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GEOGRAPHY

Was there a '4.2 kyr event' in Great Britain and Ireland? Evidence from the peatland record

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I certify that all material in this thesis which is not my own work has been identified and that no material has previously been submitted and approved for the award of a degree by this or any other University.

Signature: ……………………………………………………………

Abstract

Palaeoenvironmental and archaeological data from several regions around the world show evidence of a multi-centennial climatic event occurring approximately 4200 cal yr BP. Abrupt climate change (ACC) events in the early-Holocene were dominated by meltwater pulse events associated with the final stages of deglaciation, a mechanism unlikely to have driven subsequent ACC events in the mid- and late-Holocene. A study of the '4.2 kyr event' therefore provides an opportunity to study an ACC event in the context of environmental conditions comparable to those of the modern day, thus providing valuable lessons for the future. Whilst the climatic change and/or impact of the 4.2 kyr event is clear in certain regions, such as western Asia, more work must be done to disentangle the timing and magnitude of change at this time in other regions, including northwest Europe. A more comprehensive reconstruction of the event's spatial and temporal variability will help determine the likely drivers of this event.

Presented here are the results of a multi-proxy examination of two peat sequences from Sluggan Moss and Fallahogy Bog in the North of Ireland. A range of palaeohydrological proxy analyses have been undertaken, including: peat humification, plant macrofossil and testate amoebae analyses. Furthermore, stable isotopic analysis (^{13}C and ^{18}O) of *Sphagnum* α-cellulose was included to determine whether this novel technique can, as a proxy for changing atmospheric circulation and/or bog surface wetness, contribute to our understanding of the nature and/or cause of the 4.2 kyr event. The chronological resolution of these sequences is exceptionally high, with radiocarbon dating supplemented by the excellent tephrochronology of the region. Together, these high-resolution palaeoecological and stable isotopic records provide the best opportunity to examine this event in northwest Europe, in terms of their potential for climatic sensitivity and chronological constraint.

After inter-site comparison, plant macrofossil and peat humification records were found to be climatically complacent. Testate amoebae records, however, were regionally coherent and were subsequently combined to produce a regional climatic record. From this, it was concluded that there was no compelling evidence to support the existence of a 4.2 kyr event in Great Britain and Ireland. In addition, data suggested that peat-based stable isotopic analysis cannot currently be accepted as a robust proxy for past palaeohydrological change. It is proposed that a lack of biomechanical understanding and standardised methodology is significantly hindering the potential of the technique's application in peat-based palaeoclimatic studies.

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List of Contents

List of Figures

List of tables

Chapter 1 – Introduction

1.1 Project rationale

The prospect of future climatic change is one of the key environmental, social and political issues facing modern society. In particular, abrupt climate changes with potentially significant ramifications for human societies are increasingly recognised as important foci for palaeoclimatic research (National Research Council, 2002; Alley *et al*., 2003).

Mid- to late-Holocene abrupt climatic changes are of particular scientific and public interest owing to their occurrence under a set of environmental parameters, which are closer to those of the present day. Conversely, abrupt changes in the early-Holocene are typically associated with the deglaciation of continental ice-sheets in the Northern Hemisphere at the end of the last glacial period. Consequently, the examination of mid-Holocene climatic events may provide better analogues for future episodes of abrupt climatic change, enhancing our understanding of past climate dynamics and climate response to past perturbations, which will help inform our ability to predict and potentially mitigate future change, through comparison with model simulations.

One mid-Holocene climatic event warrants particular scientific and public interest. The '4.2 kyr event' has been shown to be a period of abrupt and significant climatic change in most regions of the globe, and is reflected in the proxy-climate archives of North America (Booth *et al*., 2005), South America (Marchant and Hooghiemstra, 2004), Africa (Thompson *et al*., 2002), western Asia (Cullen *et al*., 2000), eastern Asia (Liu and Feng, 2012), Europe (Drysdale *et al*., 2006) and the North Atlantic (Bond *et al*., 2001).

Interest in the event was initially archaeological, as urban abandonment and societal collapse in Mesopotamia were linked directly to a severe multi-centennial drought (Weiss *et al*., 1993). The event has since been linked to widespread societal disruption in the other three centres of early civilisation, including Ancient Egypt, India and China (Weiss and Bradley, 2001; Stanley *et al*., 2003 Staubwasser and Weiss, 2006; Liu and Feng, 2012), and has consequently been dubbed the 'First Dark Age' (Thompson *et al*., 2002).

Walker *et al*. (2012) have recently identified the 4.2 kyr event, alongside the 8.2 kyr event, as suitable markers with which to subdivide the Holocene period. However, despite its broad spatial nature and archaeological importance, it is still poorly understood when compared with other Holocene events. The 4.2 kyr event has a complex set of potential driving factors but, as is the case with many climatic events, has been shown to have been heavily influenced by ocean-atmospheric circulation changes associated with the North Atlantic region.

The potential of northwest European peatlands, and those of Great Britain and Ireland in particular, as archives for recording climatic change in the North Atlantic, is increasingly well documented (Charman *et al*., 2006). As systems that are essentially governed by a combination of precipitation and temperature, proxy records developed from ombrotrophic raised bogs have the potential to possess a strong climatic signal (Charman *et al*., 2009). After recent rapid progress in the discipline, multiple quantitative estimates of past palaeohydrological change are achievable from peat records through a range of proxy techniques (e.g. Blackford and Chambers, 1993; Barber *et al.*, 1994; Charman *et al.*, 2007; Loisel *et al.* 2010), facilitating cross-proxy validation (e.g. Charman *et al.*, 1999; Chiverrell, 2001; Langdon *et al.*, 2003; Blundell and Barber, 2005; Hughes *et al.*, 2006). These archives also possess considerable chronological potential, with rapid accumulation, a wealth of datable organic material and the incorporation of known-age volcanic tephra horizons, providing an opportunity for the development of high-resolution palaeohydrological reconstructions with high-quality, robust chronologies. In addition, novel methodologies have been developed for the stable isotopic analysis of raised bog material, reflecting changes in the source of atmospheric moisture, which are strongly linked variation in atmospheric circulation patterns (Daley *et al.*, 2010).

As a result of these methodological benefits and developments, the peatlands of Great Britain and Ireland present an excellent opportunity to study the potential manifestation of the 4.2 kyr event in the region, potentially documenting the timing and nature of its palaeohydrological effects and its likely drivers. Consequently, this examination will form the basis of this study. Furthermore, this study aims to evaluate the potential for the use of stable isotope analysis alongside the traditional suite of palaeoecological techniques associated with multi-proxy peat-based palaeoclimate research.

1.2 Research questions

This study therefore has a number of research questions, aimed at addressing both the methodological and climatological themes outlined in the previous section. They are:

1. Based on existing evidence, is it possible to characterise the timing and nature of a global '4.2 kyr event'?
2. Can a coherent '4.2 kyr event' signal be characterised in Great Britain and Ireland?
3. Can peat-based stable isotopic analysis contribute to our understanding of the nature and/or cause of the '4.2 kyr event'?

To address these questions, this project developed two high-resolution, multi-proxy palaeoecological and stable isotopic records from sites in the North of Ireland. The region was primarily chosen for its exceptionally long peat records (Smith and Goddard, 1991), which could provide sufficient climatic context in which to examine any potential 4.2 kyr event signal. In addition, Ireland's position on the eastern seaboard of the North Atlantic makes it extremely sensitive to changes in North Atlantic oceanic-atmospheric circulation, which largely dominate the region's climate (McDermott *et al.*, 2001; Turney et al, 2005; Swindles *et al.*, 2010a). The chronological potential of the sites chosen was also deemed exceptional given not only the excellent opportunities for radiocarbon dating afforded by ombrotrophic raised bogs, but also as a result of a well-dated, ubiquitous Hekla 4 tephra horizon, occurring just prior to the hypothesised occurrence of the 4.2 kyr event, based on timing of the event in other regions of the world. In addition, the region's potential sensitivity to changes in the ocean-atmosphere circulation system presents an exceptional opportunity to test the potential benefit of stable isotopic analysis of peat material for palaeoclimatic reconstruction, alongside the other multi-proxy palaeoecological techniques.

1.3 Terminology

Throughout this study dates are expressed as calibrated calendar years before present (i.e. cal yr BP), where the present can be defined as AD 1950. Where previous researchers have expressed dates using a calendar age scale (i.e. cal BC/AD) this has been converted to the before present scale, again using AD 1950 as the base date. Individual radiocarbon dates were calibrated using the program CALIB v. 6.1.0 (Stuiver *et al.*, 2011) and the IntCal09 calibration curve (Reimer *et al.*, 2009).

Terms that are used with frequency in this study are given acronyms (e.g. bog surface wetness, BSW; water table depth, WTD). The full term is always given in the first instance, with the acronym to be used thereafter shown in brackets. In addition, where taxonomic names are repeated frequently, they will be shortened after their first instance (e.g. *Sphagnum austinii* to *S. austinii*).

1.4 Thesis structure

This chapter has outlined the rationale for the research and provided a context for the research questions that will be addressed. Chapter 2, the first of two review chapters, presents a discussion of the 4.2 kyr event in terms of its global climatic significance. This includes an extensive review of the palaeoclimatic evidence for its existence and an evaluation of the likely driving factors associated with its occurrence. Chapter 3 presents a synthesis of relevant published research concerning the role of peatlands in palaeoclimatic research, with a brief overview of the history, current methods and future directions of peat-based palaeohydrological reconstructions. In addition, a rationale for the study of mid-Holocene abrupt climatic changes in peatland archives is also presented. Chapter 4 provides a rationale behind study site selection, as well as a detailed description of each site including a review of any previous palaeoecological and palaeoenvironmental research and modern climate data. Chapter 5 provides detailed methodologies of the field, laboratory and statistical methods employed in this study. Chapter 6 provides a review of and rationale for the laboratory and statistical techniques associated with the construction of the chronologies, including radiocarbon dating, tephrochronology and age-depth model construction. The chapter also presents chronological results and final age-depth models. Chapter 7 presents all palaeoecological and stable isotopic results developed during this study. Chapter 8 presents a detailed discussion of these results, with a particular focus on their methodological evaluation and the incorporation of stable isotopic analyses in multi-proxy peat-based palaeoclimatic studies. Chapter 9 considers the evidence for a manifestation of the 4.2 kyr event in Great Britain and Ireland, incorporating data developed during the course of this study with other records from the region and the wider North Atlantic realm. This chapter also presents a methodological appraisal for the study of mid- to late-Holocene abrupt climatic change events in peat records. Finally, Chapter 10 summarises the conclusions that may be drawn from this study, together with suggestions for the future direction of this research.

Chapter 2 – The 4.2 kyr Event

2.1 Introduction

This chapter provides a synthesis of the current evidence for the existence of a significant climatic event c. 4200 cal yr BP, hereafter referred to as the '4.2 kyr event'. This review aims to collate a large number of palaeoclimate records in an effort to characterise the timing, nature and likely causes of climatic change around this time. In addition, a rationale for the study of this event in the peat archive of northwest Europe is presented.

2.2 Holocene climatic variability

Previously, the Holocene was thought to be period of relative climatic stability (Dansgaard *et al.,* 1993). However, it is now widely accepted that the current interglacial has been a period of considerable climatic variability (Mayewski *et al.*, 2004; Wanner *et al.*, 2008). Whilst climatic shifts may have been smaller in relative magnitude than those experienced at other points in recent geological history (e.g. the last glacial cycle), Holocene climatic variations have been undeniably larger, more frequent and more spatially complex than previously thought.

Evidence for pervasive millennial-scale Holocene climatic cycles has been identified throughout the North Atlantic region (Denton and Karlén, 1973; Bond *et al.*, 1993, 1997, 2001; O'Brien *et al.*, 1995; Bianchi and McCave, 1999; Langdon *et al.,* 2003; Turney *et al.*, 2005) and arguably throughout the mid-latitudes and tropics of the Northern Hemisphere (Hu *et al.*, 2003; Magny *et al.*, 2003). Global-scale teleconnections have also been demonstrated between this cyclicity in the North Atlantic and the monsoonal systems of Asia and Africa (Street-Perrott and Perrott, 1990; Overpeck *et al.*, 1996; Hong *et al.*, 2003; Wang *et al.*, 2005).

Bond *et al.* (2001) have suggested that such cycles are amplified responses in the North Atlantic region to relatively minor changes in solar activity as recorded by cosmogenic isotopes in annually-resolved tree-ring records (^{14}C) (Bond *et al.*, 1997) or Greenland ice-cores (^{10}Be) (Finkel and Nishiizumi, 1997). The role played by solar activity in driving such cycles has been called into question, however, not least because of the significant dating uncertainties associated with many reconstructions (Turney *et al.*, 2005; Wanner and Bütikofer, 2008).

2.3 Abrupt climate change events

In addition to this climatic periodicity, a number of relatively rapid and short-term climatic shifts have occurred during the Holocene. During the Lateglacial and early Holocene, events such as the Younger Dryas and 8.2 kyr event were characterised by strong North Atlantic signatures (Daley *et* al., 2011) and were hemispheric to global in extent (Alley *et al.*, 2003; Broecker, 2003). These abrupt climate change (ACC) events were caused by meltwater outbursts (Clark *et al.*, 2002; Teller *et al.*, 2002; Bauer, 2004; Teller and Leverington, 2004; Daley *et al.*, 2009) and ice-sheet surging (Broecker, 1994; Alley and MacAyeal, 1994; Broecker and Hemming, 2001; Hemming, 2004) associated with the deglaciation of continental ice-sheets in the Northern Hemisphere at the end of the last glacial period.

Previously, ACC has been defined as occurring "*when the climate system is forced to cross some threshold, triggering a transition into a new state at a rate determined by the climate system itself and faster than the cause*" (Alley and Ágústsdóttir, 2005, p.1123). Whilst this is certainly true of Lateglacial and early Holocene events, it is less applicable for events occurring from the mid-Holocene onwards, which have continued to occur in the absence of major ice sheets and their associated influence. With their causes less well defined, and their effects possibly more gradual in influence, the nature of mid- to late-Holocene climatic events requires that the exact definition of ACC be reconsidered.

Current definitions are heavily influenced by studies of early Holocene climatic changes (e.g. Alley *et al*., 1997; Alley and Ágústsdóttir, 2005; Gavin *et al*., 2011), but it may be more appropriate to talk of ACC events in a glacially related or unrelated context. However, mid- to late-Holocene events are still frequently described as 'abrupt' (e.g. Blaauw *et al*., 2007a; Ran *et al*., 2008; Kaniewski *et al*., 2008) and, as a result, this study shall refer to the mid- to late-Holocene event as ACC events.

Semantics aside, by examining past ACC events in terms of their patterns, processes and impacts, we can gain crucial insight into the mechanisms of climatic and environmental change on a number of spatial and temporal scales. For example, examination of the driving factors behind past ACC could improve our understanding of future 'tipping elements' in different parts of the Earth's climate system, and potentially the future 'tipping points' within these systems (see Lenton *et al*., 2008; Schellnhuber, 2009). The same can be said for understanding the rates of change involved in such events. Ideally, this will allow us to more effectively anticipate future events and identify the potential occurrence, nature and effect, including the potential for societal mitigation of ACC events. Whilst these significant events are part of the Earth's natural climatic variability, anthropogenic forcing may alter their probability (Alley *et al*., 2003). As a result, ACC events are increasingly recognised as important foci for palaeoclimatic research, and also as subjects of great societal interest, with potentially significant consequences (National Research Council, 2002; Alley *et al*., 2003). Furthermore, studies focusing on mid- to late-Holocene ACC events can provide better analogues for future episodes of ACC, as environmental parameters are closer to those of the present day. Generally, ACC events during this period are under-investigated and, subsequently, relatively poorly understood, in comparison with the larger magnitude early Holocene events.

2.4 The 4.2 kyr event

Interest in this mid-Holocene event was initially archaeological in nature, when excavations and their associated sedimentological and pedological analyses revealed that the Akkadian city of Tell Leilan, on the Habur Plains of Syria, was abandoned during a period of dramatically increased aridity and wind circulation that had induced considerable landscape degradation. It was concluded that a multi-centennial period of ACC, the 4.2 kyr event, was responsible for the abandonment of the city, wider regional desertion and, ultimately, led to the collapse of the

Akkadian empire in southern Mesopotamia at approximately 4175 ± 150 cal yr BP (Weiss *et al.*, 1993). In fact, the socioeconomic effects of this climatic event are reputedly documented more broadly in the archaeological record by the simultaneous collapse of a number of the world's first complex civilisations (Weiss *et al.*, 1993; Dalfes *et al.*, 1997; deMenocal, 2001; Weiss and Bradley, 2001; Stanley *et al.*, 2003; Staubwasser and Weiss, 2006; Liu and Feng, 2012). Numerous archaeologists had previously put forward evidence for widespread societal collapse during the third millennium B.C. in the Aegean, Egypt, the Middle East and Indus Valley (Crown, 1972; Palumbo, 1990; Rosen, 1989; Bell, 1971; Butzer, 1976; Hassan, 1981), but Weiss *et al.* (1993) were the first to link this regional abandonment directly to climate. More recently, coincident societal disruption has been widely documented in eastern Asia, particularly in China (Liu and Feng, 2012), with Walker *et al.* (2012) subsequently dubbing the period one of the most severe climatic events of the Holocene, in terms of its effects on human communities.

These archaeological observations are supported by palaeoclimatic evidence, which has revealed the occurrence of a significant, centennial-scale global climatic event at c. 4200 cal yr BP reflected in the proxy archives of North America (Booth *et al.*, 2005), South America (Marchant and Hooghiemstra, 2004), Africa (Thompson *et al.*, 2002), western Asia (Cullen *et al.*, 2000), eastern Asia (Liu and Feng, 2012), Europe (Drysdale *et al.*, 2006) and the North Atlantic (Bond *et al.*, 2001).

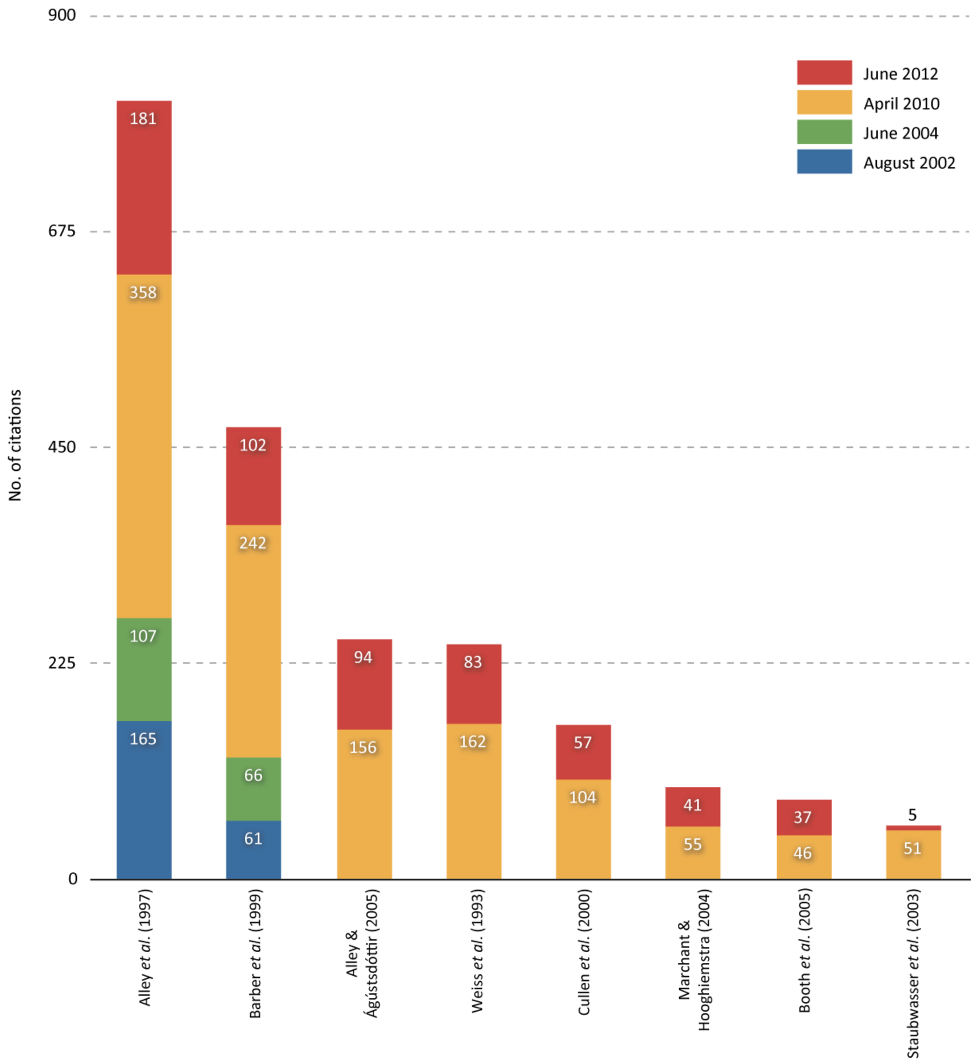
The 4.2 kyr event has recently been included, by the Subcommission on Quaternary Stratigraphy of the International Commission of Stratigraphy, in a formal discussion of the need to subdivide the Holocene Series/Epoch using new Global Stratotype Sections and Points (GSSP) (Walker *et al.*, 2012). Here, the 8.2 kyr event is presented as an Early-Middle Holocene Boundary, with the 4.2 kyr event as a Middle-Late Holocene Boundary, on account of its manifestation in a wide range of geomorphological, stratigraphical and archaeological records from many parts of the world, and thus it constitutes an excellent temporal marker for the period (Weiss, 2012; Walker *et al.*, 2012).

However, despite the broad spatial nature of the 4.2 kyr event, it is still poorly understood when compared with the 8.2 kyr event. A simple quantitative review of the literature, comparing the number of citations received by key papers associated with the 8.2 and 4.2 kyr events clearly shows the relative imbalance in discussion. As Alley and Ágústsdóttir (2005) observe, the literature surrounding the 8.2 kyr event is extensive and growing. They monitored an increasing number of citations, as recorded by ISI Web of Science, received by two key 8.2 kyr event papers, namely: Alley *et al.* (1997), which details the widespread climatic anomalies associated with the 8.2 kyr event; and, Barber *et al.* (1999), which discusses causation of the event. Table 2.1 and Figure 2.1 continue to monitor these two papers, presenting their citations as recorded in 2010 and 2012 from ISI Web of Science, alongside those of Alley and Ágústsdóttir's (2005) 8.2 kyr event review paper, and a number of key papers relating to the 4.2 kyr event.

Table 2.1 Comparison of cumulative citations for key papers relating to the 8.2 kyr (unitalicised) and 4.2 kyr events (italicised) over time. Asterisks as recorded by Alley and Ágústsdóttir's (2005). Source: ISI Web of Science.

Paper	**August 2002***	**June 2004***	**April 2010**	**June 2012**
Alley *et al.* (1997) Review of evidence for 8.2 kyr event	165	272	630	811
Barber *et al.* (1999) Discussion of likely causes for 8.2 kyr event	61	127	369	471
Alley and Ágústsdóttir (2005) Review of evidence for 8.2 kyr event	-	-	156	250
Weiss et al. *(1993)* *First identification for a 4.2 kyr event*	-	-	*162*	*245*
Cullen et al. *(2000)* *Evidence for 4.2 kyr event from Gulf of Oman*	-	-	*104*	*161*
Marchant and Hooghiemstra (2004) *Review of 4.2 kyr event evidence from South America and Africa*	-	-	*55*	*96*
Booth et al. *(2005)* *Review of 4.2 kyr event evidence from North America and Africa*	-	-	*46*	*83*
Staubwasser et al. *(2003)* *Evidence of 4.2 kyr event from Indus delta*	-	-	*51*	*56*

Figure 2.1 Comparison of cumulative citations for key papers relating to the 8.2 kyr and 4.2 kyr events over time. Asterisks as recorded by Alley and Ágústsdóttir's (2005). Source: ISI Web of Science.



This clearly shows that papers of a comparable age (e.g. Alley *et al.* (1997) and Weiss *et al.* (1993); Cullen *et al.* (2000) and Barber *et al.* (1999)) have received dramatically different numbers of citations. Papers concerned with the 8.2 kyr event receive consistently more citations than those of the 4.2 kyr event, or indeed any other Lateglacial or Holocene climatic event, apart from perhaps the Younger Dryas. Even review-based papers of similar ages (e.g. Alley and Ágústsdóttir (2005) compared to Marchant and Hooghiemstra (2004) and Booth *et al.* (2005)) receive disparate numbers of citations. Considering the apparent ramifications of the event for the world's first complex civilisations, this imbalance is surprising, and when coupled with the event's occurrence in an analogous, non-glacial period, it is clear that further examination of the event is required.

