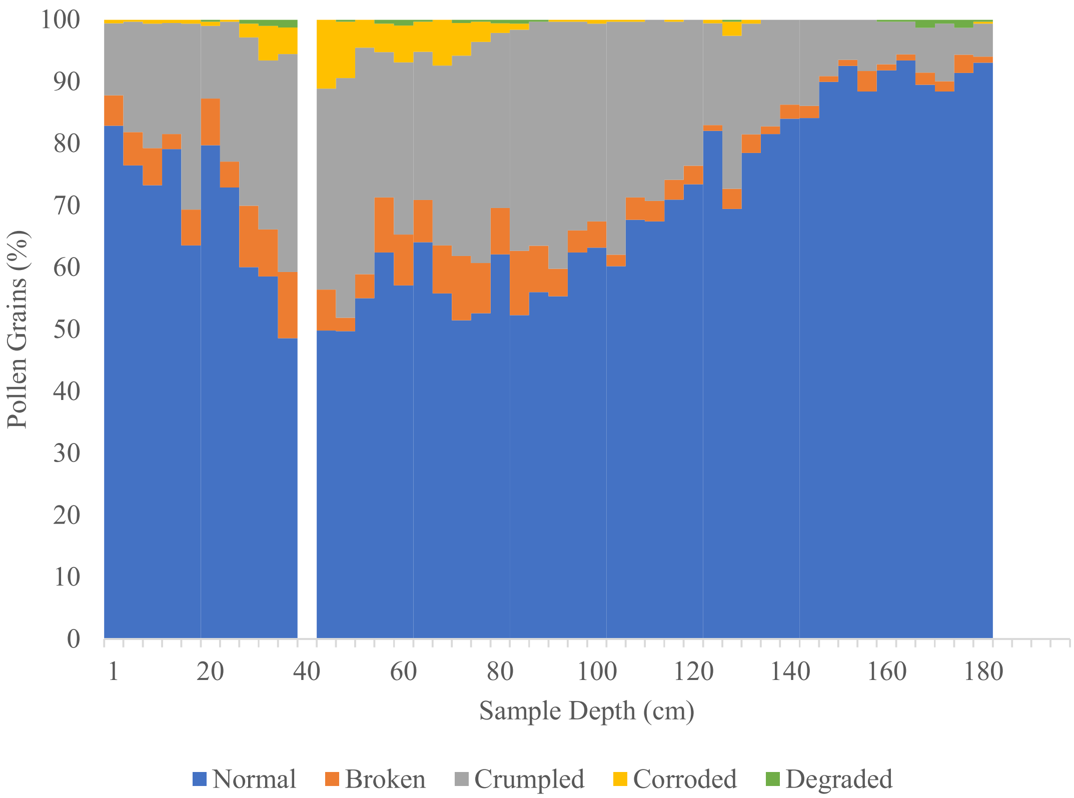

Figure C2. Condition of pollen from Loch of Loch of Clumlie. Spaces in the graph represent not enough pollen to count.