2.5 Summary of evidence for the 4.2 kyr event

Unlike early Holocene ACC, events in the mid- to late-Holocene are more commonly associated with hydroclimatic, rather than temperature, change, affecting precipitation regimes and drought frequencies (Booth *et al.*, 2005). In fact, many droughts of significant length (multi-annual to decadal) are recorded throughout this period, of which the considerable ecological and socioeconomic effects are well documented (Hodell *et al.*, 1995; Woodhouse and Overpeck, 1998; Stahle *et al.*, 2000; deMenocal, 2001: Booth *et al.* 2005). As the 4.2 kyr event was recorded so prominently in North America, Africa, and through the Middle East to China (Mayewski *et al.*, 2004; Walker *et al.*, 2012), it was originally discussed as a mid- to low latitude aridification event. However, further examination of the event reveals that its climatic manifestations are more complicated (see Section 2.6).

During the compilation of the summary presented here, a number of issues became apparent with regard to the timing, nature and nomenclature of the 4.2 kyr event. A recent, shorter review of the same event encountered a similar set of problems (Liu and Feng, 2012). As the following discussion, and summary presented in Table 2.2, demonstrates, manifestations of the 4.2 kyr event exhibit a degree of chronological inconsistency between records, both in terms of 'start' dates and duration. This could be the result of a genuinely time-transgressive event, system- or region-specific responses to climate variation, or merely of variation in the chronological quality of the records discussed. It also becomes apparent that the 'event' takes on a pattern more characteristic of a 'trend' in certain records, a tendency also observed by Liu and Feng (2012), further complicating chronological issues (see Section 9.6.5). As a result, the event has been discussed under a number of names in the existing literature.

Clearly, designating a single age estimate to any complex climatic event is undertaken with the caveat that chronological uncertainty will persist, spatially as well as temporally. Weiss *et al.* (1993) originally stated that the aridification event responsible for the abandonment of the Habur plains occurred c. 2200 BC. This age estimate was based on the stratigraphical correlation of sedimentological evidence with ^{14}C ages obtained from the Tell Leilan IIb archaeological layer, from which urban abandonment is also inferred. When recalibrated using IntCal09 these ages average 4285 ± 202 cal yr BP, demonstrating the somewhat arbitrary nature of '4.2 kyr event' title that has developed since. However, whilst acknowledging this arbitrariness, in the interests of simplicity, this study will refer to significant climatic changes during this period collectively, as the '4.2 kyr event'.

Palaeoclimatic evidence for the 4.2 kyr event is widespread but complex. The following sections will summarise the evidence from each region before outlining the key trends. It must be stressed that this review is by no means exhaustive, but rather represents the author's attempt to present the main global patterns associated with the 4.2 kyr event (see Section 2.6). As a result, the majority of the reviewed evidence is summarised in Figure 2.2 and Table 2.2, with more detailed descriptions limited to only the key records from each region.

2.5.1 North America

Historically, ice-core records have often been used as reliable markers for periods of Quaternary and Holocene climate change (e.g. Dansgaard *et al.*, 1982; O'Brien *et al.*, 1995; Alley, 2000; Masson, 2000). However, as Walker *et al.* (2012) observe, the 4.2 kyr event is not explicitly observed in the Greenland ice cores, with the 8.2 kyr event being by far the most pronounced post-Younger Dryas climatic excursion here. However, Johnsen *et al.* (2001) suggest that c. 4000 cal yr BP marks the point at which optimum Holocene warmth terminates and temperatures begin to decline as recorded in the $\delta^{18}O$ data of the Central and Southern Greenlandic ice cores.

Walker *et al.* (2012) instead propose that a clearly defined and marked excursion to lower values in a $\delta^{18}O$ data from Mount Logan, northern Canada (Fisher *et al.*, 2008; Fisher, 2011), provides an excellent regional marker for the 4.2 kyr event in North America. Coincident increases in deuterium and calcium values in the same record reflect enhanced moisture transport from the Pacific (Fisher *et al.*, 2008; Fisher, 2011). A shift to cooler and wetter conditions is seen across the Pacific Northwest region at this time, in chironomid records (Clegg *et al.*, 2010), peat inception data (Gorham *et al.*, 2007), accelerated peat accumulation rates (Yu *et al.*, 2003), widespread neoglaciation and glacial advance (Benedict *et al.*, 2008; Menounos *et al.*, 2008), marine varve thickness records (Nederbragt and Thurow, 2001) and tree-ring widths (Zhang and Hebda, 2005). Hebda and Mathewes (1984) also found this period to be one of major cultural shift along the North American Pacific coast, although information from the rest of North America with regard to cultural response to the 4200 cal yr BP remains sparse.

Mid-continent North America has experienced drought on varying temporal and spatial scales throughout the Holocene (deMenocal, 2001). Manifestations of the 4.2 kyr event in the region follow a distribution typical of North American drought, concentrated around the Great Lakes and Central Plains areas (Booth *et al.*, 2005). Evidence for a widespread, multi-centennial drought is indicated in sand dune reactivation (Stokes and Gaylord, 1993; Forman *et al.*, 1995; Loope *et al.*, 1995; Mason *et al.*, 1997; Stokes and Swinehart, 1997), increased aeolian sediments in lacustrine records (Dean, 1993, 1997), lacustrine diatom records (Bradbury and Dietrich-Rurup, 1993), reduced lake-levels (Baedke and Thompson, 2000), peatland water table records (Booth *et al.*, 2004), a speleothem $\delta^{18}O$ record (Denniston *et al.*, 1999) and reduced flood magnitudes (Knox and Kundzewicz, 1997).

Table 2.2 Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	Site/Location	Reference	Archive	Proxy	Interpretation	Associated date (cal yr BP)	Age control
1	Mt Logan, Yukon	Fisher *et al.* (2008); Fisher (2011)	Ice-core	$\delta^{18}O$	Wet event	4200	Seasonal counted years
2	Moose Lake, S Alaska	Clegg *et al.* (2010)	Lacustrine sediments	Chironomids	Cool event	4100	210Pb/Radiocarbon
3	S Vancouver Island	Zhang and Hebda (2005)	Tree-rings	Ring-width	Cool event	3875-3855	Dendrochronology
"	Saanich inlet, Vancouver Island	Nederbragt and Thurow (2001)	Marine sediments	Varve thickness	Wet event	4200	Annually varved sediments/Radiocarbon
4	W Canada	Menounos *et al.* (2008)	Detrital and in situ wood	Glacial advance	Cool/Wet event	4200-3800	Dendrochronology/Radiocarbon
5	W Canada	Yu *et al.* (2003)	Multiple peatlands	Peat accumulation	Wetting trend	4000	Radiocarbon
6	Boothia Peninsula, Nunavut	Zabenskie and Gajewski (2007)	Lacustrine sediments	Pollen	Cool event	4200	Radiocarbon
7	Ferris Dune Field, Wyoming	Stokes and Gaylord (1993)	Sedimentological	Dune reactivation	Dry/Windy event	4300-4000	Radiocarbon
8	Colorado Rocky Mountains	Benedict *et al.* (2008)	Detrital and in situ wood	Glacial advance	Cool/Wet event	4200	Dendrochronology/Radiocarbon
9	E Colorado	Forman *et al.* (1995)	Terrestrial sediments	Dune reactivation	Dry/Windy event	4300-4000	Radiocarbon
10	Sand Hills, Nebraska	Mason *et al.* (1997)	Terrestrial sediments	Dune reactivation	Dry/Windy event	4300	Radiocarbon
"	Sand Hills, Nebraska	Loope *et al.* (1995)	Terrestrial sediments	Dune reactivation	Dry/Windy event	4300	Radiocarbon
"	Sand Hills, Nebraska	Stokes and Swinehart, (1997)	Terrestrial sediments	Dune reactivation	Dry/Windy event	4300	Radiocarbon
11	Elk Lake, Minnesota	Dean (1993)	Lacustrine sediments	Clastic aeolian sediments	Dry/Windy event	4300-4100	Radiocarbon
"	Elk Lake, Minnesota	Dean (1997)	Lacustrine sediments	Clastic aeolian sediments	Dry/Windy event	4300-4100	Radiocarbon
"	Elk Lake, Minnesota	Bradbury and Dietrich-Rurup (1993)	Lacustrine sediments	Diatoms	Dry event	4300-4100	Radiocarbon
12	Upper Peninsula, Michigan	Booth *et al.* (2004)	Raised bog	Testate amoebae	Dry event	4200	Radiocarbon
13	Lake Michigan	Baedke and Thompson (2000)	Terrestrial sediments	Lake Level	Dry event	4500-4000	Radiocarbon
14	SE Minnesota and NE Iowa	Denniston *et al.* (1999)	Speleothem	$\delta^{18}O$	Moisture source shift	4100	U-series

Table 2.2 (continued) Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	Site/Location	Reference	Archive	Proxy	Interpretation	Associated date (cal yr BP)	Age control
15	Upper Mississippi River	Knox and Kundzewicz (1997)	Terrestrial sediments	Flood sediments	Dry event	4500-4000	Radiocarbon
16	WC Illinois	Forman *et al.* (2001)	Terrestrial sediments	Dune reactivation	Dry/Windy event	4200	Radiocarbon
17	WC Mexico	Figueroa-Rangel *et al.* (2008)	Terrestrial sediments	Pollen	Drying trend	4200	Radiocarbon
18	Cariaco basin	Haug *et al.* (2001)	Marine sediments	Heavy metals	Drying trend	4000	Radiocarbon
19	Colombia	Marchant *et al.* (2001)	Terrestrial sediments	Pollen	Wet event	4000	Radiocarbon
20	Colombian Amazonia	Behling *et al.* (1999)	Terrestrial sediments	Pollen	Wet event	3860	Radiocarbon
21	Lake El Caimito, W Colombia	Vélez *et al.* (2001)	Lacustrine sediments	Multi-proxy	Wet event	4200	Radiocarbon
"	Lake El Patía, W Colombia	Vélez *et al.* (2005)	Lacustrine sediments	Multi-proxy	Wet event	4200	Radiocarbon
22	SC Brazil	Behling (1993)	Highland peatland	Pollen	Wetting trend	4200	Radiocarbon
23	Ecuadorian Amazonia	Weng *et al.* (2002)	Lowland wetland	Multi-proxy	Wetting trend	4200	Radiocarbon
24	Lake Ayauchi, Ecuadorian Amazonia	Bush and Colinvaux, (1988)	Lacustrine sediments	Multi-proxy	Drying trend	4200	Radiocarbon
25	Lake Kumpak, Ecuadorian Amazonia	Liu and Colinvaux, (1985)	Lacustrine sediments	Pollen	Drying trend	4200	Radiocarbon
26	Huascarán mountains, Peru	Thompson *et al.* (2000)	Ice-core	Atmospheric dust concentration	Dry event	4500	Seasonal counted years
"	Huascarán mountains, Peru	Davis and Thompson (2006)	Ice-core	Atmospheric dust concentration	Dry event	4500	Seasonal counted years
27	Lake Titicaca, Peru	Cross *et al.* (2000)	Lacustrine sediments	Lake Level	Wet event	3800	Radiocarbon
"	Lake Titicaca, Peru	Mourguiart (2000)	Lacustrine sediments	Lake Level	Wet event	3900	Radiocarbon
"	Lake Titicaca, Peru	Baker *et al.* (2001)	Lacustrine sediments	Lake Level	Wet event	4500	Radiocarbon
27	Lake Titicaca, Peru	Tapia *et al.* 2003	Lacustrine sediments	Diatoms	Wet event	4000	Radiocarbon
28	Cordillera Real, Bolivia	Abbott *et al.* (2000)	Lacustrine sediments	$\delta^{18}O$	Wet event	4000	Radiocarbon

Table 2.2 (continued) Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	Site/Location	Reference	Archive	Proxy	Interpretation	Associated date (cal yr BP)	Age control
29	Bolivian Altiplano	Argollo and Mourguiart (2000)	Highland peatland	Multi-proxy	Wetting trend	3900	Radiocarbon
30	C Brazil	Bertaux *et al.* (2002)	Speleothem	Laminae thickness	Wetting trend	3800	U-series
31	Salitre, SC Brazil	Alexandre *et al.* (1999)	Terrestrial sediments	Pollen	Wetting trend	4000	Radiocarbon
32	Chilean Coast	Maldonado and Villagran (2006)	Lowland wetland	Pollen	Wetting trend	4200	Radiocarbon
33	La Plata basin, Uruguay	Iriarte *et al.* (2004)	Lowland wetland	Multi-proxy	Dry event	4500	Radiocarbon
34	Isla Navarino, Chile	Sapkota *et al.* (2007)	Raised bog deposit	Atmospheric dust concentration	Windy event	4200	Radiocarbon
35	Antarctic Peninsula	Hjort *et al.* (1998)	Terrestrial sediments	Glacial retreat	Warming trend	4200-3000	Radiocarbon
36	Antarctic Peninsula and Victoria Land	Ingolfsson *et al.* (1998)	Terrestrial sediments	Glacial retreat	Warming trend	4000-3000	Radiocarbon
37	Greenland	Johnsen *et al.* (2001)	Ice-core	$\delta^{18}O$	Cooling trend	4000	Seasonal counted years
38	E Greenland shelf	Jennings *et al.* (2002)	Marine sediments	Carbonate	Cooling trend	4700	Radiocarbon
39	Denmark Strait	Andresen and Bjorck (2005)	Marine sediments	Multi-proxy	Cool event	4200-4000	Radiocarbon
40	NW and C Icelandic shelf	Castaneda *et al.* (2004)	Marine sediments	$\delta^{18}O$ and carbonate	Cool event	4200-4000	Radiocarbon
41	N Icelandic Shelf	Ran *et al.* (2008)	Marine sediments	Diatoms	Cool event	4300-4100	Radiocarbon
42	Faroe Islands	Andresen *et al.* (2006)	Lacustrine sediments	Multi-proxy	Cooling trend	4200	Radiocarbon
43	N Atlantic	Bond *et al.* (2001)	Marine sediments	Ice-rafted debris	Cool event	4200	Radiocarbon
44	N Icelandic Basin, N Atlantic	Hall *et al.* (2004)	Marine sediments	$\delta^{18}O$	Cool event	4000	Radiocarbon
45	Abyssal NE Atlantic	Smart *et al.* (2008)	Marine sediments	Benthic foram accumulation rate	Cool event	4000	Radiocarbon
46	Sanday and Scara Brae	Sommerville *et al.* (2007)	Terrestrial sediments	Aeolian sand layers	Stormy event	4000	OSL
47	N Scotland	Anderson *et al.* (1998)	Peatlands	Humification	Cool/Wet event	3900-3500	Radiocarbon

Table 2.2 (continued) Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	Site/Location	Reference	Archive	Proxy	Interpretation	Associated date (cal yr BP)	Age control
47	N Scotland	Tipping and Tisdall (2004)	Peatlands	Multi-proxy	Cool/Wet event	4300-3800	Radiocarbon
48	Walton Moss	Hughes *et al.* (2000)	Raised bog	Plant macrofossils	Cool/Wet event	4410-3990	Radiocarbon
"	Kirkpatrick Fleming	Tipping (1995)	Raised bog	Multi-proxy	Cool/Wet event	4000	Radiocarbon
"	Butterburn Flow	Mauquoy *et al.* (2008)	Raised bog	Testate amoebae	Cool/Wet event	4100	Radiocarbon
"	Butterburn Flow	Mauquoy *et al.* (2008)	Raised bog	Plant macrofossils	Cool/Wet event	4000	Radiocarbon
"	Bolton Fell Moss	Barber *et al.* (2003)	Raised bog	Plant macrofossils	Cool/Wet event	4350-4000	Radiocarbon
49	Abbeyknockmoy	Barber *et al.* (2003)	Raised bog	Plant macrofossils	Cool/Wet event	4400-4000	Radiocarbon
50	C and N Norway	Nilssen and Vorren (1991)	Raised bog	Humification	Cool/Wet event	4000	Radiocarbon
51	C Sweden	Gunnarson (2008)	Detrital and in situ wood	Lake -level	Wet event	4150-4050	Dendrochronology/Radiocarbon
"	C Sweden	Gunnarson *et al.* (2003)	Raised bog deposit	Multi-proxy	Cool/Wet event	4150-4050	Dendrochronology/Radiocarbon
52	Finland	Korhola *et al.* (2002)	Lacustrine sediments	Chironomids	Cool trend	4200	Radiocarbon
53	Finland	Korhola (1995)	Raised bog deposits	Peat initiation	Cool/Wet event	4300	Radiocarbon
54	Voldafjorden, W Norway	Sejrup *et al.* (2001)	Marine sediments	Multi-proxy	Cool event	4200	Radiocarbon
"	Sygneskardvatnet, W Norway	Nesje *et al.* (2000)	Lacustrine sediments	Glacial advance	Cooling trend	4500	Radiocarbon
"	Jostedalsbreen, W Norway	Nesje *et al.* (2001)	Lacustrine sediments	Glacial advance	Cooling trend	4000	Radiocarbon
55	Värmland, SC Sweden	Borgmark and Wastegård (2008)	Raised bogs	Humification	Cool/Wet event	4200	Radiocarbon
"	SC Sweden	Borgmark (2005)	Raised bog	Humification	Cool/Wet event	4300-4200	Radiocarbon
56	Halland, SW Sweden	de Jong *et al.* (2006)	Raised bog	Pollen	Cool/Wet event	4300	Radiocarbon
"	Halland, SW Sweden	de Jong *et al.* (2006)	Raised bog	Aeolian sand	Stormy event	4200	Radiocarbon
56	Halland, SW Sweden	Bjork and Clemmensen (2004)	Raised bog	Aeolian material	Stormy event	4200-4100	Radiocarbon
57	Jutland, Denmark	Clemmensen *et al.* (2009)	Coastal dunefields	Aeolian material	Stormy event	4150	Radiocarbon
58	Draved Mose	Aaby (1976)	Raised bog deposit	Multi-proxy	Cool/Wet event	4300	Radiocarbon

Table 2.2 (continued) Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	Site/Location	Reference	Archive	Proxy	Interpretation	Associated date (cal yr BP)	Age control
59	Engbertsdijksveen	van Geel *et al.* (1996)	Raised bog deposit	Plant macrofossils	Cool/wet event	4350	Radiocarbon
60	Jura Mountains	Magny (2004)	Lacustrine sediments	Lake-level	Wet event	4150-3950	Dendrochronology/Radiocarbon
61	N Alps	Röthlisberger (1986)	Detrital and in situ wood	Glacial retreat	Dry event	4000	Dendrochronology/Radiocarbon
62	Alpi Apuane, CW Italy	Drysdale *et al.* (2006)	Calcite flowstone	Multi-proxy	Dry event	4200	U-series
63	C Italy	Magri and Parra (2002)	Lacustrine sediments	Pollen	Dry event	4200	Radiocarbon
"	Lagi di Vico, C Italy	Narcisi (2000)	Lacustrine sediments	Aeolian material	Dry event	4400	Radiocarbon/Tephra
"	Lake Accesa, C Italy	Magny *et al.* (2007)	Lacustrine sediments	Lake-level	Wet event	4200	Radiocarbon/Tephra
64	SE France - SE Spain	Jalut *et al.* (2000)	Lacustrine/Terrestrial sediments	Pollen	Dry trend	4300-3400	Radiocarbon
65	Lagoa Grande, NW Spain	Leira (2005)	Lacustrine sediments	Diatoms	Cool event	4200	Radiocarbon
66	Lake Siles, S Spain	Carrión (2002)	Lacustrine sediments	Multi-proxy	Drying trend	4100	Radiocarbon
67	Almeria, Spain	Pantaléon-Cano *et al.* (2003)	Lacustrine sediments	Multi-proxy	Drying trend	4500	Radiocarbon
68	W Africa	deMenocal *et al.* (2000)	Marine sediments	Foraminifera	Peak SST	4200	Radiocarbon
69	Lake Bosumtwi, Ghana	Talbot *et al.* (1984)	Lacustrine sediments	Multi-proxy	Dry event	4200-4000	Radiocarbon
70	Lake Barombi Mbo, Cameroon	Maley and Brenac (1998)	Lacustrine sediments	Pollen	Drying trend	4000	Radiocarbon
71	Manga grasslands, N Nigeria	Holmes *et al.* (1999)	Lacustrine sediments	Aeolian material	Drying trend	4100	Radiocarbon
72	W Africa	Gasse and van Campo (1994)	Lacustrine sediments	Multi-proxy	Dry event	4000	Radiocarbon
73	Sahara	Kropelin *et al.* (2008)	Lacustrine sediments	Multi-proxy	Dry event	4300	Radiocarbon
74	Batéké Plateaux, Congo	Schwartz *et al.* (1996)	Terrestrial sediments	Pollen	Drying trend	3940	Radiocarbon
75	Burundi Highlands	Aucour *et al.* (1999)	Highland peatland	$\delta^{13}C$ and TOC	Wet event	4200	Radiocarbon
76	Lake Edward, Uganda	Russell and Johnson (2005)	Lacustrine sediments	Geochemistry	Dry event	4100	Radiocarbon
77	Kilimanjaro	Thompson *et al.* (2002)	Ice-core	Multi-proxy	Cool/Dry event	4000	Glacier accumulation model

Table 2.2 (continued) Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	Site/Location	Reference	Archive	Proxy	Interpretation	Associated date (cal yr BP)	Age control
78	Lake Turkana, Kenya	Johnson (1996)	Lacustrine sediments	Lake Level	Drying	4200	Radiocarbon
79	Lake Tilo, Ethiopia	Lamb *et al.* (2000)	Lacustrine sediments	Lake-level	Drying	4000	Radiocarbon
80	E Africa	Butzer *et al.* (1972)	Lacustrine sediments	Lake-level	Dry event	4200	Radiocarbon
"	E Africa	Fraedrich *et al.* (1997)	Lacustrine sediments	Lake-level	Dry event	4200	Radiocarbon
"	E Africa	Said (1993)	Lacustrine sediments	Lake-level	Dry event	4200	Radiocarbon
81	Madagascan plateaux	Gasse and van Campo (1998)	Lacustrine sediments	Multi-proxy	Drying trend	4000-2300	Radiocarbon
82	C Red Sea	Edelman-Furstenberg *et al.* (2009)	Marine sediments	Foraminifera	Dry event	4200-3400	Radiocarbon
83	Shaban Deep, N Red Sea	Arz *et al.* (2006)	Marine sediments	$\delta^{18}O$	Dry event	4200	Radiocarbon
84	Nile Delta, Egypt	Stanley *et al.* (2003)	Marine sediments	Multi-proxy	Dry event	4200-4000	Radiocarbon
85	Dead Sea	Migowski *et al.* (2006)	Terrestrial sediments	Lake Level	Dry event	4200	Radiocarbon
"	Soreq Cave, Israel	Bar-Matthews *et al.* (2003)	Speleothem	$\delta^{18}O$ and $\delta^{13}C$	Dry event	4200-4050	U-series
85	Dead Sea	Frumkin *et al.* (2001)	Terrestrial sediments	Lake-level	Dry event	4000	Radiocarbon
"	Dead Sea	Enzel *et al.* (2003)	Terrestrial sediments	Lake-level	Dry event	4000	Radiocarbon
"	Dead sea	Frumkin (2009)	Tree-rings	$\delta^{13}C$ / $\delta^{15}N$	Dry event	4000	Dendrochronology/Radiocarbon
86	Syrian coast	Kaniewski *et al.* (2008)	Alluvial sediments	Pollen	Dry event	4200-3900	Radiocarbon
87	Tell Leilan, Syria	Weiss *et al.* (1993)	Terrestrial sediments	Aeolian material	Dry event	4200	Radiocarbon
88	Lake Van, Turkey	Lemcke and Sturm (1997)	Lacustrine sediments	Lake sediments	Dry event	4200	Radiocarbon
89	High Plateau, Yemen	Wilkinson (1997)	Terrestrial sediments	Aeolian material	Dry event	4350	Radiocarbon
90	Arabian sea	Gupta *et al.* (2003)	Marine sediments	Foraminifera	Dry event	4300-4100	Radiocarbon
91	Gulf of Oman	Cullen *et al.* (2000)	Marine sediments	Aeolian material	Dry event	4025	Radiocarbon
92	SE Arabia	Parker *et al.* (2006)	Lacustrine sediments	Geochemistry	Dry event	4200	Radiocarbon

Table 2.2 (continued) Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	**Site/Location**	**Reference**	**Archive**	**Proxy**	**Interpretation**	**Associated date (cal yr BP)**	**Age control**
92	NE Rub'al-Khali desert, Arabia	Parker *et al.* (2004)	Lacustrine sediments	Multi-proxy	Dry trend	4100	Radiocarbon
93	Indus Delta	Staubwasser *et al.* (2003)	Marine sediments	$\delta^{18}O$	Dry event	4200	Radiocarbon
94	NW India	Prasad and Enzel (2006)	Lacustrine sediments	Multi-proxy	Dry event	4200	Radiocarbon
95	Tibetan plateau	Mügler *et al.* (2010)	Lacustrine sediments	Lake-level	Drying trend	4000	Radiocarbon
96	Tibetan plateau	Gasse and van Campo (1994)	Lacustrine sediments	Multi-proxy	Dry event	4000	Radiocarbon
"	Tibetan plateau	Wei and Gasse (1999)	Lacustrine sediments	$\delta^{18}O$	Dry event	4500-3500	Radiocarbon
97	Tibetan plateau	Mischke and Zhang (2010)	Lacustrine sediments	Multi-proxy	Cool event	4200-2800	Radiocarbon
98	W Siberia	Pitkanen *et al.* (2002)	Peatland	Pollen	Cool/Dry event	4300-4100	Radiocarbon
99	Taimyr Peninsula, N Russia	Laing and Smol (2003)	Lacustrine sediments	Diatoms	Cool event	4200	Radiocarbon
100	Ugii Nuur Basin, C Mongolia	Schwanghart *et al.* (2008)	Lacustrine sediments	Lake-level	Drying trend	4200-2800	Radiocarbon
101	W Loess Plateau, China	An *et al.* (2005)	Terrestrial sediments	Multi-proxy	Dry event	4090-3600	Radiocarbon
102	Daihai lake, Inner Mongolia	Peng *et al.* (2005)	Lacustrine sediments	Grain size	Dry event	4400-4200	Radiocarbon
103	Qishuihe River, China	Huang *et al.* (2011)	Fluvial sediments	Flood sediments	Dry/flooding event	4300-4000	Radiocarbon
"	Jinghe River, China	Huang *et al.* (2010)	Fluvial sediments	Flood sediments	Dry/flooding event	4200-4000	Radiocarbon
104	Mawmluh cave, India	Berkelhammer *et al.* (2011, 2012)	Speleothem	$\delta^{18}O$	Monsoonal shift	4059-3860	Absolute counted age/U-series
105	Hongyuan Bog, Tibetan Plateau	Hong *et al.* (2003)	Peatland	$\delta^{13}C$	Weak monsoon event	4200-4000	Radiocarbon
106	Dongge cave, SW China	Wang *et al.* (2005)	Speleothem	$\delta^{18}O$	Weak monsoon event	4200-4000	U-series
107	Yishu River Basin, Shandong, China	Gao *et al.* (2007)	Terrestrial sediments	Multi-proxy	Cool event	4200	Radiocarbon
108	Tung-Yuan Pond, S Taiwan	Yang *et al.* (2011)	Lacustrine sediments	Multi-proxy	Monsoonal shift	4200	Radiocarbon

Table 2.2 (continued) Summary of 150 palaeoclimatic records that exhibit evidence for significant climatic change c. 4200 cal yr BP. Numbers correspond to Figure 2.2.

Map No.	**Site/Location**	**Reference**	**Archive**	**Proxy**	**Interpretation**	**Associated date (cal yr BP)**	**Age control**
109	Jinchuan, NE China	Jiang *et al.* (2008)	Peatland	Pollen	Cool/Dry event	4200	Radiocarbon
110	Mutsu Bay, Japan	Kawahata *et al.* (2009)	Marine sediments	Multi-proxy	Cool event	4100-2300	Radiocarbon
"	Lake Tougou-ike, Japan	Kato *et al.* (2003)	Lacustrine sediments	Multi-proxy	Cool event	4000	Varve counted years/Radiocarbon
111	SW/SE Australia	Baker *et al.* (2005)	Marine sediments	SST	Warm event	4200	Radiocarbon
112	Australian Lowlands	Shulmeister and Lees (1995)	Terrestrial sediments	Pollen	Dry event	4000	Radiocarbon
113	Southern Ocean	Moros *et al.* (2009)	Marine sediments	Foraminifera	Cool event	4300	Radiocarbon
114	S Australia	Quigley *et al.* (2010)	Speleothem	$\delta^{18}O$ and $\delta^{13}C$	ENSO onset	5000	U-series
115	North Island, New Zealand	Gomez *et al.* (2004)	Marine/Terrestrial sediments	Multi-proxy	Stormy event, ENSO onset	4000	Radiocarbon
116	EPICA Dome C, Antarctica	Masson-Delmotte *et al.* (2004)	Ice-core	δD	Cool event	4100	Ice-flow model, cross-dating with Vostok
117	Ile de la Possession, Iles Crozet	van der Putten et al (2008)	Peatland	Multi-proxy	Drying trend	4100	Radiocarbon

NB. Single point age estimates, as opposed to age ranges, offered in the associated date column, represent the inferred or specified start date of climatic change c. 4200 cal yr BP. Age ranges are only presented when explicitly specified by the authors themselves. It is important to note that a range of 'event'-type signals have been identified in the proxy record review, but that patterns indicative of changing climatic trends, associated with the time period, have also been identified and presented.

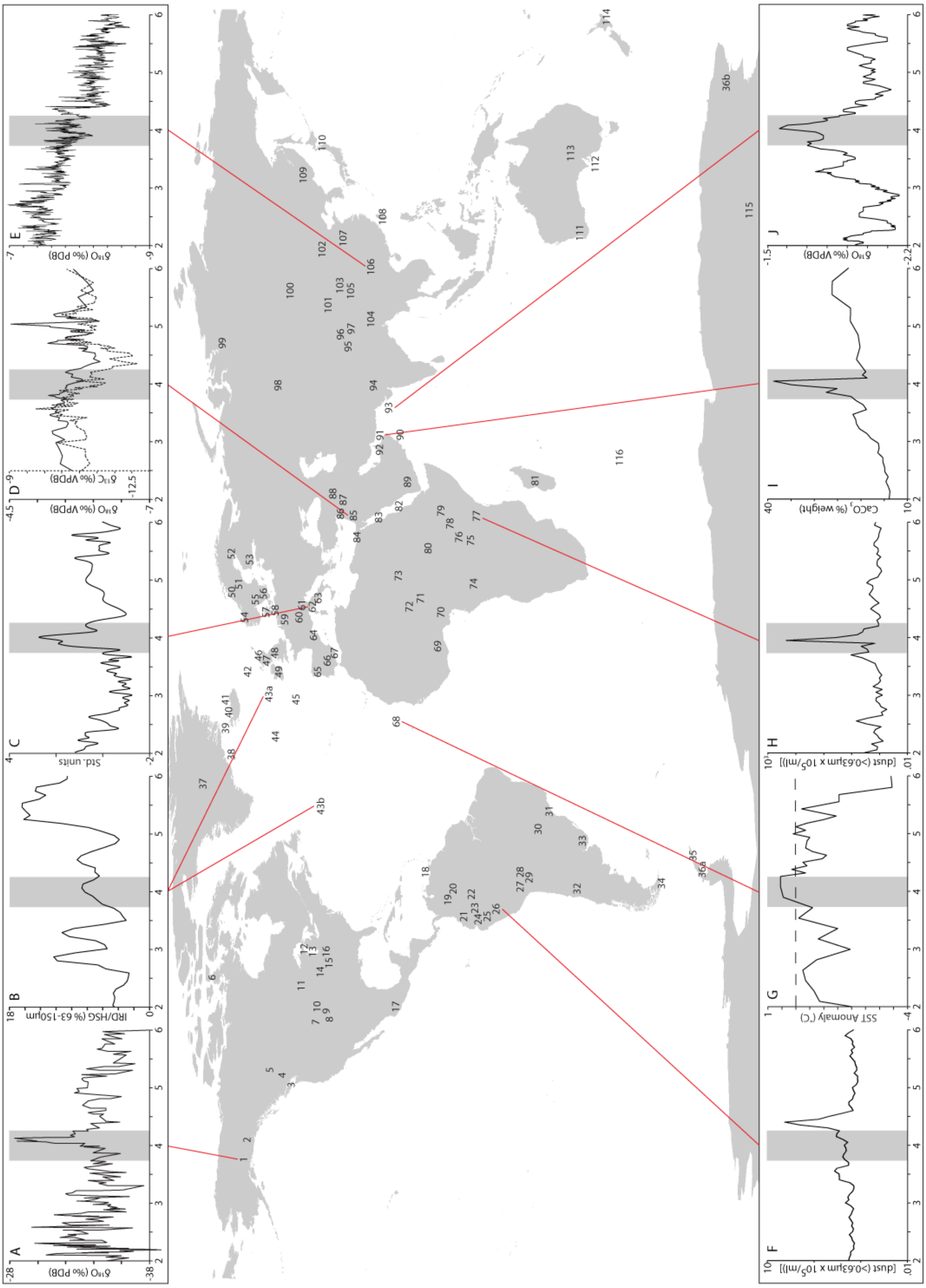


Figure 2.2 Evidence for a global 4.2 kyr event. Sites outlined in Table 2.2 (numbered), with selected key proxy records (a-j) presented on an age axis (2000 – 6000 cal yr BP). a) $\delta^{18}O$ data from the PRCol ice core, Mt Logan, Yukon territory, Canada (Fisher *et al*., 2008; Fisher, 2011); b) Ice-rafted debris events from the North Atlantic, as recorded by haematite-stained grains in marine sediments (Bond *et al*., 2001); c) Stacked data (inc. $\delta^{18}O$, $\delta^{13}C$, Mg-Ca and Fluorescence) from a cave flowstone in Northern Italy (Drysdale *et al*., 2006); d) $\delta^{18}O$ and $\delta^{13}C$ from a speleothem in Soreq Cave, Israel (Bar-Matthews *et al*., 2003); e) $\delta^{18}O$ from a speleothem in Dongge Cave, China (Wang *et al*., 2005); f) levels of dust recorded in the Huascarán ice-core, Peru (Thompson *et al*., 2002) displayed on a log-scale due to the magnitude of the peaks.

Figure 2.2 (continued) g) Mean SST anomalies from Atlantic marine records off the coast of West Africa (deMenocal *et al.*, 2000); h) levels of dust recorded in the Kilimanjaro ice-cores, Tanzania (Thompson *et al.*, 2002), displayed on a log-scale due to the magnitude of the peaks; i) CaCO3 and percentage dolomite of Mesopotamian origin in marine sediments from the Gulf of Oman (Cullen *et al.*, 2000); j) Planktonic foraminifera $\delta^{18}O$ values from marine sediments in the Indus Delta (Staubwasser *et al.*, 2003).

2.5.2 South America

A distinct increase in dust content in the Nevado Huascarán ice core, northern Peru, occurs c. 4500 cal yr BP (Thompson *et al.*, 2000; Davis and Thompson, 2006), indicating the presence of a significant arid event, apparently prior to that of North America. Subsequently, the period associated with the 4.2 kyr event is generally indicative of a wet phase, as demonstrated in pollen records from Colombia (Behling *et al.*, 1999; Marchant *et al.*, 2001), south-central Brazil (Alexandre *et al.*, 1999; Behling *et al.*, 1993) and the central Chilean coast (Maldonado and Villagran, 2006). Lake-level transgressions are also documented at Lake Titicaca (Cross *et al.,* 2000; Mourguiart, 2000; Baker *et al.*, 2001) and in Ecuadorian Amazonia (Weng *et al.*, 2002). Furthermore, wet conditions are indicated by increases in fluvial activity in Colombia (Velez *et al.*, 2001, 2005), stalagmite growth in central Brazil (Bertaux *et al.*, 2002) and lacustrine $\delta^{18}O$ data from Bolivia (Abbot *et al.*, 2000). In southern South America, records suggest a shift to drier and/or windier conditions, as indicated by multi-proxy records from Uruguayan wetlands (Iriarte *et al.*, 2004) and atmospheric dust concentrations in a raised bog in Tierra del Fuego (Sapkota *et al.*, 2007).

2.5.3 Africa

Gasse and van Campo (1994) were among the first to identify a dry period in Africa c.4 ka BP, when a study in the northern monsoon domain revealed simultaneous lowering of lake-levels at this time. In a further review of hydrological changes in the African tropics, from the lake records of central and eastern Africa, Gasse (2000) found two major dry spells, more notable than other hydrological fluctuations in the Holocene, occurring 8400 - 8000 cal yr BP and 4200 - 4000 cal yr BP. The former of these two events is clearly aligned with the 8.2 kyr event (Alley and Ágústsdóttir, 2005), and the latter, with the 4.2 kyr event discussed in this review. Evidence of increased African aridity and aeolian activity during both these time periods is found both within the continent in northern Nigeria (Holmes *et al.*, 1999), and as wind-blown sediment in records from Mediterranean Europe (Magri and Parra, 2001).

Evidence for reduced Main Nile flow is found in the lakes of East Africa (Butzer *et al.*, 1972; Said, 1993; Fraedrich *et al.*, 1997) and the Nile delta (Stanley *et al.*, 2003), thus supporting archaeological hypotheses of climate-induced societal disruption and collapse in the Old Kingdom in Egypt (Hassan, 1997). Lake-level regression is evidenced from across the region indicated, including Kenya (Johnson, 1996), the Congo/Uganda (Russell and Johnson, 2005), Ethiopia (Lamb *et al.*, 2000), Ghana (Talbot *et al.*, 1984) and many of the large lakes in the rift valleys of central and east Africa (Gasse and Descourtieux, 1979; Damnati, 2000). In addition, palynological records indicate an arid event in records from Madagascar (Gasse and van Campo, 1998), the Congo (Schwartz *et al.*, 1996) and Cameroon (Maley and Brenac, 1998).

The 4.2 kyr event represents the period of greatest historically recorded drought in Africa, leading Thompson *et al*. (2002) to describe it the 'First Dark Age'. A significant depletion in $\delta^{18}O$ from the Kilimanjaro ice-core record is coincident with an exceptional peak in the major aerosol species, including a 30mm thick dust layer, and interpreted by Thompson *et al*. (2002) as indicating cooler and drier conditions not only in Africa, but broader scale trends in tropical temperature.

2.5.4 Western Asia

Echoing Thompson *et al*. (2002), Frumkin (2009, p. 319) describe the "*most severe long-term historical drought [to affect] the region in the mid-late Holocene*". The Dead Sea is one of the largest hydrological systems in the Levant and, as a terminal lake in an arid environment, it is extremely climatically sensitive. Frumkin (2009) observe the largest fall of the Dead Sea level during the mid-late Holocene, based on a comprehensive review of regional lake-level studies (Kadan, 1997; Frumkin *et al*., 1991, 2001; Enzel *et al*., 2003; Bookman (Ken-Tor) *et al*., 2004; Migowski *et al*., 2006). In addition, Frumkin (2009) present a dendrochronologically derived $\delta^{13}C/\delta^{15}N$ record from the Dead Sea region, indicating extreme drought c. 4265 - 3930 cal yrs BP. It has been suggested that this drought in the Dead Sea catchment resulted in the entire southern basin drying up (Neev and Emery, 1967) and the northern basin becoming distinctly more evaporitic (Migowski *et al*., 2006). Bar-Matthews *et al*. (2003) also found evidence for a significant reduction in regional precipitation based on $\delta^{18}O/\ \delta^{13}C$ data from a speleothem deposit from Soreq Cave, Israel. Evidence of coincident increased evaporation rates and precipitation minima have been found elsewhere in the region, including the Red Sea (Arz *et al*., 2006), Yemen (Wilkinson, 1997) and the United Arab Emirates (Parker *et al*., 2006). Vegetational shifts towards xerophytic taxa are also documented in the climatically sensitive regions of coastal Syria (Kaniewski *et al*., 2008).

A lake-level regression of 30-60m was documented at Lake Van, located in the headwaters of the Tigris and Euphrates rivers (Lemcke and Sturm, 1997). This supports the archaeological hypotheses of Weiss *et al*. (1993), discussed earlier, that the abandonment of the once fertile Habur plains, located between these two rivers, was climatically driven. In addition, marine cores from the Gulf of Oman register an abrupt increase in aeolian dust of Mesopotamian origin, lasting c. 300 years (Cullen *et al*., 2000). Furthermore, tephra shards found in these marine sediments were geochemically similar to shards found at the Tell Leilan archaeological site (Weiss *et al*., 1993), providing further evidence of temporal synchroneity. Further evidence, collated by Weiss and Bradley (2001) and Staubwasser and Weiss (2006), demonstrates that drought and societal disruption was, at that time, a common theme across much of southwestern Asia and the eastern Mediterranean, so much so that Frumkin (2009) refers to the period as the 'Intermediate Bronze Age climatic crisis'.

2.5.5 Eastern Asia

Abundant evidence of societal disruption during the 4.2 kyr period permeates the archaeological record of eastern Asia, and is thought to be the consequence of the severe climatic deterioration seen in other parts of the world at this time (Yasuda *et al*., 2004). Comprehensive reviews of societal disruption and collapse are offered by Yasuda (1997), An *et al*. (2005) and Liu and Feng (2012) and are therefore not repeated here.

These archaeological hypotheses are supported by a wealth of palaeoclimatic evidence indicating a significant climatic event from across the region. Here, the 4.2 kyr event generally correlates with a period of increased aridity, largely attributed to a weakening of the monsoonal circulations which dominate regional meteorology (Guo *et al*., 2000). Hong *et al.* (2003) present a $\delta^{13}C$ time series from Hongyuan peatland, Tibet, demonstrating marked reductions in precipitation c. 4 kyr BP, indicating a weakening of the Indian Ocean monsoon. Coincidently, a weak monsoon event is registered in $\delta^{18}O$ data from the Dongge cave speleothem record in Southern China (Wang *et al*., 2005). Conversely, Taiwanese palaeoprecipitation records demonstrate a shift reflecting a strengthening of the East Asia summer monsoon c. 4200 cal yr BP, following a prolonged period of reduced precipitation, c. 5000 - 4200 cal yr BP (Yang *et al*., 2011). Grain size analyses in the lake sediments of Daihai Lake, north-central China, as a proxy for monsoon precipitation, also demonstrate a marked reduction during the 4.2 kyr event (Peng *et al*., 2005).

Corroborating this pattern is evidence of a distinct cool/dry event seen in palynological records from the Yishu River basin, Shandong, China (Gao *et al*., 2007); the western Loess Plateau (An *et al*., 2005); Yinchuan, China (Jiang *et al*., 2008); and western Siberia (Pitkanen *et al*., 2002). In addition, lacustrine proxy records have also exhibited evidence of a cool dry period in Tibet (Gasse and van Campo, 1994; Wei and Gasse, 1999; Mischke and Zhang, 2010; Mügler *et al*., 2010), northern Russia (Laing and Smol, 2003) and central Mongolia (Schwanghart *et al*., 2008). Fluvial sediments from the middle reaches of Yellow River also demonstrate evidence of a significant dry event, punctuated by a series of extreme floods (Huang *et al*., 2010, 2011). Evidence of reduced temperatures is also present in Japanese marine (Kawahata *et al*., 2009) and lacustrine sediments (Kato *et al*., 2003).

Significant societal disruption, again a theme across the region, is also well documented in the Indus Valley of west Pakistan and northwest India, driven by the onset of drought conditions, as recorded in $\delta^{18}O$ data from the marine sediments of the Indus delta (Staubwasser *et al*., 2003), and lacustrine palaeoprecipitation proxy data from northwest India (Prasad and Enzel, 2006). Notably, Berkelhammer *et al*. (2011, 2012) present a 4.2 kyr event signal, similar to that of the Mount Logan ice-core (Fisher *et al*., 2008; Fisher, 2011), in a speleothem isotope record from Mawmluh Cave, northeast India, which is interpreted as a proxy for changes in past monsoonal conditions. Walker *et al*. (2012) subsequently propose that this record should provide the GSSP for the 4.2 kyr event, and therefore the Middle-Late Holocene boundary, arguing that its low-latitudinal location is in keeping with the concentration of evidence for the event.

2.5.6 Europe

Perhaps owing to the relatively high density of palaeoclimatic studies conducted in Europe the climatic signal c. 4200 cal yr BP is somewhat more complex than in other areas of the globe. However, two broad patterns can be extracted from the palaeoclimatic data: a dry event in southern Europe and a wet event in northern Europe.

Drysdale *et al*. (2006) were the first to explicitly link aridification in Europe, c. 4200 cal yr BP, with the multi-centennial drought responsible for the collapse of Old World civilisations. They found evidence for a severe drought in an Italian calcite flowstone when stacked stable isotope, trace element and organic fluorescence data showed a brief period but extreme excursion, indicating reduced moisture availability.

Pseudoschizaea, a non-pollen palynomorph indicative of arid conditions (Scott, 1992; Carrión and Navarro, 2002), were found in southern Spain, alongside palynological evidence of a shift to drier conditions and expansion in the Mediterranean scrub ecosystem (Carrión, 2002; Pantaléon-Cano *et al*., 2003). Similarly, lacustrine diatom assemblages from northwest Spain indicate a shift to lower lake productivity and cold water conditions at c. 4200 cal yr BP, alongside other noted periods of Holocene climatic change, c. 8200, 6000 and 2600 cal yr BP (Leira, 2005). Further palynological evidence of shifts in vegetation cover, indicative of regional arid episodes, has been presented by Jalut *et al*. (2000), including a phase which commenced c. 4300 cal yr BP.

Magny *et al*. (2002) identified a number of phases of decreased river activity in the western Mediterranean region, one of which occurred c. 4000 cal yrs BP. This period is subsequently correlated, by the authors, with other indicators of a coincident dry event, including glacial retreat in the northern Alps (Röthlisberger, 1986). However, a prolonged lake-level transgression in the Jura Mountains, eastern France (Magny, 1999, 2004) and central Italy (Magny *et al*., 2007) are also documented, c. 4200 cal yr BP. After comparison with the atmospheric residual ^{14}C record from tree-ring records (Stuiver *et al*., 1998), Magny *et al*. (2007) suggest that higher lake-level conditions developed in both records in response to cooling periods associated with phases of decreased solar activity, and in some early Holocene cases (e.g. 8.2 kyr event) to meltwater pulses during the final stages of deglaciation.

A wet phase, c. 4200 cal yr BP, is recorded in a number of records in northern Europe, particularly in the region's peatlands, which are a major source of palaeoclimatic information. In a review of the wet shifts recorded in this archive, Hughes *et al*. (2000) identify a number of records that indicate a climatic change to wetter conditions at a similar time to their own plant macrofossil sequence from Walton Moss, northern England, c. 4400 - 4000 cal yr BP. These included records from elsewhere in northern England (Barber *et al*., 1994), as well as southern Scotland (Tipping, 1995), south Cumbria (Wimble, 1986), the Netherlands (van Geel, 1978; van Geel *et al*., 1996), Finland (Korhola, 1995), Denmark (Aaby, 1976), and central and northern Norway (Nilssen and Vorren, 1991), although considerable variation in chronological quality prevents the presentation of a more precise date. This broad regional trend is supported by the findings of Barber *et al*. (2003) who, comparing radiocarbon dated plant macrofossil records from three oceanic raised bogs in Great Britain and Ireland, were able to identify a prominent and broadly coincident wet event 4400 - 4000 cal yr BP. Caseldine *et al*. (2005) describe this body of evidence as a strong indication of a synchronous deterioration in climate during this period.

More recently, a number of peat humification records from south-central Sweden have demonstrated similar patterns (Borgmark, 2005; Borgmark and Wastegård, 2008). Similarly, Mauquoy *et al*. (2008) identified wet shifts c. 4100 cal yr BP at sites in northern England and Denmark, based on plant macrofossil and testate amoebae data. The review presented by Hughes *et al*. (2000) found evidence for coincident wet shifts in a multitude of palaeoenvironmental archives from NW Europe including: lake-level and sedimentation data (Pennington *et al*., 1972; Digerfeldt, 1988; Yu and Harrison, 1995), blanket mire humification records (Blackford, 1990; Nilsson and Vorren, 1991; Anderson *et al*., 1998), tree-line studies (Gear and Huntley, 1991) and peat initiation archives (Korhola, 1995). Gunnarson *et al*. (2003) and Gunnarson (2008) also present dendrochronologically dated evidence for a lake-level highstand during the period.

In addition, peat records in southwest Sweden provide evidence of increased winter storminess during the 4.2 kyr event, as indicated by an increased presence of aeolian sediment in raised bog deposits (Bjork and Clemmensen, 2004; de Jong *et al*., 2006). Clemmensen *et al*. (2009) found corroborative evidence for such an event from the same region, presenting a combined ^{14}C and optically stimulated luminescence (OSL) chronology of dune formation, resulting from high energy events, followed later by dune stabilisation and soil formation. Lacustrine sediments from the Faroe Islands also contained evidence for increased aeolian activity. Andresen *et al*. (2006) present sedimentological, geochemical and magnetic susceptibility data documenting a marked climatic deterioration c. 4200 cal yr BP. Sejrup *et al* (2001) present $\delta^{18}O$ data from the marine sediments in western Norway, indicating a c. 2°C sea surface temperature (SST) drop, c. 4200 cal yr BP. This also coincides with a low stand (c. -1.9m) in sea levels, reconstructed from sedimentological analyses in Skagen Odde, Denmark (Clemmensen *et al.* 2001).

At Sanday, Orkney, a combination of ^{14}C and OSL dating was used to estimate the age of sand layers, overlain by organic soils, indicative of periodic enhancement of aeolian activity (Sommerville *et al*., 2007). Ages of c. 4000 cal yr BP were found to be consistent with the abandonment of the famous Skara Brae archaeological site, suggestive of a potential link between the increased aeolian activity recorded at Sanday and societal disintegration in the area. In fact, Tipping and Tisdall (2004) describe this period as one of climate change-driven crisis for many early Bronze Age communities across northwestern Europe; a theory supported by Magny *et al*. (2012).

2.5.7 North Atlantic

The influence of ocean-atmosphere circulation changes in the North Atlantic on the climate deteriorations in northern Europe is well documented (e.g. McDermott *et al*., 2001, Turney *et al*., 2005, Charman *et al*., 2006; Marshall *et al*., 2007; Jonsson *et al*., 2010), and is discussed with particular reference to the region's peatlands in Chapters 3 and 4. However, links between the North Atlantic and climate change associated with the 4.2 kyr event have been proposed further afield, as far away as China (Gupta *et al*., 2003; Hong *et al*., 2003).

Bond *et al*. (1997) identified a number of ice-rafted debris (IRD) events in North Atlantic marine sediments, characterised by lithic and petrologic changes indicative of increased ice-rafting, which reflect a cooling of the ocean surface, caused by substantial changes in North Atlantic surface circulation; one of which occurs c. 4000 cal yr BP. This event's timing and likely spatial extent of influence has led some to suggest that it should be considered synonymous with the 4.2 kyr event (Liu and Feng, 2012).

Other marine records in the region exhibit evidence from a range of proxies for a significant cool event in the North Atlantic, coincident with the 4.2 kyr event, including studies from the East Greenland shelf (Jennings *et al*., 2002), the Denmark Strait (Andresen and Bjorck, 2005), the northern Icelandic shelf (Castaneda *et al*., 2004; Ran *et al*., 2008), the northern Icelandic basin (Hall *et al*., 2004) and the abyssal northeast Atlantic (Smart, 2008).

2.5.8 Southern Hemisphere

Outside the main continental landmasses of South America and Africa, evidence for the 4.2 kyr event in the Southern Hemisphere is sparse. However, it is likely that this is a sampling artefact, reflecting the relative bias of palaeoclimatic studies the Northern Hemisphere has enjoyed until relatively recently. Evidence of pronounced warming, c. 4200 cal yr BP, has been found across Antarctica, in both the Antarctic Peninsula (Hjort *et al.*, 1998, Ingolfsson *et al.*, 1998) and Victoria Land (Ingolfsson *et al.*, 1998). Furthermore, Baker *et al.* (2005) found rising sea levels and warmer-than-present SST occurring simultaneously in southwestern and southeastern Australia at approximately 4200 cal yr BP. However, the EPICA ice core appears to suggest a cooling during this period (Masson-Delmotte *et al*., 2004).

Peatlands in the sub-Antarctic Iles Crozet have also demonstrated a coincident shift to drier conditions (van der Putten *et al.,* 2008). Quigley *et al*. (2010) interpreted mid-Holocene drying, as witnessed in speleothem-derived isotopic data from southern Australia, as evidence for the onset of 'modern' El Niño-Southern Oscillation (ENSO) dominated ocean-atmosphere interactions conditions in the Southern Hemisphere.

This suggestion is supported by multi-proxy evidence from neighbouring New Zealand, which indicates increased storminess, characteristic of prevailing ENSO dominance (Gomez *et al*., 2004). Furthermore, palynological evidence from the Australian lowlands indicates the existence of a multi-centennial dry shift c. 4000 cal yr BP (Shulmeister and Lees, 1995), again interpreted as the Holocene onset of ENSO conditions.

2.6 Key patterns

By plotting the evidence discussed in Section 2.5 it is possible to highlight a number of key climatic patterns, associated with the 4200 cal yr BP period (Fig. 2.3). The 4.2 kyr event can therefore be characterised by clear drying in much of the mid-latitudes, across Africa, Asia and North America. Northern Europe is dominated by wet and stormy conditions and, as in the North Atlantic, there is evidence of a significant cooling event. A subtle drying trend may be present on the western coast of South America, but overall the continent is dominated by a wetting trend. Elsewhere in the Southern Hemisphere coherent trends are more difficult to characterise, highlighting the need for further work in the region.

As different regions and different climate archives are likely to have been affected and/or responded differently it is very difficult to assign confidently a likely duration of the 4.2 kyr event. It could be argued that the somewhat arbitrary nature of the 4.2 kyr event's title, in terms of its chronological implications (see Section 2.5), persist in the discussion of its duration. The majority of studies that explicitly discuss the 4.2 kyr event present it as a multi-centennial climatic event, with this assumption based largely on initial definitions of the event as a period of aridification in western Asia, lasting approximately 200 to 300 years (cf. Weiss *et al.*, 1993; Cullen *et al.*, 2000; Weiss and Bradley, 2001; Staubwasser and Weiss, 2006). However, as demonstrated in Table 2.2, for the majority of records from which an estimate the event's duration can be derived this estimate appears reasonable, with specified durations averaging c. 200 years. Clearly, where climatic variations associated with the 4.2 kyr event appear to have triggered the beginning of a climatic or environmental trend this estimated duration is not applicable. Subsequently, the extraction of robust dates for the 'start', 'end' or 'duration' of an event from palaeoclimatic archives is a major conceptual issue and is a theme discussed further in Section 9.6.5.

2.7 Potential causes

Examining palaeoclimatic studies of the period c. 4200 cal yr BP event provides an opportunity to study the causes and effects a globally abrupt climatic change, which occurred against background conditions similar to those today. In short, the analogy provided by the 4.2 kyr event has the potential to enhance our understanding of past climate dynamics and response to past perturbations, which will help inform our understanding of potential future changes through comparison with model simulations.

As outlined earlier, alongside its global extent and societal repercussions, much of the 4.2 kyr event's climatic significance lies in its occurrence in the absence of major ice-sheets and their associated forcing mechanisms (e.g. ice-sheet surging, glacial meltwater outburst floods). Consequently, a number of mechanisms have been postulated as to its origins, but as yet no consensus has been reached. Booth *et al.* (2005) outlined four broad possibilities as to the cause of the 4.2 kyr event, including non-linear responses to Milankovitch forcing, solar variation, volcanic events and variability in the ocean/atmosphere circulation system.

Whilst variation in Milankovitch forcing was minimal during the mid- to late-Holocene, it is still possible that base-state climate could have been affected by changes in orbital forcing. However, climate models which deal with variation in such factors (i.e. those which assess the likely effect of changes orbital forcing on the magnitude and frequency of abrupt decadal- to centennial-scale changes) have not been exercised to a level necessary upon which to base conclusions (Booth *et al.*, 2005).

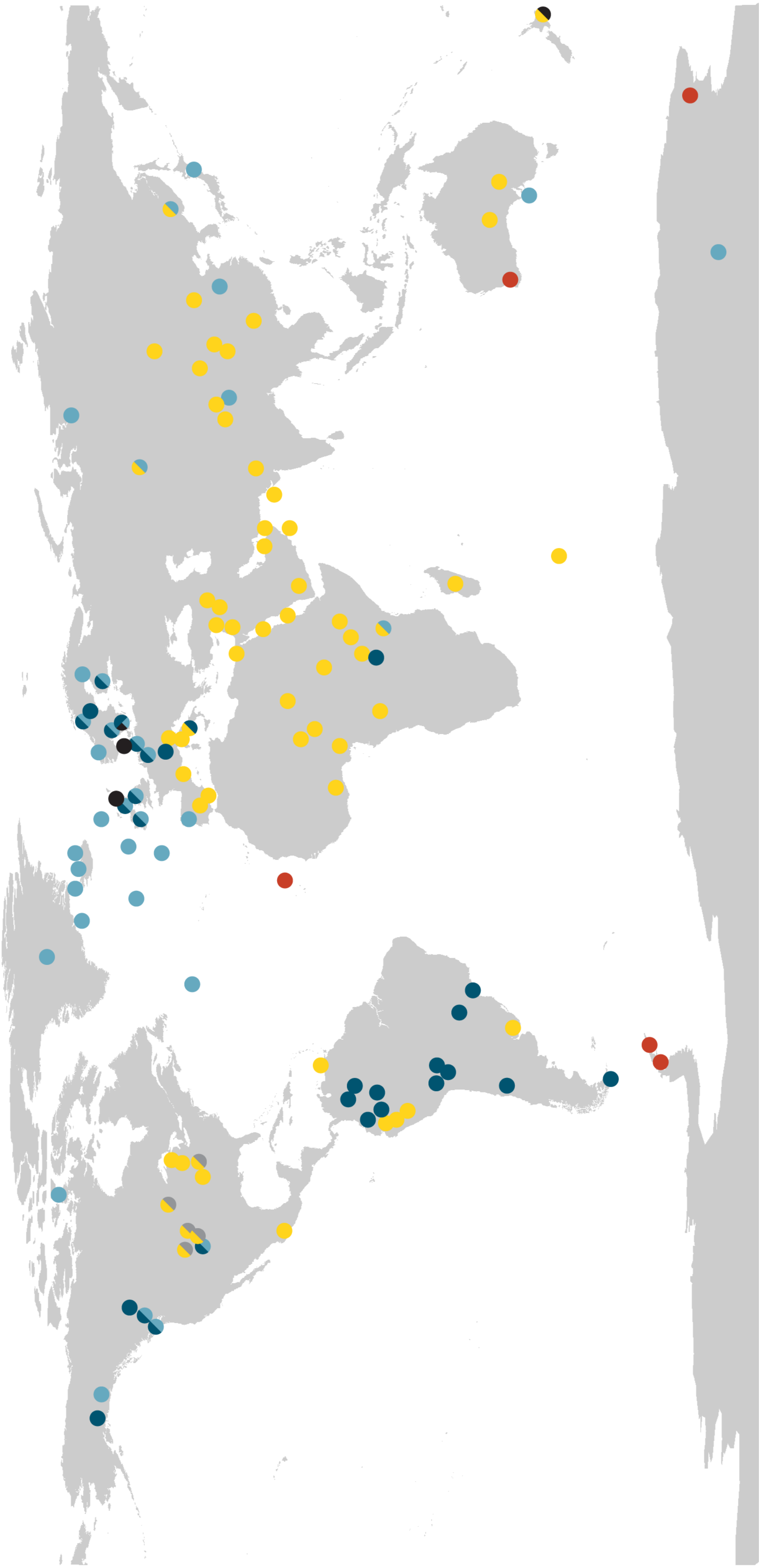


Figure 2.3 Global trends during the 4.2 kyr event. Dots correspond to sites shown in Figure 2.2 and detailed in Table 2.2. Yellow = drying; dark blue = wetting; light blue = cooling; red = warming; grey = increased aeolian activity; black = increased storminess.

Bond *et al*. (2001) suggest that a change in solar forcing occurred c. 4200 cal yr BP, indicated by correspondence between the IRD 3 cold event in the North Atlantic and an increase in cosmogenic isotopes, which are indicative of reduced solar radiation. In addition, Staubwasser *et al*. (2003) link late-Holocene drought cycles in southern Asia to cosmogenic ^{14}C production rates. Links have also been made between BSW in northern Europe and solar activity, for changes similar to those seen in the region c. 4200 cal yr BP (e.g. Chambers and Blackford, 2001; Blaauw *et al*., 2004; Plunkett *et al*., 2008). However, as Booth *et al*. (2005) describe, climate models, which focus on climate system response to variation in solar isolation, tend to focus on temperature and/or circulation responses, rather than precipitation. The 4.2 kyr event, however, appears to be a hydroclimatic event, as opposed to a temperature-based event. When changes in precipitation have formed the focus for these models, they have tended to extend only to tropical areas (e.g. Meehl *et al*., 2003).

Whilst Bryson (1988) notes an increase in volcanic activity at this time, no global-scale volcanic signal has yet been found in the world's ice-core records (Zielinksi, 2000), although high-resolution studies of these records exist only for the last thousand years (Crowley, 2000; Ammann and Naveau, 2003). Again, global climate models do not yet sufficiently assess the effect of increased volcanism on large-scale climatic changes, including variation in precipitation, but rather focus on temperature and circulation (e.g. Amman *et al*., 2003; Shindell *et al*., 2003; Rind *et al*., 2004). Although, it has recently been suggested that the onset of the Little Ice Age was triggered by volcanism and sustained by ocean feedbacks (Miller *et al*., 2012), suggesting further research is required with regard to the role of volcanism in other Holocene ACC events.

2.7.1 Ocean-atmosphere circulation and the importance of the North Atlantic

Perhaps most plausibly then, many of the explanations as to the 4.2 kyr event's origins offered by studies thus far focus on variations in ocean-atmosphere circulation. Indeed, Marchant and Hooghiemstra (2004) suggest that given the apparent uniformity in the timing of this event, together with its wide spatial extent, we must look to the large-scale aspects of general ocean-atmosphere circulation for explanations. The global ocean-atmosphere circulation system is made up of complex interactions between a large number of variables, including, among others, the Arctic Oscillation/North Atlantic Oscillation (AO/NAO), Thermohaline Circulation (THC), migration of the Intertropical Convergence Zone (ITCZ), ENSO, variation in SST, the westerlies and various monsoon systems. Recently, climatic changes in the North Atlantic have emerged as a fundamental component of many of these ocean-atmosphere circulation sub-systems, asserting the importance of palaeoclimatic study here beyond time periods associated with the region's continental ice sheets and the subsequent deglaciation.

One of the key trends during the 4.2 kyr event was continental scale aridification in the world's mid-latitudes (Booth *et al*., 2005, Staubwasser and Weiss, 2006). The majority of major modern mid-latitude droughts have been linked to internal variability of, and interaction between, the Earth's oceanic and atmospheric systems (e.g. Enfield *et al*., 2001; McCabe *et al*., 2004; Schubert *et al*., 2004). Links between SST and drought, for example, have been made in the instance of the global 1998-2002 drought. During this event, SST changes resulted in persistent high pressure over the Northern Hemisphere's mid-latitudes, causing widespread synchronous drought (Hoerling and Kumar, 2003). Interestingly, this period also witnessed increased wetness in some areas in the mid- to high-latitudes, in a broadly similar pattern to that of the 4.2 kyr event. Whilst not on the temporal scale of the 4.2 kyr event the 1998-2002 drought clearly shows that such processes are capable of causing significant climate change across broad regions. Gupta *et al.* (2003) also found a correlation between SST variation and the Indian Ocean Monsoon, a potentially key circulatory system with regard to the 4.2 kyr event. Similarly, links have been demonstrated between timing of Asian monsoonal variations and climatic events in Africa (Guo *et al.*, 2000). Cooler temperatures in the North Atlantic, such as those experienced during IRD events (Bond *et al.*, 1997), have also been linked to drought in North America during the Lateglacial, Holocene and in GCM results (Benson *et al.*, 1997; Peteet *et al.*, 1997; Yu *et al.*, 2002). Interestingly, historical observations suggest that the opposite relationship is more accurate, with warm phases in the Atlantic multi-decadal oscillation resulting in drought in North America (Enfield *et al.*, 2001; McCabe *et al.*, 2004). Regardless, stable isotope analysis of a speleothem in Iowa indicates a brief negative spike in $\delta^{18}O$ at 4.1 ka BP (Denniston *et al.*, 1999), which has been interpreted as indicating a shift in the seasonality of precipitation and an atmospheric change to more distal sources of moisture, typical of drought climatology (Woodhouse and Overpeck, 1998). Widespread lake-level minima in the northern monsoon domain of Africa suggest a weakening of monsoonal rainfall in the region, which could realistically be related to cooler SSTs in the North Atlantic (Gasse and van Campo, 1994; Gasse, 2000). Harrison *et al.* (2003) state that drought in these regions can be linked with the summer insolation cycle, land-ocean temperature differences, large-scale circulation of the ocean and the occurrence of strong summer and winter precipitation maxima. Weakening in the Asian monsoon systems has also been linked to IRD events in the North Atlantic (Gupta *et al.*, 2003; Hong *et al.*, 2003).

Cooler tropical SSTs would result from a reduction in deep-water formation in the northern North Atlantic (i.e. during IRD events; Bond *et al.*, 1997, 2001), possibly resulting from a reduction in the northward advection of equatorially warm water (Nyberg *et al.*, 2002). Precipitation regimes in the west African tropics are strongly influenced by variation in Atlantic SST (Stager and Anfang-Sutter, 1999). It has been suggested that cooler SSTs here would result in an enhancement of the north-south pressure gradient and a reduction marine evaporation (Stager *et al.*, 2002), thus leaving less atmospheric moisture for convective precipitation. However, as Figure 2.2g demonstrates, Atlantic SSTs off the coast of tropical west Africa reached a mid-Holocene peak c. 4200 - 4000 cal yr BP, during which time major regional drought is well documented (Gasse and van Campo, 1994; Gasse, 2000; Damnati, 2000). In addition, Baker *et al.* (2001) have shown that cold high-latitude SSTs can be associated with contemporaneous increases in moisture in tropical South America, characteristic of the 4.2 kyr event (see Section 2.5.2). deMenocal (2000) suggests a degree of synchroneity between this SST record and the IRD event record from the North Atlantic (Bond *et al.*, 2001). However, this correlation appears to collapse during the period associated with the 4.2 kyr event. It is clear, however, that the relationship between SSTs and regional precipitation changes, although not fully understood, must be considered carefully as a potential driver, or contributory factor, of the 4.2 event.

Higher NAO values have been linked with above average aridity in northwestern Africa (Hurrell, 1995). Tourre *et al*. (1999) have also shown that NAO can be linked with tropical SSTs in the South Atlantic. In a discussion of the 4.2 kyr event in the eastern Mediterranean, Arz *et al*. (2006) suggest that changes seen in the regions influenced by the Asian and African monsoonal systems as well as in Northern Hemisphere mid-latitudes, can all be linked to long-term change in the AO/NAO. The NAO system has also been found to be a key modulator of the THC system, and subsequently of ocean heat transfer in Atlantic (Marshall *et al*., 2007). Furthermore, Smart (2008) document a major decrease in the number and accumulation rate of planktic and benthic foraminifera in the abyssal northeastern Atlantic, symptomatic of increased dissolution of deep-sea carbonates, controlled by changes in ocean water chemistry. This has been linked to short-term shutdown of North Atlantic Deep Water (NADW) formation as a result of meltwater input to the eastern Atlantic during the Younger Dryas (Sarnthein *et al*., 1994).

The influence of the strength and position of the westerlies on prevailing climate in northern Europe is well documented (e.g. Charman and Hendon, 2000; Magny, 2004; de Jong *et al*., 2006; Blundell *et al*., 2008; Daley *et al*., 2010). During the 4.2 kyr event, evidence from the region is strongly indicative of an increase in westerly strength and subsequent aeolian activity (Bjork and Clemmensen, 2004; Sommerville *et al*., 2007; Clemmensen *et al*., 2009). However, the influence of these variations is more widespread. Staubwasser and Weiss (2006) suggest that a displacement of the Mediterranean westerlies, and possibly the Indian summer monsoon, could explain reductions in seasonal precipitation in western Asia. In addition, Staubwasser and Weiss (2006) assert that precipitation patterns in the subtropical latitudes of Asia, Africa and the Mediterranean are significantly affected by Rossby wave pattern in upper-level westerly flow, related in turn to the intensity of tropical convection, thus providing a link between the climates of eastern Africa and the Middle East. Furthermore, the authors suggest that changes in southern east African tropical convection intensity c. 4200 cal yr BP could realistically have altered upper level flow and wave pattern in northern hemispheric subtropical regions, and in particular the Middle East.

Marchant and Hooghiemstra (2004) have also suggested that the teleconnections seen between Africa and South America during the 4.2 kyr event, can be explained by variation in the ITCZ system. As previously outlined, Thompson *et al*. (2002) highlighted a broad coincidence of large dust peaks found in the ice-core record from the Nevada Huascarán, northern Peru, and the Kilimanjaro ice-core, suggesting they were the result of a global climatic event. However, Davis and Thompson (2006) date the South American dust peak to c.4.5 ka BP, suggesting that Africa and South America were experiencing contrasting climatic conditions during the 4.2 kyr event, assuming both sets of dates are accurate. It also suggests that the wetness displayed in other records from South America during the 4.2 kyr event was preceded by a dry period. Indeed, Marchant and Hooghiemstra (2004, p.234) came to similar conclusions, in a review of palaeoclimatic change in the tropics of Africa and South America around 4 ka BP. The authors state that "*in both hemispheres of the South American tropics, there was a major synchronous change to increased precipitation and/or decreased evaporation, possibly due to a shorter dry season, about 4000 years BP. In comparison to South America, the environmental shift in equatorial Africa is towards dry conditions, this is likely to reflect* *reduced precipitation, increased evaporation, and/or an extension of the dry season*". It has been demonstrated that the ITCZ varied synchronously with North Atlantic climate over a range of timescales throughout the Holocene and Last Glacial Maximum (LGM) (Chiang *et al*., 2003), and it is possible that the time-transgressive nature of this dry period may have been caused by alternating climatic conditions resulting from ITCZ migration. Subsequently, Marchant and Hooghiemstra (2004) suggest that variation in North Atlantic ocean-atmosphere circulation could feasibly alter evaporation, moisture and precipitation regimes, and could, in large part, be responsible for the 4.2 kyr event. It should be noted, however, that this 300 year discrepancy could be the result of chronological error, which are potentially considerable for these records (Davis and Thompson, 2006).

In conclusion, evidence outlined here, and in Section 2.5, suggests it is feasible that variation in SST, NAO/AO, THC, ITCZ, ENSO, the westerlies and/or the various monsoon systems form a complex set of contributory drivers of the climatic change seen at 4200 cal yr BP. As a result, circulation changes in the North Atlantic appear to have played a major role in triggering, or at least amplifying the 4.2 kyr event, and other periods of ACC in the Holocene, even in the tropical realm (Marchant and Hooghiemstra, 2004). Whilst we cannot underestimate the ability of the southern Atlantic, and/or possibly ENSO, to force change in the North Atlantic, understanding of long-term ENSO dynamics is still poorly resolved. Despite this, Osborn *et al*. (1999) found that recently, there have been instances where the ENSO system has driven the NAO into a positive mode, resulting in warmer Northern Hemisphere winters. There are suggestions that the ITCZ and ENSO are linked in some way, as correlation has been shown between variation in ENSO and African precipitation (Cole *et al*., 2000; Diaz *et al*., 2001). Convection intensity in the southern East African tropical region responds to ENSO variability, as does the entire Indian Ocean-Western Pacific monsoon system (Meehl, 1994). In addition, it has also been found that weaker ENSO variability, and a correspondingly persistent La Niña can be linked with prolonged drought in palaeoclimate records of continental North America (Forman *et al*., 2001; Menking and Anderson, 2003), including the famous 'Dustbowl' event of the 1930s (Schubert *et al*., 2004).

Recent modelling work by Mann *et al*. (2009), however, demonstrates that we must think of the relationship between 'event' and 'driver' as one of dynamic response, with the authors demonstrating that the Little Ice Age and Medieval climate anomaly were most likely the result of such a response of climate to natural radiative forcing changes involving both ENSO and AO/NAO. In short, SST variation clearly plays a key role in the 4.2 kyr event, with the primary sources of atmospheric variability most likely being ENSO and NAO. However, what remains unclear is the precise nature of the inter-hemispheric and global teleconnections, which occur during the event. Oceanic circulation is clearly an active and highly non-linear component of the earth's global climate system (Rahmstorf, 2002) and interpretation of variation in its many integrated systems, interaction with atmospheric systems, and their subsequent effect on climate is extremely complex.

2.8 The potential of peatland palaeoclimate records in NW Europe at 4.2 kyr

The importance of the role played by changes in the North Atlantic during the 4.2 kyr event is clear. As palaeoclimatic archives strongly influenced by the North Atlantic (Charman *et al.*, 2006), the peatlands of northwest Europe possess significant potential for the examination of the 4.2 kyr event in terms of both, its regional timing and nature, and the role played by the North Atlantic in its origins.

The efficacy of palaeoclimatic studies examining abrupt climatic events is limited by a number of issues. When displayed as a function of time, palaeoclimatic records inherently exhibit anomalies from a mean value or trend. However, such anomalies can be caused by 'true signal', noise and/or error. In addition, there is frequently variation in the sampling and temporal resolution between palaeoclimatic records. Finally, there is significant disparity in the precision and accuracy of chronologies. Combined, these problems present significant challenges to those wishing to identify and correlate specific perturbations in the Earth's climate system, as represented in palaeoclimatic archives. In order to minimise these issues, it is proposed here that several criteria must be set for records that are intended for use in the examination of abrupt climate events. Therefore, records must: i) represent a climate dominated signal; ii) be high-resolution; iii) have a high-quality chronology; and iv) produce a semi-quantitative reconstruction. In addition, in order to ascertain the degree to which noise and/or error is masking the climatic signal at a site, it is preferable that multiple records from each region are prepared together (cf. Payne and Blackford, 2008), using comparative sampling, chronological and analytical strategies.

Whilst discussed in greater detail in the next chapter, it is important to note here that the ombrotrophic raised bogs of northwest Europe are able to meet these criteria. These systems are governed by a combination of precipitation and temperature, and therefore possess a strong climatic signal (Charman *et al.*, 2009). Recent developments in the field of peat-based palaeoecology have also enabled the quantitative estimates of past changes in prevailing hydrological conditions (e.g. Blackford and Chambers, 1993; Barber *et al.*, 1994; Charman *et al.*, 2007; Loisel *et al.* 2010). When combined with their rapid accumulation, wealth of datable organic material and incorporation of known-age volcanic tephra horizons, this enables the development of high-resolution palaeohydrological reconstructions with high-quality, robust chronologies. In addition, novel methodologies have been developed for the stable isotopic analysis of raised bog material, reflecting changes in the source of atmospheric moisture, which are strongly linked variation in atmospheric circulation patterns (Daley *et al.*, 2010).

Multi-proxy palaeohydrological reconstructions from peatlands in the region have already suggested the occurrence of a multi-centennial wet phase associated with the 4.2 kyr event (see Section 2.5.6). However, these studies lack the sampling resolution, chronological extent and temporal focus necessary to provide an accurate assessment of the 4.2 kyr event's manifestation in the region (e.g. Anderson *et al.*, 1998; Hughes *et al.*, 2000; Barber *et al.*, 2003; Mauquoy *et al.*, 2008). As a result, the 4.2 kyr event is comparatively underrepresented with studies focused on other similar climatic deteriorations relatively common (e.g. Little Ice Age, Mauquoy *et al.*, 2002a, Barber *et al.*, 2000; '2.8 kyr event', van Geel *et al.*, 1996, Blaauw *et al.*, 2004a, Swindles *et al.*, 2007a). In addition, studies of northwest European raised bogs have adopted a variety of methodological techniques and interpretive approaches, many of which present issues, reviewed in Chapter 3, for the evaluation of abrupt climatic events. Consequently, Section 3.5 presents a rationale for an optimal methodology for the examination of these climatic downturns, characterised by the development of comprehensive chronology and multi-proxy palaeohydrological analyses, including that of the stable isotopes $\delta^{18}O$ ands $\delta^{13}C$.

Finally, Walker *et al*. (2012) argue that the Blytt-Sernander scheme for the subdivision of the Holocene interglacial, developed in the peatlands of northwest Europe, is not applicable over anything more than regional spatial scales as a result of their association with radiocarbon chronologies and potentially time-transgressive vegetational responses. As a result, this scheme has been largely abandoned (see Section 3.3), and Walker *et al*. (2012) propose that their tripartite subdivision, using the 8.2 and 4.2 kyr events as GSSP boundaries, holds more validity (see Section 2.4). However, for this to be effective, these events must not only be readily identifiable, but universally definable within the palaeoclimatic archive. The ombrotrophic raised bog records of northwest Europe present an excellent opportunity to test this possibility in the same archives used to develop the original Blytt-Sernander model.

Chapter 3 – Peat-based palaeoclimatic research

3.1 Introduction

This chapter provides a synthesis of the current literature concerning the role of peatlands in palaeoclimatic research, with a brief overview of the history, current methods and future directions of peat-based palaeohydrological reconstructions. A rationale for the study of mid-Holocene abrupt climatic change in peatland archives is also presented.

3.2 Peatlands

In the context of the climatic trends presented by the IPCC in the 4th Assessment Report (2007), the potential of peatlands as carbon 'sinks', but also as sources of increased methane emissions, has raised the profile of peat-based studies. Barber and Charman (2003) observe that the global distribution of peatlands highlights their reliance on optimal moisture and temperature conditions and that, consequently, the occurrence, rate and nature of peat accumulation is clearly determined, at least in part, by climate. Scientific interest in peatlands stems both from their role in the global carbon and methane cycles, and their importance as archives of past climate change (de Jong *et al.*, 2010; Jackson *et al.*, 2010). It is the latter, which is the focus of this study.

The precise definition of peat and peatlands varies with context. Charman (2002) offers a detailed account of the terminology associated with various disciplines, whilst also providing a lexiconic framework for use in peat-based palaeoclimatic research. It is this framework that will be followed in this study. In the broadest context, peat can be described as forming wherever decay of plant material is exceeded by production (Barber and Charman, 2003). This is typically facilitated by the inhibition of decay owing to wet and/or cold conditions, coupled with the maintenance of a productive growing season. Peatland systems can be broadly divided into two main hydromorphological types. Ombrotrophic peatlands, or bogs, obtain water solely from precipitation. Water balance as determined by precipitation and evaporation is therefore largely a function of climate. Conversely, water supply in a minerotrophic peatland or fen, is dependent on surface run-off and groundwater as well as precipitation, and changes in water balance cannot therefore be as clearly linked to climatic variation.

Traditionally, palaeoclimate data from peatlands has taken one of two forms. The first concerns the location and timing of peat initiation, which is considered a proxy for a shift towards a colder and/or wetter climate (e.g. Korhola, 1990, 1994, 1995; Halsey *et al.*, 1998; Yu *et al.*, 2003; Gorham *et al.*, 2007). Such studies are able to employ a range of sites, minerotrophic and ombrotrophic. However, vegetation, substrate, local hydrological position and climate are all contributory factors to peat initiation and it is difficult to explicitly attribute peat initiation at a given site to a specific climatic deterioration (Barber and Charman, 2003). Regional records of simultaneous or time-transgressive peat initiation are more helpful evidence of climatic change (e.g. Halsey *et al.*, 1998).

The second form of peat-based palaeoclimate data, and also the focus of this study, documents the variations in peat composition and characteristics reflecting changes in precipitation and temperature over time (Barber and Charman, 2003). Ombrotrophic raised or blanket bogs are used as the basis for the majority of peat-based proxy palaeoclimate research, owing to the coupling of atmospheric water balance and prevailing climatic conditions within these systems.

3.2.1 Ombrotrophic raised bogs

Raised bogs frequently take precedence in palaeohydrological reconstruction studies due to their exceptionally high accumulation rates, when compared to blanket bogs. Raised bogs tend to experience relatively consistent accumulation at a rate of approximately 10-20 years/cm; compared to 20-50 years/cm in blanket bogs, within which accumulation rates can also vary considerably (Blackford, 2000; Barber and Charman, 2003).

Barber and Charman (2003) state that in order to fully understand the peat-based palaeoclimatic record, we must first understand the functioning of the peatland system in question. Peatlands have an upper 'active' layer and a lower 'inert' layer (Clymo, 1971, 1984; Belyea and Clymo, 2001). The upper layer, or 'acrotelm', possesses high hydraulic conductivity and a fluctuating water table. The majority of peat decay takes place in this relatively shallow layer, owing to its periodic aeration and biologically active state. The lower 'catotelm', corresponds to the permanently saturated main body of peat (Ivanov, 1948, 1953, 1981; Lopatin, 1949; Romanov, 1968). Decomposition is far slower in this zone, as a result of its anoxic nature.

Ombrotrophic raised bogs possess a distinct convex profile, which isolates them from the influence of the underlying groundwater. Their water supply is derived solely from the atmosphere, with surface runoff sometimes forming a lagg at the perimeter of the bog system, but not affecting the WTD at the bog surface. High-water tables in these systems above that of the surrounding landscape are maintained by the low hydraulic conductivity of the catotelm, owing to the collapse of peat structure and resultant reduction in pore space from the incompletely decayed and looser surface vegetation of the acrotelm (Ingram, 1982, 1983; Clymo, 1991).

Ombrotrophic raised bogs are largely acidic, anoxic and nutrient-poor ecosystems, which form as a result of the progressive accumulation of dead plant material. These ecosystems are typically dominated by moss species of the genus *Sphagnum*, although other species from the families Ericaceae and Cyperaceae are also commonly found, albeit in lesser proportions. The high cation exchange capacity possessed by *Sphagnum* causes the active acidification of their surrounding environments, as a result of the extraction nutrients (e.g. potassium, magnesium) and the excretion of hydrogen ions (Barber, 1982). This process often takes place to the detriment of other plant species, facilitating the ecological dominance of the genus (van Breemen, 1995).

Sphagna possess no leaf stomata and cannot regulate water loss. Despite physiological adaptations to reduce or negate the effects of evaporation, Sphagna remain heavily reliant on a favourable climate (Barber, 1982; Rydin and Jeglum, 2006), re-emphasising the importance of a positive water balance for the formation and growth of raised bog systems.

Owing to their reliance on precipitation, ombrotrophic raised bogs are only able to form in regions that experience a positive water balance, with precipitation exceeding evaporation and runoff. Consequently, these systems are typically distributed in northwest Europe, the northern portions of the Atlantic and Pacific seaboards in North America, New Zealand, southern Patagonia and parts of southeast Asia. When combined with the historic advantage Europe has enjoyed, in terms of science and academia, the majority of early peat studies took place in northwest Europe.

3.3 Peatlands as archives of climatic change

Dau (1823, 1829) was arguably the first to observe the occurrence of discrete 'layers' in the peat sequences of Zealand, Denmark. Famously, and rather later, Blytt (1876, 1882) hypothesised that these shifts from dark, well-humified *Sphagnum* peat, often rich in *Calluna*, to lighter, comparatively unhumified *Sphagnum* peat were physical manifestations of a climatic change from cooler and drier to warmer and wetter conditions. Blytt termed these contrasting climatic characteristics the 'Boreal' and 'Atlantic', respectively. This scheme was extended by Sernander (1908) to include the 'Sub-Boreal', 'Sub-Atlantic' and 'Lateglacial'. A combined Blytt-Sernander scheme served to define regionally transgressive periods of climatic change within peat sequences, subsequently enabling the correlation of raised bog stratigraphies across northwest Europe (e.g. Granlund, 1932).

Granlund's (1932) study of Swedish raised bogs, and discussion of *Grenzhorizont* ('boundary horizons') encouraged Godwin (1946) to compile his own study of, what he termed, 'recurrence surfaces' – periods of apparent increases in surface wetness, succeeding drier periods, in which a rejuvenation of peat growth was enabled. Barber (1982) talks of 'the search for fixed points' when discussing this period of peat stratigraphic study, and describes how the theory of recurrence surfaces subsequently dominated the field for many years. Of course, much of the science involved in these theories has since been invalidated or developed. These early stratigraphic correlations have, with the advent of radiocarbon dating, been shown to be overly simplistic (Smith and Pilcher, 1973; Birks, 1975) and modern studies now employ a far more rigorous correlative procedure, both in terms of palaeoclimatic evidence and chronology (e.g. Hughes *et al*., 2000; Blaauw *et al*., 2007a, 2008; Charman *et al*., 2009).

The recurrence surface hypothesis, promoted by the early researchers, held that the majority of changes in peat stratigraphy were the result of autogenic forcing factors, namely the internal dynamics of the growth of bog vegetation, and cycling between hummocks and hollows. They suggested that only large-scale stratigraphical changes, between humified and unhumified peat, could be attributed to allogenic factors, specifically climatic change. Building on the work of Walker and Walker (1961), these ideas were formally falsified by Barber (1981), who demonstrated clear correspondence between stratigraphic changes in the peat record and the winter severity and summer wetness indices of Lamb (1977). This confirmed the significant role climatic change has played in peat formation, a conclusion that has been supported by numerous studies since (Barber and Charman, 2003). It has been suggested that misplaced emphasis on an autogenic peat development process, may have hindered the development of peat palaeoclimate records in the decades between the development of the early theories and Barber's (1981) rebuttal (Barber, 1994; Barber and Charman, 2003; Chambers *et al.*, 2012). Charman (2002) emphasises that autogenic development and change provide background 'noise' against which any palaeoclimatic signal must be evaluated. Quantification of this biologically-induced noise (Barber *et al.*, 1998; Woodland *et al.*, 1998), combined with an increased understanding of the role of climatic variables in peatland hydrology (Schoning *et al.*, 2005; Charman, 2007; Charman *et al.*, 2009; Booth, 2010), has strengthened our knowledge of the link between BSW and climate.

Recently, Swindles *et al.* (2012a) have suggested that ecohydrological feedbacks inherent in peatland development can induce water table homeostasis; a concern that has been suggested previously (Barber, 1981, 1993). This proposed disconnection from external climatic influences, it is argued, prevents reliable palaeoclimatic reconstructions being inferred from records of past water table change. The authors present a well-dated, multi-proxy palaeohydrological record, of approximately 500 cal yr in length, from a site in North Yorkshire. This record is then compared with 8 other sites from two previous studies (Blundell and Barber, 2005; Langdon and Barber, 2005) finding that 5 out of 8 sites did not demonstrate the same hydrological pattern as their own. It is then suggested that regional climatic differences or autogenic factors could be responsible. Furthermore, Swindles *et al*. (2012a) present results from a 'virtual bog' model (Morris *et al*., 2011), which suggest that climatic wet shifts actually result in an initial lowering of WTD, as the increase in peat accumulation associated with the bog causes a relative decrease in water table. Conversely, the model predicted that shifts in climate to drier conditions, would also result in homeostatic water table rise, although no mechanism is offered for this. Finally, the model results presented also find that the response of the bog to climate perturbations is non-linear (i.e. water table response to 40% shifts are not twice the size of those associated with 20% shifts).

It is not acknowledged by the authors, however, that palaeohydrological proxies have the potential to exhibit climatic complacency (e.g. Blundell and Barber, 2005; Hughes *et al*., 2006), or outperform one another in terms of climatic sensitivity under different palaeohydrological conditions (e.g. Charman *et al*., 1999; Booth and Jackson, 2003). In addition, multi-proxy studies have also found inconsistencies between proxy data from the same peat sequence (e.g. Charman *et al*., 1999; Chiverrell, 2001; Mauquoy *et al*., 2002b; Langdon and Barber, 2004; Blundell and Barber, 2005 Yeloff and Mauquoy, 2006). These issues are discussed further in Section 3.4, but it is important to note that until additional modelling, observational and palaeoecological evidence are presented which support the conclusions of Swindles *et al*. (2012a), Charman's (2002) description of autogenic 'noise' is likely to persist in the literature. With an appropriate sampling and analytical approach (see Section 3.5), however, numerous studies have shown that raised bogs can be used successfully to identify regional palaeoclimatic change. Indeed, recent peat-based palaeoclimate studies maintain that "*proxy records derived from ombrotrophic peatlands provide important insights into climate change over decadal to millennial timescales*" (Swindles *et al*., 2012b, p.94). Regardless, the initial findings presented by Swindles *et al*. (2012a) suggest that additional research into this issue would certainly be warranted. For that reason, it should be acknowledged their study should represent a summary of the starting rather than the end point of this line of research.

Attempts to understand the role played by climate in BSW variability have persisted since the early works of Aaby (1976), van Geel (1978) and Barber (1981), through to more recent studies, such as Hendon and Charman (2004), Schoning *et al*. (2005) and Charman (2007). The summer wetness and winter severity indices developed by Lamb (1977) showed a positive correlation with a plant macrofossil-based BSW reconstruction from northern England (Barber, 1981). Furthermore, using a generalised linear model, Barber *et al*. (1994) compared this macrofossil record with the climatic indices, finding that taxa responses matched variation in the indices, thus demonstrating that the resultant climate response model exhibited a very good fit.

It may seem intuitive that prolonged, increased rainfall should influence BSW. However, Barber *et al*. (2000) hypothesised that temperature represented a dominant driving force, in the face of varying precipitation patterns between two systematically and climatically contrasted sites, 300km apart and differing by 900m attitudinally. Hendon and Charman (2004) also suggested that summer temperature may be the main driving factor behind BSW changes at sites in the Anglo-Scottish border region, over the last two centuries. Furthermore, Barber and Langdon (2007) compared BSW records from the mid-Holocene with mean July temperature records as inferred from chironomid-based lake records, finding that shifts in BSW coincided with shifts in temperature.

Whilst some evidence suggests that temperature may be the dominant variable, logic states that precipitation must also contribute significantly. Charman *et al*. (2009) conclude that BSW should be considered a function of both temperature and precipitation, caused by a complex interaction affecting annual deficit (i.e. precipitation minus evapotranspiration). Consequently, the duration of the summer water deficit (i.e. where evapotranspiration is greater than precipitation) has been shown to play a critical role in bog hydrology (Charman, 2007). Whilst temperature variation is a plausible explanation of variability in BSW, Charman *et al*. (2009) suggest that this relationship varies with time, and the magnitude of temperature change in the mid- to late-Holocene is insufficient to explain the corresponding magnitude in BSW changes. For example, Charman *et al*. (2004) compared recent, high-resolution BSW records with instrumental meteorological records, finding stronger correlations with summer precipitation than summer temperature. In fact, there is a negative correlation between modern instrumentally observed summer temperatures and summer precipitation, with both driving moisture deficit in the same direction (Charman *et al*., 2009), highlighting the difficulty in separating the forcing mechanisms from one another, given that, despite the dominance of precipitation over temperature, they appear to be intrinsically linked in their influence on WTD (Charman, 2007; Charman *et al*., 2009).

The climate record archived in peat sequences, in various proxy forms, is essentially a representation of biological and chemical response to changing BSW conditions over time, and therefore a reflection of changing atmospheric moisture balance. Variations in summer precipitation, particularly in western areas of Great Britain and Ireland, are strongly governed by the strength of westerly airflow, bringing cool, moist air from the Atlantic. BSW records from northwestern Europe can therefore be interpreted as proxies for the strength and position of the westerlies, based on the results of past studies (Charman and Hendon, 2000; Magny, 2004; de Jong *et al*., 2006; Blundell *et al*., 2008; Daley *et al*., 2010).

The increased use of peat-based stable isotopic analysis may provide further insight into the key drivers of changes in BSW. Potentially, the technique could record shifts in the strength of westerly airflow, characterised by changes in temperature and/or precipitation (Daley *et al*., 2009, 2010), thus enabling comparison of isotopic records with other BSW proxy records from the same sequences. By testing the relationships between records of BSW and temperature/precipitation, it may be possible to gain greater understanding of the drivers of BSW change in ombrotrophic raised bogs.

However, the over-arching forcing mechanism of regional patterns of BSW variability remains contentious. Two primary and related mechanisms have been proposed: solar variability and oceanic forcing. Solar variability is an established explanation, with initial suggestions (e.g. van Geel *et al*., 1996) more recently supported by well-dated records (e.g. Mauquoy *et al*., 2002a, 2004; Blaauw *et al*., 2004b) and data compilations (Charman *et al*., 2006). Furthermore, other proxy data, such as ice cores, lake varves, and treeline and glacier fluctuations also demonstrate evidence for cyclicity linked to solar variability (Barber *et al*., 2004).

However, it has been suggested that major changes in temperature and precipitation in northwestern Europe are likely to be driven by changes in the North Atlantic (i.e. NAO, THC) (Barber *et al*., 1994; Barber, 2006). Studies have shown links between changing BSW and oceanic forcing (e.g. Blundell and Barber, 2005), as indicated by indices of IRD (Bond *et al*., 2001) and Iceland-Scotland Overflow Water (Bianchi and McCave, 1999). However, changes in oceanic forcing in the region are likely to be intrinsically linked with solar variability and therefore the precise relationship between the two, and their subsequent influence on regional BSW patterns remains complicated (e.g. Turney *et al*., 2005; Swindles *et al*. 2007a).

Scientific interest in the identification of cycles within the peat-based palaeoclimate record has grown in recent years, given its potential relationship with the forcing mechanisms associated with regional BSW variation. Cyclic change in peat bogs was first identified by Aaby (1976), who detected a cyclicity of 260 years at a Danish site. Similar conclusions have been reached since (e.g. Chambers and Blackford, 2001; Barber *et al*., 2004), with some studies linking peat-based records from Great Britain and Ireland to Holocene oceanic cycles of 550, 800 and 1100 years (Barber *et al*., 1994; Hughes *et al*., 2000; Langdon *et al*., 2003), as detected by Chapman and Shackleton (2000). Where raised bog records have are of sufficient length and have accumulated in an uninterrupted fashion, time-series analysis of the subsequent climatic proxy data can be used to identify dominant periodicities (Barber *et al*., 1994). Recent chronological improvements, often enabled by wiggle-match dating (WMD), have led the number of studies looking to examine for periodicity to increase (e.g. Blackford, 2000; Langdon and Barber, 2005; Turney *et al*., 2005; Snowball and Muscheler, 2007; Mauquoy *et al*., 2008, Swindles *et al*., 2012b).

Swindles *et al*. (2012b) suggest that periodicities found in peatland records, which are replicated across multiple sites, clearly represent forcing by allogenic, rather than autogenic factors. However, Charman *et al*. (2009) state that given the range of periodicities detected thus far, a simple explanation as to the principal driver of BSW, whether oceanic or solar, cannot yet be offered. Rather, it is suggested that a complex combination of both drivers, together with associated feedback mechanisms, was responsible for the changes witnessed in the northwest European peat record (Wanner *et al*., 2008; Charman *et al*., 2010).

3.3.1 Potential of peatlands for the analysis of mid-Holocene climate events

Climatic change is now one of the major social, political and environmental issues faced by modern society. The uncertainty and potential risk associated with climatic changes has meant that interest in abrupt climate change events has grown dramatically in recent years, throughout the global palaeoclimate community (e.g. Broecker, 2000, 2006; Clark *et* al., 2002; Alley *et al.*, 2003; Schneider, 2004; Bradley, 2008). In addition, these events (e.g. 4.2 kyr and 8.2 kyr events) are also being used to formally subdivide geological epochs (Walker *et al*., 2012) and, therefore, an understanding of their timing and nature must be developed in all regions of the globe, using the best palaeoclimatic tools available in each location.

Barber (2006) describes how the focus of peat-based palaeoclimatic research is also increasingly turning to detailed analyses of individual events and periods of particular interest. This is, in part, a by-product of the community's attempts to improve understanding of the forcing mechanisms of BSW, with examinations of abrupt climate changes witnessed on regional scales and hypothesised to be associated with solar and/or oceanic forcing mechanisms (e.g. van Geel *et al*., 1996; Chambers *et al*., 2007; Swindles *et al*., 2007a; Plunkett and Swindles, 2008).

The Holocene peat-based palaeoclimate archive from northwest Europe is punctuated by a series of well-documented climatic downturns, driven by changes in the North Atlantic (e.g. Blackford and Chambers, 1995; Hughes *et al*., 2000; Langdon *et al*., 2003; Barber *et al*., 2004; Blundell and Barber, 2005; Borgmark, 2005; Brown, 2006; Charman *et al*., 2006; de Jong *et al*., 2006). A number of studies have shown these deteriorations to be broadly coherent in BSW records from across Europe (van Geel *et al*., 1998, Hughes *et al*., 2000; Barber and Charman, 2003; Barber, 2006).

Many of these climatic deteriorations have been the subject of focused studies, in which researchers attempt to examine the timing, nature and magnitude of the events. Of particular prominence in the palaeoecological literature are the Little Ice Age (Mauquoy *et al*., 2002a; De Vleeschouwer *et al*., 2009), the 2800 yr BP event (van Geel *et al*., 1996; Plunkett, 2006; Chambers *et al*., 2007; Swindles *et al*., 2007a; Plunkett and Swindles, 2008), the 5.2/3 kyr event (Magny and Haas, 2004; Caseldine *et al*., 2005), and the 8.2 kyr event (Hughes *et al*., 2006; Daley *et al*., 2009). As Section 2.8 outlines, the 4.2 kyr event remains comparatively underrepresented in the literature, with an explicit examination of its timing and nature in northwest Europe currently absent, despite repeated suggestions of its existence (e.g. Hughes *et al*., 2000; Barber *et al*., 2003; Mauquoy *et al*., 2008).

So far, this section has presented a case for the use of peat records for the analysis of Holocene climatic change, and the same logic can be extended to the analysis of abrupt climatic change. The peat bogs of northwest Europe, particularly of western Great Britain and Ireland (see Chapter 4), are highly sensitive to the changes in precipitation and temperature caused by the atmospheric and oceanic circulation variations often associated with abrupt climatic events, such as the 4.2 kyr event (see Section 2.7). Methods of palaeohydrological reconstruction, discussed in Section 3.4, together with the rapid accumulation and excellent chronological potential (see Chapter 6) associated with peat bogs in northwest Europe, present the optimal opportunity to examine the timing and nature of these events in the region.

3.4 Palaeohydrological reconstructions

Chambers *et al*. (2012) and de Jong *et al.* (2010) provide excellent and broad critical reviews of developments in proxy-climate reconstructions from peat records. However, the principal approach of peat-based palaeoclimate studies has been to reconstruct changes in bog hydrology over time, using one or more proxies of BSW, and it is these techniques upon which this review will focus. Of the three standard principal proxy-climate indicators from peats (cf. Hughes *et al*., 2006; de Jong *et al*., 2010), plant macrofossil analysis has been the most widely used, followed by humification analysis and, more recently, testate amoebae analysis. Blackford (2000) outlines two assumptions that underpin the use of plant macrofossils as proxies for BSW, which can also be extrapolated to apply to testate amoebae. Firstly, it is assumed that botanical and faunal assemblages respond to variation in the depth of the water table, which is itself responding to changes in climate. Secondly, it is assumed that the sub-fossil remains of vegetation and organisms preserved as part of the peat matrix are an accurate representation of vegetation cover or faunal composition at the time of peat deposition. If it is assumed that a species' ecological preference has not changed over time, it becomes possible to infer past changes in BSW from down-core assemblage variation. Blackford (2000) adds that humification analysis is based on the assumption that decomposition increases with dry conditions, resulting in darker, more humified peats than in peat that accumulates whilst the bog is wetter. In addition, peat-based records of temperature and/or precipitation change, potentially independent of changing BSW, have been sought, based on the isotopic analysis of hydrogen (^{2}H; also deuterium, D), oxygen (^{18}0) and carbon (^{13}C) in bulk peat and selected botanical components. The latter technique's use in these studies has been implemented with varying success, and its efficacy is still debated.

Many studies have advocated the need for a multi-proxy approach to palaeoclimatic reconstruction, and this study will incorporate all the above proxies to produce a comprehensive palaeohydrological reconstruction from each of the study sites (see Section 3.5). The following sections of this chapter serve to provide a brief outline of the background, application and limitations of each of the four proxy techniques employed in this study.

3.4.1 Plant macrofossil analysis

3.4.1.1 Background and principles

Plant macrofossil analysis could be considered the 'founding' peat-based palaeohydrological proxy, and remains one of the most commonly used. Since its inception, plant macrofossil analysis has been employed successfully to reconstruct palaeoclimates, particularly in northwestern Europe, in both single (e.g. Barber *et al*., 1994; 1998; 2000; 2003; 2004; Hughes *et al*., 2000) and multi-proxy contexts (e.g. Charman *et al*., 1999; Blundell and Barber, 2005; Langdon and Barber, 2005; Swindles *et al*., 2007a, b; Sillasoo *et al*., 2007; Blundell *et al*., 2008; Lamentowicz *et al*., 2008; Mauquoy *et al*., 2008). They have also been used to trace mire development pathways (Hughes *et al*., 2000; Hughes and Barber, 2003), to examine long-term vegetation development to inform conservation management (Chambers *et al*., 1997) and to examine the rate and nature of carbon sequestration in peat deposits (Heijmans *et al*., 2008). Although generally employed in European and North American peat deposits, plant macrofossil analysis also has a global application, with successful examples from southern South America (Mauquoy *et al*., 2004), Ile de la Posession (van der Putten *et al*., 2008) and South Georgia (van der Putten *et al*., 2009).

Plant macrofossils can be defined as the vegetative parts of plants, along with seeds and fruits, visible, but often not necessarily identifiable by eye with a median size range of 0.5-2mm (Birks, 2007). They are considered in contrast to pollen and spores from the same deposits, which are referred to as microfossils (Charman, 2002). Plant macrofossils from ombrotrophic raised bogs typically include *Sphagnum* mosses, ericaceous species such as *Erica tetralix* and *Calluna vulgaris* and sedges from the family Cyperaceae, including *Eriophorum* and *Rhynchospora* species. The anaerobic conditions present in ombrotrophic raised bogs allow for good preservation of plant material once it enters the saturated catotelm. As a result, plant material is often identifiable to species level, despite the loss of mass that occurs during peat formation (Mauquoy and van Geel, 2006).

Plant macrofossils operate as a climate proxy under the premise that BSW is the main controlling factor on botanical composition, responding along a well-documented bog surface microtopographical hummock-hollow gradient (e.g. Ratcliffe and Walker, 1958; van Geel, 1978; Andrus *et al*., 1983; van der Molen, 1988; Malmer, 1986; McMullen *et al*., 2004). It is also assumed that the ecological niches expressed by modern examples of peat-based species have not changed over time. Based on this ecological uniformitarianism, inferences can be made as to the likely changes in moisture based on variations in fossil assemblages.

Variation in the composition of *Sphagnum* species along the hummock-hollow gradient of a raised bog system is a product of their variable tolerance of water (Andrus *et al*., 1983; Vitt and Slack, 1984; Daniels and Eddy, 1990). Species occupy differing niches, at specific heights above or below the water table, based on a number of variables including their cation-exchange capacity, desiccation tolerance and resistance, capacity for water retention, rate of desiccation and their ability to photosynthesise at varying levels of inundation (Clymo and Hayward, 1982; Andrus, 1986). It has, however, been demonstrated that hummock-forming species (e.g. *Sphagnum austinii*) are tolerant of a broader range of ecological conditions than their hollow-dwelling equivalents (e.g. *Sphagnum* section *Cuspidata*; Clymo, 1970; Andrus *et al*., 1983). Consequently, when inferring climatic variation from the plant macrofossil record, entire species assemblages must be considered, rather than relying on the occurrence of 'indicator species'. This is especially true in instances where hummock-forming species are dominant.

Based on a comprehensive plant macrofossil dataset from Bolton Fell Moss, northern England, Barber (1981) developed the phasic theory of bog growth, whereby peat growth involving one botanical community will continue until WTD exceeds its threshold. Beyond this, a different botanical assemblage, ecologically appropriate for the new water-table position, will emerge (McMullen *et al*., 2004).

After preparation and analysis, (see Section 5.3.1) the raw plant macrofossil data are typically subjected to statistical analyses, with the aim of producing a single palaeoclimatic curve. Statistical transformations commonly employed include weighted-average ordination (e.g. Dupont, 1986) and detrended correspondence analysis (DCA) (e.g. Barber *et al*., 1994). However, Barber *et al*. (2003) have since observed that responses of botanical communities to changes in BSW are not linear, unlike those exhibited in other subfossil communities, such as testate amoebae (Charman *et al.*, 1999). Rather, they involve threshold changes, particularly when species appear and disappear from the peat record. As a result, the curves produced by these statistical transformations cannot be considered strictly quantitative reconstructions, in as much as they are not capable of reproducing, say, WTD. They do, however, as semi-quantitative plots, replicate the relative magnitude of past changes.

These threshold changes, coupled with the relatively low species diversity found in these environments, and with many species possessing relatively large hydrological/ecological niches, have hindered the development of quantitative transfer functions to reconstruct quantitative changes in BSW, as is commonly practised in the field of testate amoebae analysis (Daley and Barber, 2012). However, recent attempts have been made to extract transfer function based WTD reconstructions, comparable with those from testate amoebae, from plant macrofossil data (Väliranta *et al*., 2007, 2012). Again, these statistical methods will be reviewed at greater length, and their relevance to this study discussed further, later in Section 5.4.1.

3.4.1.2 Limitations

There are, however, a number of limitations to the plant macrofossil analytical technique, which must be born in mind when considering results. The first limitation is the differential decay rates of various peat-forming plant species. In this instance, the taphonomy of peat plant material is such that a fossil sample is unrepresentative of the assemblage, from which it originally derived, thus reducing the accuracy of any palaeoclimatic inferences based upon it. It has been found, for example, that *Cladonia*, a lichen often found on drier bog surfaces is only very rarely preserved in the fossil record. Conversely, the resistant nature of *Eriophorum* spp. root and leaf epidermis material often results in it being comparatively overrepresented in the fossil record. Indeed, root material generally, as it is frequently submerged in waterlogged conditions by its very nature, is often relatively well preserved in comparison with other plant components (Brown, 2006). A number of studies have observed inter-species differential decay rates, not least in the prominent peat-forming genus *Sphagnum*. Microtopographically, drier bog hummocks provide the most favourable conditions for decay. Within such features, Johnson *et al.* (1990) found that *Sphagnum balticum* is the least resistant to decay, with *Sphagnum rubellum* following, and *Sphagnum fuscum* being the most resistant; the latter of which was confirmed by Johnson and Damman (1991). However, Johnson and Damman also found *Sphagnum cuspidatum* to decay relatively quickly in wet hollow features. Furthermore, Clymo (1965) suggested that *Sphagnum acutifolium* and *Sphagnum cuspidatum* decayed at twice the rate comparative to *Sphagnum papillosum*. The issue of differential decay rates also has ramifications for the humification technique, which are discussed in Section 3.4.2.2.

The assumption of ecological uniformitarianism, necessary to employ the plant macrofossil record as a proxy for climatic change, can also present problems for the technique. A particularly well-documented example is the decline and subsequent local extinction of *S. austinii* (Hughes *et al.*, 2007). The dramatic decline has been observed in peat stratigraphies across Great Britain and Ireland (Godwin and Conway, 1939; Green, 1968; Dickson, 1973; Barber, 1981; Smith, 1985; Wimble, 1986; van Geel and Middeldorp, 1988; Stoneman, 1993; Stoneman *et al.*, 1993; Mauquoy and Barber, 1999b; Langdon and Barber, 2005; Hughes *et al.*, 2007). Factors suggested to have contributed to this decline include: climatic change (Godwin and Conway, 1939; Barber, 1981; Mauquoy and Barber, 1999b), drainage, burning and grazing (Pearsall, 1956; Piggot and Piggot, 1963; Hughes *et al.*, 2007), and airborne eutrophication (van Geel and Middeldorp, 1988).

However, two recent studies highlight the continued uncertainty regarding the species' demise. Mauquoy and Barber (1999a) found the decline in *S. austinii* to be associated with wet shifts in four out of six sites in a study across northern England and the Border region, suggesting a climatic driver. At the remaining two sites, Walton Moss and Bolton Fell Moss, however, the decline could be attributed to interspecific competition between different *Sphagnum* species. Such conclusions have been supported by evidence from four sites in northern Scotland, where Langdon and Barber (2005) found coincident increases in BSW, interspecific competition and loss of genetic diversity in *S. austinii*. Hughes *et al.* (2007), however, found evidence in Wales to suggest that the decline in *S. austinii* was most likely caused by increased agricultural pressures and land use intensity, as opposed to climate. They argued that changes in bog surface hydrology could not adequately explain the disappearance, as the species would have survived numerous other palaeohydrological variations in the early- to mid-Holocene, repeatedly re-establishing itself in the botanical assemblage.

Slater and Slater (1978) found that this species, where still present on the surface of northwest European raised bogs, was distributed almost exclusively in high hummock environments, unless in highly oceanic sites. This contrasts with findings from the mid- to late-Holocene in Great Britain and Ireland, when the species was found to inhabit far wetter environments (Stoneman *et al.*, 1993).

In a subsequent review of the species' past and present ecology, Stoneman *et al*. (1993) concluded that as the ecological/hydrological niche occupied by the species appears to have varied over time. This precluded the ecological uniformitarian principle from being easily applicable in this instance. As a result, caution should be taken when making climatic inferences based on a sequence where this species dominates. Understandably, this leads to significant problems for the interpretation of its presence within the plant macrofossil record (Barber and Charman, 2003; Barber, 2006).

Lastly, due to the aforementioned wide water-table tolerance of many species found in ombrotrophic raised bogs, plant macrofossil records possess the potential for climatic insensitivity, with changes of lower magnitude or shorter duration not being clearly displayed in some records (e.g. Hughes *et al*., 2006). Barber *et al*. (1994) concluded that plant macrofossil records were at their most sensitive when biodiversity was relatively high, particularly with regard to *Sphagnum* mosses. Subsequent interaction and competition between these species, as BSW fluctuated, would result in far more climatic sensitivity than in a sequence, which was dominated by a single species. In a multi-proxy study, Hughes *et al*. (2006) found the plant macrofossil record to express significantly less climatic variability than other proxies on a number of occasions. These periods occurred during phases of *Sphagnum fuscum* domination and it was concluded that climatic sensitivity had been reduced by this species' wide hydrological tolerance (Blundell and Barber, 2005).

Following a study at Tore Hill Moss, Scotland, Blundell and Barber (2005) also found evidence to suggest that dominant periods of *Sphagnum* section *Acutifolia* and *S. austinii* caused climatic complacency, limiting the palaeohydrological use of the record, as these species possess broad hydrological tolerances. It was suggested that this complacency could occasionally cause the plant macrofossil record to lag behind other palaeohydrological proxy records in terms of its response to climatic changes. Furthermore, it was suggested that by grouping *S.* s. *Acutifolia* species, due to the difficulty associated with identifying them to species level at a microscopic scale, further climatic insensitivity was introduced to the record. The section generally inhabits drier environments, but ecological differences do exist. For example, Daniels and Eddy (1990) found *Sphagnum subnitens* to often be present in slightly wetter environments, whilst *Sphagnum capillifolium* was almost exclusively present in high hummock features.

Equally, a plant macrofossil record dominated by indeterminate monocotyledon remains has the potential to produce limited information regarding past hydrological change (Chiverrell, 2001). This is a problem enhanced by the resistance to decay of many vascular plant roots, which are often difficult to identify to species level. Root material, by its nature, generally grows downwards into older sediments, often dominating levels at which the plant itself was not present on the bog surface. As a result, Mitchell *et al*. (2000) suggest that deep-rooted plants (e.g. *Eriophorum* spp.; certain mosses) should be considered as two separate function groups, and interpreted appropriately.

In spite of these issues, plant macrofossil analysis has been shown to have excellent application in northern European raised bogs, where species composition is sufficiently broad, but may prove to be less effective in other locations where single species dominance of a stratigraphy is more common, such as southern Patagonia (Pendall *et al*., 2001; Mauquoy *et al*., 2004) and eastern Canada (Hughes *et al*., 2006).

3.4.2 Peat humification analysis

3.4.2.1 Background and principles

The use of peat humification as a proxy for palaeoclimatic change is based on the premise that decomposition of peatland vegetation, and therefore degree of humification is primarily a function of BSW (Blackford and Chambers, 1993). In ombrotrophic raised bogs BSW is determined by climate (Barber, 1981) and, subsequently, humification analyses can be used to infer past changes therein. Chambers *et al*. (2011) provides an extensive review of the technique's theoretical and methodological details.

Generally speaking, decay of peatland vegetation is dependent on the amount of time spent in the biologically active acrotelm, before passing into the relatively inactive and saturated catotelm (Clymo, 1984; Belyea and Clymo, 2001). As a result, variation in the degree of decomposition over time is driven by changes in WTD (i.e. the acrotelm/catotelm boundary). During drier periods of reduced BSW, the water table is at its lowest, thus maximising the potential for decomposition in the acrotelm before plant material enters the permanently saturated catotelm. Conversely, increased BSW and higher water tables reduce the duration of vegetation decomposition, thus producing poorly humified peats (Barber and Langdon, 2001).

As evaporation increases in the summer months, water tables often become more variable, allowing for increased bacterial activity and subsequent aerobic vegetational decomposition in the aerated acrotelm. During the winter months when evaporation generally decreases, water tables stabilise and generally maintain higher levels (Anderson, 1998). As a result, it is generally accepted that variation in humification is representative of changes in average summer effective precipitation (i.e. precipitation which is actually added to, and stored in, the bog) (Barber and Langdon, 2001). However, it has been suggested that in colder periods and/or regions, temperature can become the dominant factor affecting peat humification. Roos-Barraclough *et al*. (2004) compared humification data from an ombrotrophic bog in the Swiss Jura mountains, with a nearby lacustrine oxygen isotope record, from the Lateglacial/Younger Dryas time period, finding an excellent correlation. However, in mid-Holocene northwest Europe, conditions cold enough to promote this type of response are extremely unlikely to have persisted for any significant amount of time.

As peat decomposes, the proportion of dark brown, humic acids increases, and so extracting and quantifying these acids, through a peat sequence, can produce a record of humification. Humic acids are extracted from a sample using sodium hydroxide (NaOH) and it is assumed that the colour of these NaOH extracts is indicative of the degree of humification. The extracts are measured colorimetrically and the level of light transmission is recorded, with high transmission values generally representative of low humification and, therefore, a relatively wet bog surface.

Conversely, low transmission values are indicative of higher humification levels and, therefore, a relatively dry bog surface (Blackford and Chambers, 1993). As the exact mathematical relationship between humic acid levels and BSW is not clear, humification analyses remain only semi-quantitative, and serve to provide data on relative, rather than absolute, water table changes (Yeloff and Mauquoy, 2006).

Initially, humification was assessed in the field using von Post's (1922, 1924) scaling method. Humification analysis by colormetric measurements was first introduced by Melin and Odén (1916), before being developed by Springer (1938) and Souci (1938). However, the method practised today is derived from the work of Danish researchers of the 1960s and 1970s, of which Aaby and Tauber (1975) is probably the best-known example. Modern protocols are generally considered to be adaptations of the refined methodology published by Blackford (1990) and Blackford and Chambers (1993). Indeed, this study follows an adapted methodology of Blackford and Chambers (1993). Blackford and Chambers (1993) review a number of other methods for determining peat humification, but the colorimetric technique prevails due to its ability to provide robust and replicable data in the most quantitative format possible. A more detailed history of the technique can be found in other publications (e.g. Yeloff and Mauquoy, 2006).

3.4.2.2 Limitations

Whilst fairly ubiquitous in peat-based palaeoclimate studies since the 1990s, humification analysis is not without its criticism and apparent limitations. The primary criticism is that results can be affected by differential decay rates in the various peat-forming plant species (see Section 3.4.1.2), with variations in humification level potentially reflecting merely a change in the botanical assemblage, rather than any direct climatic signal. A study by Yeloff and Mauquoy (2006) combined plant macrofossil and humification data in an examination of the issue and findings suggested that even subtle changes in species composition could have significant effects on the humification profile of a peat sequence. This potentially limits the inferences that can be made, with regard to past climatic variations, from a sequence in which the plant assemblage has varied significantly over the course of the profile. There are however, a number of studies in which variation in botanical composition does not appear to have had significant effects on the humification record (e.g. Chambers *et al.*, 1997; Blundell, 2002). In addition, Hughes *et al.* (2012) present a novel technique whereby *k*-values of fresh plant material (cf. Overbeck, 1947; i.e. light transmission values of alkali extracts from modern samples) can be used to correct the impact of botanical variation on transmission values over time.

There are also concerns with regard to the effects of secondary decomposition. This can occur when water tables are lowered during relatively dry periods, subsequently resulting in older peats being subjected to a second instance of decomposition (Borgmark and Schoning, 2006). The extent to which this occurs is unclear, often inconsistent and, as a result, difficult to correct for in the final analyses. However, many palaeoecologists have attempted to statistically remove obvious long-term decay signal by detrending the data where appropriate (e.g. Blackford and Chambers, 1995; Langdon *et al.*, 2003; Blundell and Barber, 2005; Borgmark and Wastegård, 2008).

Humification has the benefit of being possible in blanket peats (e.g. Blackford and Chambers, 1993), where other palaeoclimatic proxies, such as testate amoebae and plant macrofossils, are frequently not possible due to poor preservation (Yeloff and Mauquoy, 2006). The caveat of this is that with no plant macrofossil data available, it would be difficult to assess the influence of any botanical changes on the final humification record.

Despite these limitations, humification analysis remains a relatively quick and simple way to produce high-resolution, contiguous datasets, which, when incorporated as part of a comprehensive palaeohydrological multi-proxy study, can aid palaeoclimatic reconstruction. Using humification as part of a multi-proxy protocol also allows for the comparison of both the sensitivity and reliability of the various palaeoclimatic proxies involved. For example, after comparing peat humification and testate amoebae records, Charman *et al.* (1999) suggested that in wet conditions the former might be more sensitive to variations in WTD. Booth and Jackson (2003) found evidence to support this, extending the theory by suggesting that during drier periods, testate amoebae served as the more sensitive proxy. This merely highlights the importance of multi-proxy studies and inconsistencies between peat-based proxies from the same sequence have been found elsewhere (Charman *et al.*, 1999; Chiverrell, 2001; Mauquoy *et al.*, 2002b; Langdon and Barber, 2004; Blundell and Barber, 2005 Yeloff and Mauquoy, 2006). As Caseldine *et al.* (2000) observe, owing to the relative lack of understanding of the processes involved, compared to, say, testate amoebae or plant macrofossil records, incoherent humification data can be difficult to interpret, sometimes to the detriment of the final climatic reconstruction.

Caseldine *et al.* (2000) also highlight a final limitation of the technique, which occurs during the humic acid extraction itself. They found that when using NaOH as an extractant, organic decay products reflecting both humic and fulvic acids in the sample are produced, as well as enabling further breakdown of the peat during the extraction process. The authors conclude that as long as the extraction procedure is conducted rigorously and uniformly, it may still provide a climatic proxy. As a result of these findings, and possibly the other limitations discussed thus far, Caseldine *et al.* (2000) warn against the over-interpretation of small scale changes in the humification record, which could easily be the result of analytical noise. Instead, they suggest that only shifts of relatively large magnitude should be interpreted as demonstrating significant changes in BSW. A number of efforts were made to introduce new methodologies for the measurement of humification, using luminescence spectroscopy (Caseldine *et al.*, 2000) and near-infrared reflectance spectroscopy (McTiernan *et al.*, 1998), but these have not been adopted, despite promising initial results.

Studies are still being published that contain only humification data (e.g. Roos-Barraclough *et al.* 2004; Borgmark, 2005; Borgmark and Wastegård, 2008), but these are increasingly rare. More often than not, humification is included in multi-proxy, peat-based climate reconstruction studies alongside one (e.g. Barber *et al.*, 2000; Ellis and Tallis, 2000; McCulloch and Davies, 2001; Mauquoy *et al.*, 2002a, b) or more (e.g. Baker *et al.*, 1999; Charman *et al.*, 1999; Mauquoy and Barber, 1999a, b, 2002a, b; McGlone and Wilmshurst, 1999; Caseldine *et al.*, 2000; Charman *et al.*, 2001; Chiverrell, 2001; Booth and Jackson, 2003; Langdon *et al.*, 2003; Langdon and Barber, 2004; Mauquoy *et al.*

2004; Blundell and Barber, 2005; Vorren *et al*., 2007; Loisel *et al*., 2010; Amesbury *et al*., 2012; Daley and Barber, 2012) proxy techniques, often with a good level of coherence between the proxy records (e.g. Chiverrell, 2001; Langdon *et al*., 2003; Blundell and Barber, 2005; Hughes *et al*., 2006; Chambers *et al*., 2007).

3.4.3 Testate amoebae analysis

3.4.3.1 Background and principles

Testate amoebae are a diverse, polyphyletic group of unicellular, shell-forming protozoan organisms, which inhabit a wide range of terrestrial and aquatic habitats, including lakes, rivers, salt marshes, freshwater tidal marshes, wetlands, wet soils, organic litter and moss habitats (Charman, 2001; Mitchell *et al*., 2008; Charman *et al*., 2010; Ooms *et al*., 2011). Testate amoebae have also been referred to as thecamoebians (e.g. Medioli and Scott, 1983), testaceans (e.g. Tolonen *et al*., 1992), arcellaceans (e.g. Patterson *et al*., 1985) and rhizopods (e.g. Tolonen, 1986). However, some of these names are suggestive of taxonomic association, which is unrepresentative of all testate amoebae and therefore, following recommendations (e.g. Charman, 2001; Mitchell *et al*., 2008), this study will use the unambiguous term, testate amoebae, to refer to all shell-forming non-marine taxa of single celled protists.

Testate amoebae taxonomy is complicated, partially due to the polyphyletic nature of the group. A detailed account of testate amoebae taxonomy, biology and ecology can be found in Charman (1999), Mitchell *et al*. (2008) and Smith *et al*. (2008). Put simply, testate amoebae are phylogenetically divided into those with lobose and filose pseudopodia (Cavalier-Smith, 2004), which protrude from the test and possess dual functionality of mobility and feeding. Many of the shells, or ‘tests’, of a testate amoeba can be further subdivided into two groups. Xenosomic tests are formed by the agglutination of material from the surrounding environment, including siliceous or calcareous material, such as silica grains, other tests, pollen grains and diatom frustules, held together by secreted cement. Conversely, idiosomic tests are produced entirely by materials secreted by the parent amoeba during reproduction, frequently constructed from uniformly shaped plates. There are also examples of intermediary tests, which possess both idiosomic and xenosomic characteristics (e.g. of the genus *Nebela*). Finally, some tests cannot be placed in either group, and are without plates or other particles but rather formed entirely of smooth, proteinaceous material secreted by the amoeba itself (Charman, 2001). Tests are generally 20 – 200 μm (Mitchell *et al*., 2008), although can occasionally be larger or smaller (Charman *et al*., 2000). The broad size range and high morphological diversity associated with these tests means that they are frequently identifiable to species level. In addition, the decay-resistant nature of their composition means they are often well preserved in peat and lake sediment sequences (Warner, 1990). The group possesses a rapid reproduction rate of between 10 and 27 generations per year (Charman, 2001) and a typically high species diversity of c. 15 species per 4 cm^3 sample (Woodland *et al*., 1998), enabling sensitivity and rapid response to environmental changes. Consequently, they are ideal subjects for palaeoenvironmental proxy-based investigations.

The value of testate amoebae for ecological and climatic applications has been recognised in a range of archives, including: lakes (Beyens and Meisterfeld, 2002), fens (Payne, 2011), salt marshes (Charman *et al*., 1998; Gehrels *et al*., 2001; Charman *et al*., 2010) and freshwater tidal marshes (Ooms *et al*., 2011). However, whilst present in most wet soils, testate amoebae have been shown to be particularly abundant in *Sphagnum*-dominated habitats (Booth, 2002), and are therefore especially ubiquitous in raised bog systems. The bryophytes and vascular plants in such environments possess films of water in which testate amoebae can thrive (Charman *et al*., 2001). In addition, the saturated, anoxic and low energy sedimentary environment experienced within these systems provides excellent conditions for test preservation, although there are issues of preferential preservation (Payne, 2007; Swindles and Roe, 2007). The well-preserved nature of typical raised bog peat enables testate amoebae to be isolated using a simple sieving procedure, with no requirement for any chemical reagents (Hendon and Charman, 1997; Charman *et al*., 2000; Booth *et al*., 2010), which has clear benefits both for the examiner and for test preservation.

It has been observed for some time that testate amoebae assemblages vary with BSW (Steinecke, 1913; Harnisch, 1925, 1927). Early work was dominated by qualitative assessments of these variations, such as the moisture classes presented by Jung (1936), which ranged from I (wettest) to VIII (driest). However, this knowledge and understanding of testate amoebae ecology, and its relationship to prevailing hydrological conditions, has improved considerably over the last four decades, and the first attempts to extract semi-quantitative hydrological data from changes in testate amoebae assemblages were presented by Schönborn (1962) and Meisterfeld (1977). In Great Britain and Ireland, the study of testate amoebae in peatlands was pioneered by Heal (1961, 1962, 1964) who investigated the distribution of species across the topographical gradient, from pool to hummock. Techniques that measured moisture content of surface mosses enabled the identification of species niches more precisely (Tolonen *et al*., 1992, 1994; Warner, 1987), and subsequent studies have also included WTD measurements (Charman and Warner, 1992, 1997; Tolonen *et al*., 1992).

Furthermore, techniques that enabled the multivariate classification of communities (Meisterfeld, 1978, 1979; Mitchell *et al*., 1999; Tolonen *et al*., 1994) and fully quantitative techniques, involving univariate and multivariate statistics (Charman and Warner, 1992, 1997; Charman, 1997; Bobrov *et al*., 1999; Mitchell *et al*., 1999; Booth, 2001; Lamentowicz and Mitchell, 2005; Mitchell *et al*., 2008) have afforded a more comprehensive understanding of testate amoebae ecological preferences. These studies have demonstrated that testate amoebae community structure responds, principally, to some measure of moisture, whether moisture content or WTD.

Alongside substrate moisture (Tolonen, 1986; Warner, 1987, 1990), a number of other controls on taxa distribution patterns have been investigated, including trophic status (Tolonen *et al*., 1992), pH (Charman and Warner, 1992; Mitchell *et al*., 1999; Booth, 2001) and other aspects of water chemistry (Woodland *et al*., 1998; Mitchell *et al*., 2000). Whilst pH, in particular, has been found to be a considerable contributory control on assemblage variation, it is generally accepted that these factors are second to hydrology, in terms of their influence, in a variety of geographical settings

(Tolonen *et al.*, 1992; Charman and Warner, 1992; Charman, 1997; Bobrov *et al.*, 1999; Booth, 2001).

In addition, the relationship between temperature and testate amoebae assemblage is still relatively poorly understood (Smith *et al.*, 2008) and it has been suggested that improved comprehension of the relationship would further increase the palaeoclimatic value of testate amoebae (Mitchell *et al.*, 2008). However, Mitchell *et al.* (2008) highlight the difficulty associated with extricating a temperature control from a complex set of variables, outlined above, which also vary along latitudinal and altitudinal gradients (Odum, 1971). Subsequently, the scientific focus remains on the relationship between testate amoebae and hydrology.

As surface wetness in ombrotrophic raised bogs is principally a function of prevailing climate (see Sections 3.3 and 3.4), records of past palaeohydrological change, based on the testate amoebae assemblages contained within such sequences, can subsequently be interpreted as proxies for past palaeoclimatic change. When this opportunity is combined with the favourable depositional and compositional characteristics of these peatlands, as described earlier, the exceptional potential afforded by testate amoebae analysis from such sequences becomes clear. This has been recognised by the scientific community, and renewed interest has stimulated further ecological and palaeoecological work in the discipline, with significant progress being made in the last two decades (e.g. Tolonen *et al.*, 1992, 1994; Charman, 1997, 1999, 2001; Charman and Warner, 1997; Bobrov *et al.*, 1999; Hendon *et al.*, 2001; Booth, 2002; Charman *et al.*, 2007; Mitchell *et al.*, 2008; Smith *et al.*, 2008; Charman *et al.*, 2010; Markel *et al.*, 2010; Ooms *et al.*, 2011; Sullivan and Booth, 2011; Charman *et al.*, 2012).

The dominance of hydrology on species composition, together with the availability of quantitative hydrological data, has facilitated the development of testate amoebae-based transfer functions for the reconstruction of palaeohydrological changes. Species response to mean annual WTD and percentage soil moisture is well documented (e.g. Tolonen *et al.*, 1992, 1994; Woodland *et al.*, 1998; Mitchell *et al.*, 1999; Bobrov *et al.*, 1999), but the former is broadly considered a more reliable variable on which to base transfer function reconstructions (Woodland *et al.*, 1998). Attempts have been made to model testate amoebae response to other environmental variables including percentage soil moisture, pH, conductivity and a number of climatic parameters (Tolonen *et al.*, 1992, 1994; Charman, 1997; Charman and Warner, 1997; Woodland *et al.*, 1998; Mitchell *et al.*, 1999; Wilmshurst *et al.*, 2003; Payne *et al.*, 2006; Charman *et al.*, 2007; Booth, 2008), and across a range of wetland environments, across the trophic gradient (Payne and Mitchell, 2007; Payne and Pates, 2009; Booth, 2010; Payne, 2010; Payne *et al.*, 2010). However, greatest success in testate amoebae-based transfer function development remains in the modelling of species response to, and subsequent reconstruction of, WTD in ombrotrophic raised bogs (e.g. Charman *et al.*, 2007). Warner and Charman (1994) were the first to employ the technique in Canada, but transfer functions have now been developed and validated for many peatland regions, including: New Zealand (Charman, 1997); Russia (Bobrov *et al.*, 1999); North America (Charman and Warner, 1997; Booth, 2002, 2008; Payne *et al.*, 2006); mainland Europe (Mitchell *et al.*, 1999; Lamentowicz and Mitchell, 2005; Payne *et al*., 2008); and Great Britain and Ireland (Woodland *et al*., 1998; Charman *et al*., 2007; Swindles *et al*., 2009).

Alongside transfer functions, ordination techniques originally used with plant macrofossils, such as DCA, have also been applied to testate amoebae records. Together, these techniques provide promising quantitative methods by which ecological gradients, such as WTD, can be reconstructed. The main advantage of this is the ability to express assemblage variation, as a proxy for environmental change, on a linear scale, thus enabling direct comparison with records from other sites (Charman *et al*., 1999). The finer details of the application of DCA and transfer functions to testate amoebae datasets can be found in Sections 5.4.3.1 and 5.4.3.2, respectively.

In addition, testate amoebae have also been used to assess the affects of tephra loading on peatlands (Dwyer and Mitchell, 1997; Payne and Blackford, 2005), and to monitor peatland management and regeneration (e.g. Davis and Wilkinson, 2004; McMullen *et al*., 2004; Vickery and Charman, 2004). Finally, Charman *et al*. (2012) present the first attempt to provide fully quantitative palaeoclimatic reconstructions from testate amoebae data, reconstructing summer precipitation over the last 1000 years in Ireland.

3.4.3.2 Limitations

Despite significant progress in the analysis of testate amoebae as palaeoenvironmental indicators, the technique is not without its limitations. Many of the taphonomic and ecological problems outlined here are shared with plant macrofossil analysis (see Section 3.4.1.2). However, their combination with the taxonomic complexity of the testate amoebae group, already been described earlier in this section, can make certain issues particularly acute in this discipline (Charman, 1999).

Many of the common testate amoebae can be readily identified with confidence, but Mitchell *et al*. (2008) insist on the urgent need for a taxonomic revision and synthesis of current data. A wealth of identification materials exist for these organisms, but much of it is relatively, or very, old (Penard, 1902; Cash and Wailes, 1915; Deflandre, 1928, 1929, 1936; Grospietsch, 1958, 1964; Harnisch, 1958; Chardez, 1969, 1991; Corbet, 1973; Ogden and Hedley, 1980; Ogden, 1983; Schönborn *et al*., 1983). However, Charman *et al*. (2000) provided a valuable guide, detailing and standardising nomenclature and size divisions associated with testate amoebae taxonomy and identification. Nevertheless, Mitchell *et al*. (2008) maintain that considerable further work is needed to ensure the consistent and correct identification of testate amoebae.

Unfortunately, identification problems extend beyond the lack of modern, detailed monographs. The degree to which intra-species variation in testate amoebae affects accurate identification has not yet been properly assessed (Mitchell *et al*., 2008). Testate amoebae morphospecies are principally defined by size, composition and morphology of the test. Many taxa display gradational morphological change, and so divisions between taxa can all too frequently be considered as arbitrary (Charman, 1999). This has led to a number of problems associated with the naming of new species, and their subsequent identification, as what are effectively subjective judgements, can be predictably varied. If intra-specific variation within a particular taxon is not properly understood, it can lead to the erroneous allocation of species status to individuals, which are simply atypical examples of their species, as a result of exceptional ecological conditions (Charman, 2001). It is likely that some examples, which have been assigned species status, are in fact morphotypic extremes of the same variable species (Bobrov and Mazei, 2004). It has been shown that a high degree or morphological variability can exist both within and between populations (Bobrov and Mazei, 2004). This can be caused by a range of biotic and abiotic factors (Chardez, 1989; Schönborn, 1992; Wanner, 1999; Wanner *et al.*, 1994; Wanner and Meisterfeld, 1994), as well by test enlargement over time (Schönborn, 1992). Bobrov and Mazei (2004) suggest that this variation is caused by phenotypic plasticity, as species adapt their morphological characteristics to their surrounding environment. In addition, new species are often assigned upon examination of a relatively small number of individuals (Charman, 2001), and it is not unknown for new species to be identified, only for them to never be found again! Mitchell *et al.* (2008) reassert, however, that this should not detract from the value of testate amoebae for palaeoenvironmental reconstruction, and that greater taxonomic understanding would enhance their effectiveness further.

Medioli and Scott (1983) argued in favour of taxonomically grouping testate amoebae, however Charman (2001) suggests that this may result in valuable environmental information, related to minor morphological variation, being lost. Indeed, Bobrov *et al.* (1999) have shown that 'splitting' certain taxonomic groups can affect the environmental indicator values associated with those taxa. Clearly further work is required, so that the situation can be resolved, maximising the environmental information gained from correct identifications; although Smith *et al.* (2008) advise that species 'grouping' may be more genetically valid than species 'splitting'.

Workers are, therefore, faced with a decision as to which taxonomic scheme to adopt. At this stage in testate amoebae analytical research, Charman (1999) emphasises the importance of consistency in taxonomic criteria when presenting four schemes of increasing taxonomic subdivision, varying either side of the 'original' scheme, following Corbet (1973) and Grospietsch (1958) with modifications by Warner (1987) and Charman and Warner (1992). These schemes extend from the 'conservative' to the 'optimistic', in terms of their likelihood to combine similar taxa. Mitchell *et al.* (2008) suggest that the adoption of a more conservative approach to testate amoebae identification may avoid many of the problems associated with intra-specific variation and problematic taxonomy.

Despite generally good levels of testate amoebae preservation, especially in peatlands, there are taphonomic issues. Whilst it may be logical to suggest that preservation is linked to test morphology, in as much as idiosomic tests are more likely to be preserved than xenosomic tests (Medioli and Scott, 1983), it has been shown that this is not necessarily the case (Charman, 2001). However, Charman (1999) suggests that xenosomes on certain tests may suffer disproportionately in terms of number and density in fossilised individuals. Differential preservation between species has been demonstrated in mineral soil environments (Lousier and Parkinson, 1981), and evidence suggests this may also occur in peatlands (Steinecke, 1927; Tolonen, 1968; Wilmshurst *et al.*, 2003), particularly after laboratory preparations involving harsh chemical treatments (Hendon and Charman, 1997). The genus *Euglypha*, for example, has been described as the most abundant and widespread of testate amoebae genera (Payne and Mitchell, 2007; Smith *et al.*, 2008), but it is frequently absent from fossil records (Payne, 2005; Payne *et al.*, 2006). Levels within peat sequences occasionally have such poor test preservation that assemblage quantification is impossible (Beyens and Chardez, 1987; Charman, 2001; Swindles *et al.*, 2007b).

Studies that are explicitly concerned with testate amoebae preservation and taphonomy are increasing, and acidity, desiccation and nutrient status are all found to be contributory factors (Payne, 2006; Swindles *et al.*, 2007b). However, Mitchell *et al.* (2008) highlight the continued and relative paucity in our knowledge of the taphonomic processes associated with peat formation, and subsequently the transformation of modern to fossil testate amoebae assemblages, and suggest that further work is required to improve understanding.

The number of testate amoebae that need to be counted to accurately represent an assemblage remains a point of contention. Until recently, counts of 150 tests were deemed adequate enough to represent the main faunal changes between assemblages (Warner, 1990; Woodland, 1996). Charman (1999) observes that although larger count sizes would often increase the number of rare taxa identified, they were seldom conducted, as the small benefit to environmental interpretations would be deemed negligible when the additional time commitments were considered. However, more recently, after a systematic investigation into the impacts of count totals on the number of taxa identified, quantitative inferences of environmental variables and the strength of WTD predictive power, Payne and Mitchell (2009) found that counts of 100 were sufficient to detail relevant ecological community changes in the majority of circumstances. It is suggested that counts as low as 50 can be accepted as statistically representative if preservation conditions are particularly poor (Swindles *et al.*, 2007b; Payne and Mitchell, 2009)

Another problem in testate amoebae analysis is the lack of good modern analogues for past assemblages. This is of most concern in the construction and application of transfer functions (see Section 5.4.3.2) and whilst, in some instances, it may be caused by taphonomic bias in severely degraded peat (Wilmshurst *et al.*, 2003), it is primarily the result of poor representation of certain taxa in modern calibration datasets. For example, fossil assemblages in Ireland are frequently dominated by *Hyalosphenia subflava* and *Difflugia pulex* (Caseldine and Gearey, 2005), for which there are few or no modern analogue assemblages. It is not yet clear whether this absence of modern analogues is genuine or whether it is merely an artefact of incomplete sampling strategies. Mitchell *et al.* (2008) propose that such communities could inhabit degraded niches, which are intentionally or unintentionally avoided during surface sampling. For example, *Hyalosphenia subflava* is found to be associated with forested (Charman and Warner, 1992) and drained (Tolonen *et al.*, 1994) peatlands. *D. pulex*, however, are not known to dominate in any specific modern niche. Mitchell *et al.* (2008) suggest that a conscious effort should be made to sample unusual habitats, in an attempt to find a modern analogue for the species. In the first transfer function for Great Britain and Ireland, both contentious species were poorly represented. The current transfer function for the region, however, includes both species, with *H. subflava* being indicative of very dry conditions, and *D. pulex* of moderately dry to intermediate levels of BSW. Quite how accurate the representations of the species are in this training set are, however, debatable, potentially leading to problematic reconstructions where these species occur in abundance.

There are also issues of microhabitat specific and/or regionally diverse responses to ecological variation. Testate amoebae are assumed to be cosmopolitan; however, if this is not the case, or if they respond to temperature in addition to moisture and pH, then it is possible for the ecological preferences of a particular morphological species to vary geographically (Mitchell *et al.*, 2008). Further research is required to explore these possibilities. Similarly, the varied distribution of testate amoebae along *Sphagnum* stems is well documented, suggesting the organisms are sensitive to vertical micro-environmental gradients such as light and moisture availability (Chacharonis, 1954, 1956; Bonnet, 1958; Heal, 1962, 1964; Schönborn, 1963; Meisterfeld, 1977; Booth, 2002; Mitchell and Gilbert, 2004). Complex spatial patterns of community structure have been observed, even within mono-specific and relatively flat *Sphagnum* lawn environments, hypothesised to be the result of small variations in water chemistry and moisture conditions (Mitchell *et al.*, 2000).

Lastly, it has been shown that testate amoebae assemblages exhibit seasonal variation, in terms of relative abundance and community composition, most likely owing to hydrological changes through the year (Heal, 1964; Quinn, 2003; Warner *et al.*, 2007). This complicates the generation of representative testate amoebae training sets for transfer function development as, generally, a single surface sample is taken and analysed for testate amoebae assemblage and coupled with either a single point measure of environmental variables (WTD, pH, conductivity etc.), taken at the time of sampling or, in some instances, an annually averaged WTD (see Section 5.4.3.2). It has been proposed, however, that the sensitivity of testate amoebae to short-term environmental variability, such as this, could have positive implications for the amount of environmental information we are able to glean from studying the organisms (Sullivan and Booth, 2007). Nevertheless, Warner *et al.* (2007) suggest that a greater understanding of the organisms' relationship with microenvironmental variables is needed to fully assess the function of seasonality in testate amoebae.

Whilst the problems associated with testate amoebae analysis, as discussed here, present a series of challenges that future research must address, they by no means reduce the potential of the technique for reconstructing past environmental change. Before testate amoebae analysis became widely employed, Blackford (2000) outlined a major limitation of existing peat-based palaeoclimatic research. It was suggested that climatic reconstruction should ideally provide quantitative climatic data such as temperature or precipitation, and that the research at that time often resulted in rather vague and climatically imprecise descriptions of 'wetter and cooler' conditions. Testate amoebae analysis, therefore, presents an excellent opportunity to provide quantitative estimates of past climatic change, in the form of WTD reconstructions, something which has until recently been unavailable. As a result, testate amoebae are now routinely used to reconstruct past changes in WTD (e.g. Charman *et al.*, 2004; Chiverrell, 2001; Hendon and Charman, 2004; Hendon *et al.*, 2001; Langdon *et al.*, 2003; Mauquoy and Barber, 1999, 2002; McGlone and Wilmshurst, 1999; Wilmshurst *et al.*, 2002; Schoning *et al.*, 2005; Gearey and Caseldine, 2006; Booth, 2002, 2008). Frequently, these data are generated within multi-proxy studies incorporating plant macrofossil and/or humification analyses (e.g. Charman *et al.*, 1999, 2001, 2004; McGlone and Wilmshurst, 1999; Lavoie and Richard, 2000; Mitchell *et al.*, 2000; Mauquoy and Barber, 2002;

Wilmshurst *et al*., 2002; Booth and Jackson, 2003; Booth *et al*., 2004; McMullen *et al*., 2004; Blundell and Barber, 2005; Swindles *et al*., 2007a, b, 2010a; Lamentowicz *et al*., 2008), or stable isotopic analysis (Loisel *et al*., 2010; Markel *et al*., 2010). Blundell and Barber (2005) found that in some circumstances, testate amoebae reconstructions are more consistent than the other BSW reconstruction techniques, which can exhibit complacency (see Section 3.4.1), although they still advocate the use of a multi-proxy approach for the assessment of past hydrological change (see Section 3.5). It is claimed that the cosmopolitan nature of testate amoebae prevents them from being as affected by biogeographical distribution patterns as the plant macrofossil record (Mitchell *et al*., 2000). However, a possible effect of *Sphagnum* species on testate amoebae communities, due to the marginally different microenvironment provided by each species' leaf size and plant structure (Bobrov *et al*., 1999), is an area requiring more research, possibly through statistical inter-comparison of proxy records.

Comparison of transfer function reconstructions with instrumental climatic data from the recent past has confidently shown testate amoebae to be reliable even in reconstructing relatively low amplitude water table changes (Charman *et al*., 2004). This degree of sensitivity must be enabled, at least in part, by the rapid response time of testate amoebae to environmental change (Buttler *et al*., 1996), likely to be facilitated by rapid reproduction rates exhibited by the organisms (McCarthy *et al*., 1995). These characteristics are encouraging to those wishing to reconstruct periods of change, which are abrupt and/or subtle in nature. For example, McCarthy *et al*. (1995) observed a short-lived Lateglacial climatic event in the testate amoebae record of three lakes in eastern Canada. It is hypothesised that this is evidence of the testate amoebae community's response to the Younger Dryas Lateglacial stadial period.

To summarise, whilst the palaeoclimatic potential of testate amoebae in peatlands has been recognised since the early twentieth century, relatively little work has been done when compared to other established peat-based palaeoclimate proxies. Analyses are not without limitation and uncertainty, but significant progress has been made in the last two decades (Charman, 2001; Mitchell *et al*., 2008). The increasing predictive power of statistical reconstruction methods (Payne *et al*., 2011), together with its inclusion alongside new and novel palaeoclimatic techniques (e.g. Markel *et al*., 2010), has rendered testate amoebae analysis a vital part of any peat-based multi-proxy palaeoenvironmental study.

3.4.4 Stable isotopic analysis

3.4.4.1 Background and principles

The potential for reconstruction of past climatic change, based on the analysis of the stable isotopic composition of plant tissue has been the subject of scientific investigation since the discovery that the isotopic composition of meteoric water assimilated in to plant tissue varies with climate. Initially, investigations were undertaken by the tree-ring community and Libby *et al*. (1976) were among of the first to identify the potential for $\delta^{18}O$ analysis of annually-resolved tree-rings to act as an 'isotope thermometer'.

The development of a technique for the extraction of plant cellulose (Green, 1963) enabled a number of earlier studies to demonstrate empirical relationships between the isotopic ratios found in cellulose, and various climatic and environmental parameters (Libby and Pandolfi, 1974; Gray and Thompson, 1976; Libby *et al.*, 1976; Epstein and Yapp, 1977; Burk and Stuiver, 1981; Yapp and Epstein, 1982; DeNiro and Epstein, 1981; Edwards *et al.*, 1986; Sternberg *et al.*, 1986). Consequently, analysis expanded to employ a range of other materials, including: subfossil leaf cellulose from rodent middens (e.g. Terwilliger *et al.*, 2002), pollen (e.g. Loader and Hemming, 2004), grass cellulose and phytoliths (e.g. Webb and Longstaffe, 2006), and cellulose from peat bogs (e.g. Brenninkmeijer *et al.*, 1982; Aucour *et al.*, 1996; Hong *et al.*, 2000; Ménot–Combes *et al.*, 2002).

3.4.4.2 Application to peat bogs

As Brenninkmeijer *et al.* (1982) recognised, peat deposits present an ideal archive in which to combine existing palaeoecological techniques with novel isotope geochemical techniques. Furthermore, in the instance of ombrotrophic raised bogs, where moisture is obtained entirely from precipitation, the stable isotopic analysis of peat allows for the comparison of 'traditional' BSW proxies (i.e. humification, plant macrofossil and testate amoebae analyses) with a record potentially indicative of changes in temperature, precipitation and/or humidity. In addition, the rapid accumulation rates experienced in some bogs may enable direct comparison of peat isotope records with isotope signals from other high-resolution climate archives, such as tree-rings, speleothems, lake sediments and ice cores. Finally, the application of stable isotopic analysis to peat archives offers a unique opportunity to generate estimates of climatic change in a format that can be compared to these archives, as well as be used to assess global climate models (Daley *et al.*, 2009).

Many researchers have observed a strong correlation between plant tissue isotopic ratios and botanical composition in the peat record, particularly with variation in the relative abundance of vascular and non-vascular plants (Brenninkmeijer *et al.*, 1982; Brenninkmeijer, 1983; Dupont and Brenninkmeijer, 1984; Dupont, 1986; Dupont and Mook, 1987; van Geel and Middeldorp, 1988). Bulked sub-fossil peat samples often contain many different species of vascular and non-vascular plants, which possess different growth habits and isotopic discrimination mechanisms (Loader *et al.*, 2007), and can be subject to selective degradation of isotopically distinct compounds (Pancost *et al.*, 2003). The differential decay rates between peat-forming botanical taxa (Blackford and Chambers, 1993; Charman *et al.*, 1999) have already been discussed in Sections 3.4.1.2 and 3.4.2.2. Consequently, studies that have attempted palaeoclimatic and/or palaeoenvironmental reconstruction based on isotopic records obtained from cellulose extracted from bulked peat samples (e.g. Aucour *et al.*, 1996; Turney *et al.*, 1997; Hong *et al.*, 2000, 2001; Jedrysek and Skrzypek, 2005) have yielded ambiguous results. The magnitude of the isotopic variability seen in these studies caused by botanical variation is larger than the variability that could feasibly have been caused by changes in precipitation source water. For example, the difference in the isotopic ratio of ^{18}O between vascular and non-vascular plants is approximately 5‰ (Brenninkmeijer *et al.*, 1982; Ménot-Combes *et al.*, 2002), where modern annual seasonal variation in precipitation ^{18}O of northwestern Europe is just ~2-4‰ (Rozanski *et al.*, 1993). The practice of bulk sampling peat for isotopic analysis was a response to the large sample sizes required for mass spectrometry. Although van Geel and Middeldorp (1988) suggested that quantitative climatic signals could be extracted from the peat records, providing a single genus could be isolated, Brenninkmeijer (1983) found the prospect of manually isolating a single botanical component, such as *Sphagnum*, from the peat matrix to be an 'unrealistic effort'. However, significant advancements in the cellulose extraction and purification process (Loader *et al*., 1997), coupled with the development of continuous flow isotope ratio mass spectrometry (CI-IRMS), and more recently gas chromatography isotope ratio mass spectrometry (GC-IRMS), techniques, which require notably reduced sample sizes, have enabled the stable isotopic analysis of plant tissues from a single genus to become a realistic addition to palaeoclimatic studies in terms of the time and effort required (McCarroll and Loader, 2004; Filot *et al*., 2006; Loader *et al*., 2007; Woodley *et al*., 2011; Young *et al*., 2011).

The relative isotopic enrichment in vascular plants is the result of fractionation occurring during the uptake of water and carbon dioxide via complex pathways (Barbour, 2007). Opportunities for fractionation prior to cellulose synthesis are dependent on: evaporation from leaves (McCarroll and Loader, 2004); evaporation of soil water (Darling, 2004; McCarroll and Loader, 2004); and, soil water residence time (Dawson, 1993; Buhay and Edwards, 1995; Dawson and Pate, 1996). Conversely, non-vascular plants such as *Sphagnum* do not possess the physiological mechanisms to control water transport, water loss, or gas exchange, and consequently offer relative simplicity during water and carbon dioxide uptake. This makes them ideal subjects for isotopic analysis, as they directly record isotopic variation of the source water used for cellulose synthesis (Pendall *et al*., 2001; Ménot-Combes *et al*., 2002; Zanazzi and More, 2005; Loader *et al*., 2007; Daley *et al*., 2009, 2010). Furthermore, *Sphagnum* is by far the most abundant peat-forming species in the ombrotrophic bogs of northwestern Europe, and is present through extensive sections of many peat sequences, further enhancing its suitability for analysis.

In the last decade, a number of studies have demonstrated the potential of stable isotopes (i.e. δ^{18}O, δ^{13}C, δD), derived from isolated *Sphagnum*, as proxies for environmental and climatic change, supporting the reconstruction of variation in temperature, precipitation and/or humidity (Ménot-Combes *et al*., 2002; Sharma *et al*., 2004; Loader *et al*., 2007; Lamentowicz *et al*., 2008; Daley *et al*., 2009, 2010; Moschen *et al*., 2009; Loisel *et al*., 2009; Markel *et al*., 2010; Tillman *et al*., 2010; van der Knaap *et al*., 2011). The links between climate and the stable isotopic composition of precipitation, together with the subsequent incorporation of meteoric water into cellulose preserved in peat, are complicated. Nevertheless, a number of studies have examined the relationship between cellulose isotope ratios of modern *Sphagnum* and climate parameters, with promising results (Proctor *et al*., 1992; Aucour *et al*., 1996; Ménot and Burns, 2001; Ménot-Combes *et al*., 2002; Zanazzi and Mora, 2005; Skrzypek *et al*., 2007). An attempt to summarise the principles, limitations and potential applications of the technique are reviewed here. A more comprehensive account may be found in Daley (2007). As the uptake of water (containing stable isotopes δ^{18}O and δ^{2}H) and carbon dioxide (containing δ^{13}C), and subsequent isotopic fixation is different, the mechanisms and processes involved will be discussed separately.

3.4.4.3 Carbon dioxide uptake in *Sphagnum*

As Sphagna have no stomata, they are unable to physiologically regulate their uptake of CO_2 from the atmosphere. Instead, CO_2 diffuses through pores in the dead hyaline cells that surround the photosynthetic cells. The hyaline cells frequently fill to form significant water reservoirs, presenting a significant barrier to carbon assimilation, as diffusivity of CO_2 is much slower in water than in air. As a result, for *Sphagnum* mosses, the internal component of diffusional resistance is more complicated than that of vascular plants, and is a function of both diffusion through cell walls and mesophyll, and a variable component reflective of the internal water content (White *et al*., 1994; Ménot and Burns, 2001). It is this complication, which makes the isotopic signal preserved in their structure useful for palaeoenvironmental and palaeoclimatic reconstruction.

Due to the morphological differences between vascular plants and *Sphagnum* mosses, the latter can absorb and lose water far more rapidly than the former. Stable carbon isotope ($\delta^{13}C$) fractionation in *Sphagnum*, and subsequently the stable isotopic carbon ratios of the plant's cellulose ($\delta^{13}C$), is therefore heavily dependent on water availability. During periods of dry BSW, the hyaline cells are devoid of water and diffusion of CO_2 to the chloroplast is relatively high. Fixation of carbon is limited by metabolic activity, and $\delta^{13}C$ is determined by carbon isotope fractionation due to the photosynthetic enzyme. Subsequently, instances of desiccation result in increased discrimination against $^{13}CO_2$ during photosynthesis, as metabolic activity is the only limiting factor (Williams and Flanagan, 1996). Conversely, during periods of wet BSW, the hyaline cells tend to be inundated and diffusion of CO_2 to the chloroplast is reduced. Consequently, $\delta^{13}C$ is primarily determined by stable carbon isotope fractionation (Ménot and Burns, 2001). As a result, in *Sphagnum* the emptier the hyaline cell water reservoir surrounding the chloroplast is, the lower the $\delta^{13}C$ value, and *vice versa*. Stable carbon isotopic analysis of *Sphagnum* cellulose therefore possesses considerable potential for the reconstruction of BSW, which can, in turn, be related to climatic variability (Loader *et al*., 2007; Moschen *et al*., 2009). However, Markel *et al.* (2010) found that $\delta^{13}C$ records and palaeohydrological reconstructions from 12 peatlands in Alaska did not correlate well. Furthermore, Tillman *et al*. (2010) found their $\delta^{13}C$ record showed the greatest correlation with July temperatures, supporting the hypothesis that BSW may be closely related to changes in temperature (Barber and Charman, 2003; Hendon and Charman, 2004; Charman, 2007; Schoning *et al*., 2005; also see Sections 3.3 and 3.4). However, an encouraging number of other studies find good coherence between the $\delta^{13}C$ and testate amoebae records, concluding that the isotope can be employed as a proxy record for moisture changes at the bog surface (Lamentowicz *et al*., 2008; Loisel *et al*., 2009, 2010; van der Knaap *et al*., 2011).

3.4.4.4 Oxygen uptake in *Sphagnum*

Broadly speaking, the oxygen isotopic composition of plant cellulose ($\delta^{18}O$) is largely determined by the isotopic composition of the source water, the isotopic enrichment as a result of leaf water evapotranspiration and, finally, the biochemical fractionation occurring between source water and cellulose (Brenninkmeijer *et al*., 1982; Barbour, 2007). However, as Sphagna possess neither stomata nor vascular tissue, they employ a comparatively simple water use strategy. During periods of low water content, assimilation and the biosynthesis of cellulose is reduced as a result of an overall decrease in metabolic processes. However, it must also be noted that rates of photosynthesis are also inhibited as a result of the diffusional resistance imposed by excess water (Williams and Flanagan, 1996).

An overall enrichment factor between source water and *Sphagnum* cellulose during cellulose biosynthesis is 27 ± 3‰ for oxygen isotopes (Zanazzi and Mora, 2005). This is statistically identical to the enrichment factor of other terrestrial and aquatic plants operating in a range of environmental conditions and under various photosynthetic types (DeNiro and Epstein, 1979; DeNiro and Epstein, 1981; Sternberg *et al.*, 1986; Yakir and DeNiro, 1990). Because of this temperature insensitive constant fractionation factor, $\delta^{18}O$ analysis of *Sphagnum* cellulose therefore should possess a record of change in the oxygen isotopic composition of the plant's source water, thus providing extremely useful palaeoclimatic information.

Variation in the isotopic composition of source water can rarely be directly utilised as a 'palaeo-thermometer', owing to the changing nature of this relationship at different locations and within different air mass trajectories (Cole *et al.*, 1999; Araguás-Araguás *et al.*, 2000). Daley *et al.* (2010), for example, found temperature variability inferred from a $\delta^{18}O$ record derived from Walton Moss, northern England to be far greater than expected, when compared with other palaeoclimatic archives. The authors suggest interpreting these records with caution, but also observe a strong correlation between palaeoecological records and the oxygen isotopic record at the site suggesting a common climatic driver over centennial timescales. Daley *et al.* (2010) go on to describe the possibility that oxygen isotope ratios in this context could be influenced by variation in atmospheric circulation, as the isotopic composition of precipitation is known to reflect rainout history along given air mass trajectories (Cole *et al.*, 1999; Jouzel *et al.*, 1997, 2000). In addition, it has been shown that the isotopic composition of precipitation varies with atmospheric circulation changes caused by the North Atlantic Oscillation (White *et al.*, 1997) and oceanic circulation changes associated with the Great Salinity Anomaly (Serreze *et al.*, 1992; Curry *et al.*, 1998; Curry and Mauritzen, 2005).

There is a considerable body of evidence demonstrating variation of atmospheric circulation during the Holocene (Mayewski *et al.*, 2004). Changing pathways of westerly storm tracks, associated with fluctuations in the temperature gradient between high and low latitudes (Magny *et al.*, 2003, 2006), are potentially of considerable interest in peat-based palaeoclimate studies based in northwestern Europe. Variation in atmospheric circulation in Europe has also been observed, as documented in storm intensity, treeline variation, pollen stratigraphy and carbonate lake isotope records (Seppa and Birks, 2001; Hammarlund *et al.*, 2002; Veski *et al.*, 2004; de Jong *et al.*, 2006). Similar variation has also been documented in other regions of the globe, including North America (Edwards *et al.*, 1996; Dean *et al.*, 2002) and the southern mid-latitudes (van Geel *et al.*, 2000; Chambers *et al.*, 2007). This allowed Daley *et al.* (2010) to conclude that variation in oxygen isotope records obtained from *Sphagnum* in peat is likely to reflect either the cooling of prevailing air masses passing over a given site, or changes in the moisture source for these air masses. Consequently, such records possess considerable potential for the reconstruction of relative changes in atmospheric circulation over time. In addition, Daley *et al*. (2010) compared a $\delta^{18}O$ record from northern England with a composite BSW record. The two records showed good correlation, and since *Sphagnum* cellulose synthesis documents the isotopic signal varying in meteoric waters, a common climatic driver is suggested for variability of $\delta^{18}O$ of precipitation and BSW over centennial timescales.

3.4.4.5 Hydrogen uptake in *Sphagnum*

Of the three stable isotopes discussed here, understanding of the fractionation and assimilation of hydrogen in *Sphagnum* cellulose biosynthesis is the most limited. Unlike oxygen, hydrogen fixation is not the same for all plants. It varies, at least in part, according to the extent of biochemical fractionation taking place in different biosynthetic pathways (Estep and Hoering, 1980; Estep and Hoering, 1981; Yakir and DeNiro, 1990). It has, however, been found that the deuterium/hydrogen ratio contained within non-exchangeable, carbon-bound hydrogen in nitrated plant cellulose (δD) reflects the corresponding ratio in the plant's source water (Yapp and Epstein, 1982; White *et al*., 1994; Pendall *et al*., 2001). The deuterium/hydrogen ratio of meteoric water is strongly governed by surface temperature, but this relationship can be influenced by other factors including: changing storm tracks, variable moisture sources and precipitation seasonality (Dansgaard, 1964; Gat, 1980; Rozanski *et al*., 1993; Cole *et al*., 1999; Pendall *et al*., 2001). The mechanisms involved in δD in *Sphagnum* remain the least understood of the isotopes discussed here, but it is hoped that its continued use with help to elucidate its potential role in palaeoclimate science. For reasons discussed in detail in Section 5.3.4.3 this study did not collect δD data.

3.4.4.6 A multi-isotope, multi-proxy approach

It is clear that despite some uncertainty as to the interpretation of stable isotopic records from *Sphagnum*, the technique retains considerable potential as a palaeoclimatic proxy. However, in order to exploit the maximum interpretive power of plant tissue isotopic records, it has been advocated that they are best employed within multi-isotope studies (McCarroll and Loader, 2004). The emergence of novel measurement methodologies (Woodley *et al*., 2011; see Section 5.3.4), incorporating the use of a GC-IRMS machine at Swansea University, presents this study with an opportunity to compare results for different isotopes extracted from the same sample. Daley *et al* (2010) discuss the potential of stable isotopic analysis within peat-based palaeoclimatic studies, concluding that the technique should be employed alongside existing palaeoecological techniques, as part of a multi-proxy study, to obtain the most valuable results. Combining these two approaches, as is the intention in this study, can only serve to improve understanding as to the role stable isotopic analysis can play in the future of peat-based climate archives.

3.4.4.7 The importance of α-cellulose

The methodological advancements that have enabled single genus stable isotopic analysis in peat-based studies, as discussed above, have been coupled with a growing appreciation of the importance of isolating α-cellulose from plant tissue for subsequent analysis. This was a practice originally pioneered by the dendroclimatology community. Plant tissues contain a number of chemical components, which differ in their proportions from species to species. For example, wholewood contains lignins, cellulose, hemicelluloses, lipids, tannins and resin (McCarroll and Loader, 2004; Rinne *et al.*, 2005), whereas *Sphagnum*'s simpler chemical make-up contains only cellulose, hemicelluloses and galacturonic lipids (van Breemen, 1995).

The isolation of purified α-cellulose for analysis was widely adopted when it became apparent that the different chemical components of wholewood possessed differing isotopic ratios (Wilson and Grinstead, 1977). As a result, isotopic signals obtained from bulk plant tissue samples are an amalgamation of signals from each differing component. Initially, the dendroclimatology community employed α-cellulose due to the subsequent ability to link the isotope signal specifically to the growth period in the tree's annual cycle of wood formation (Helle and Schleser, 2004; McCarroll and Loader, 2004). The practice has been extended to other plant tissues, including *Sphagnum* found in peat records (e.g. Daley *et al.*, 2010). The process of extraction of α-cellulose is, however, extremely time consuming and requires significant amounts of specialist laboratory equipment and its necessity is debated (Borella *et al.*, 1998; Ménot and Burns, 2001; Cullen and Grierson, 2005; van der Knaap *et al.*, 2011), and successful palaeoclimatic reconstructions have been made using bulked *Sphagnum* material (Ménot and Burns, 2001; Loader *et al.*, 2007; Skrzypek *et al.*, 2007). However, the isolation of a single chemical compound reduces the risks associated with diagenetic variation of the relative proportions of chemical compounds over time (Suberkropp *et al.*, 1976; Spiker and Hatcher, 1987; Schleser *et al.*, 1999; Rinne *et al.*, 2005). Furthermore, extraction of the α-cellulose encourages homogeneity within the sample thereby improving estimates of mean isotopic value (Borella *et al.*, 2004; McCarroll and Loader, 2004). Consequently, many current studies continue to extract *Sphagnum* α-cellulose for analysis, with methodological advances and modifications (Daley, 2007; see Section 5.3.4). Inter-laboratory comparisons of cellulose preparation methods found α-cellulose methods preferable to those of holocellulose (Boettger *et al.*, 2007).

3.4.4.8 Single-component analysis

Previously, the selection of genus specific samples for isotope analysis was deemed unfeasible due to the large amounts of material required for processing. However, the methodological and technological advancements, summarised in this section, have enabled not only genus specific, but also component specific sampling. It has been shown that the differential decay rates of different plant components (e.g. leaves, stems, branches) can lead to systematic errors during isotopic analysis, owing to the isotopic offset between them (Loader *et al.*, 2007). As a result, a number of studies have investigated this offset in the *Sphagnum* genus, concluding that single-component sampling should be practiced where possible.

Tillman *et al.* (2010) also found less isotopic variability in replicate samples composed of branch material, than in those composed of stems. They suggest that the large number of branches required for each sample, owing to their relatively low weight, combined with the comparative ease at which their cellulose chains can be homogenised, decreases isotopic variability. However, it could be argued that if the homogenisation process were executed satisfactorily, then isotopic variability in samples containing *Sphagnum* stems would be comparatively low.

Tillman *et al.* (2010) found *Sphagnum* branch material to be less depleted in $\delta^{13}C$, but more enriched in $\delta^{18}O$ than stem material. Their results confirmed previous findings of a ~2‰ offset in carbon isotope ratios between the two components (Loader *et al.*, 2007). This trend is observable down-core and implies a common forcing mechanism, despite the offset. However, it remains preferable to solely sample either stems or branches. Where this is not possible, Loader *et al.* (2007) suggest that if components from a large number of plants are included and well homogenised, a broadly meaningful isotopic record could be produced.

Moschen *et al.* (2009) conducted a detailed inter- and intra-plant stable isotopic investigation, of the varied physical components of modern *Sphagnum* samples, confirming a significant offset between branches and stems, but failing to provide a firm conclusion as to the physiological or biomechanical cause. Loader *et al.* (2007) suggest a number of hypotheses, including: time of growth, allocation of photosynthates and secondary products, rate of photosynthesis, accessibility of chloroplast and susceptibility to drying.

Moschen *et al.* (2009) hypothesise that *Sphagnum* branches decompose relatively quickly in comparison to stems, causing differential rates of cellulose degradation, and subsequently favour the use of stem material in isotopic analysis. This reduced rate of decomposition also allows for more confident identification of *Sphagnum* stems. In addition, Loader *et al.* (2007) found an additional, if relatively small, offset between pendant and horizontal branches, further complicating the identification issues associated with the use of branch material in single-component sampling.

From a practical viewpoint, Tillman *et al.* (2010) found stems easier to pick out than the corresponding branch material. The weight ratio between the two components is such that a significantly smaller number of stems are required to reach a satisfactory sample weight. Unfortunately, stem material can often be scarce and the process can still be very time consuming. Regardless, the ease with which stem material can be decontaminated, as observed by Tillman *et al.* (2010), makes them preferable in terms of sampling efficiency. Moschen *et al.* (2009) conclude that stem material should be considered preferable to branches during sampling to both reduce the effect of any component-specific isotopic offset experienced, and minimise the effects of the time-intensive technique as far as possible.

3.4.4.9 Limitations

It is important to remember that the interpretation of the isotopic composition of *Sphagnum* cellulose is still in its infancy, when compared to other techniques employed in this study. As a result, there are a number of potential limitations that must be carefully considered during the interpretation phase of the analysis, before any climatic inferences are made. Tillman *et al.* (2010) highlight problems associated with the varying resolution of peat samples due to varying growth rates, species-specific variation, poorly understood biochemical pathways in plant tissues and signal preservation. These will be discussed at greater length in Chapter 8.

However, one of these potential limitations has methodological implications. A drawback of using stem material for isotopic analysis, as justified above, is that identification to *Sphagnum* species level is rarely possible (Moschen *et al*., 2009). *Sphagnum* cellulose from different species has been shown to exhibit isotopic signals of varying amplitudes, suggesting that this signal is species dependent (Ménot and Burns, 2001; Ménot-Combes *et al*., 2002). As species composition is likely to vary down-core, it is possible that isotopic records could be misinterpreted. This lends further credence to the suggestion that isotopic analysis of this sort should preferably be combined with other peat-based analyses as part of a multi-proxy study. Careful interpretation of the isotopic record in conjunction with the plant macrofossil record may reduce or remove the potential problems associated with inter-species variability. In addition, studies appear to suggest that effects of offsets between *Sphagnum* species are less than that of moisture changes (Rice, 2000; Loisel *et al*., 2009) and that, therefore, species do not require identification prior to analysis.

Loader *et al*. (2007) observed that *Sphagnum* macrofossils suitable for cellulose extraction (i.e. branch and stem material) contained in ancient peat samples became difficult to consistently identify with an acceptable degree of confidence when component dissociation had occurred. *Sphagnum* remains are often identified based on their leaves and where these are absent, identification is largely impossible. For this reason it was decided not to undergo species-specific sampling for these analyses.

Despite the limitations and uncertainties associated with peat-based stable isotopic analysis, as outlined in this section, significant advances have been made over the last decade, and understanding of the mechanisms and processes involved has improved greatly. The technique remains a potentially useful tool for palaeoclimatic reconstruction, and its benefit to the discipline can only increase with its continued use, especially if employed within multi-proxy studies (Daley *et al*., 2010).

3.5 Methodological rationale

Despite an overwhelming body of evidence for a global climatic event 4.2 kyr BP, together with suggestions of a potential manifestation in the peat climate archive of northwestern Europe, no peat-based palaeoclimate study has explicitly focused on the 4.2 kyr event in this region. Peat records encompassing the entirety of this period are scarce in Great Britain and Ireland, with many records focussing on the last c. 4500 cal yrs (e.g. Charman *et al*., 2006; Swindles *et al*., 2010). Where longer records exist, they often lack the sampling and chronological resolution necessary to examine the timing and nature of mid-Holocene climatic events, and are often based on single palaeoecological proxy reconstructions (e.g. plant macrofossils; Hughes *et al*., 2000; Barber *et al*., 2003; Mauquoy *et al*., 2008; see Section 2.8). However, the excellent potential of peatlands for palaeoclimatic reconstruction, discussed in this chapter, presents an optimal opportunity to examine the timing and nature of the 4.2 kyr event within the context of region heavily influenced by the atmospheric and oceanic circulation change of the North Atlantic, one of the principal mechanisms put forward as a possible cause of the event (see Section 2.7).

From the preceding sections of this chapter, it should be clear that a multi-proxy approach is generally considered preferable in modern peat-based palaeoclimate studies. Therefore, this study will adopt a high-resolution, multi-proxy palaeoecological methodology, including humification, plant macrofossil and testate amoebae analysis, focussed on the mid-Holocene. By incorporating three palaeoecological proxies for BSW/WTD, it will be possible to compare, contrast and cross-validate between records, before any conclusions are made, thus providing a more confident reconstruction of past climatic changes, potentially associated with the 4.2 kyr event.

Issues associated with these palaeoecological techniques, as outlined in this chapter, can in some situations, reduce the meaning and/or confidence that can be attached to certain records in each peat archive. As a result, by selecting sites in close geographical proximity, proxy records can be compared between sites to evaluate their reliability, reproducibility and climatic sensitivity (cf. Payne and Blackford, 2008). A more detailed discussion of the regional focus and site selection can be found in Chapter 4.

Previous studies, of lower analytical and chronological resolution, register a notable shift towards wetter and/or cooler conditions c. 4000 cal yr BP in various records from northwest Europe, including in Great Britain and Ireland (Hughes *et al*., 2000; Barber *et al*., 2003; Mauquoy *et al*., 2008). It could reasonably be expected, therefore, that the palaeoecological data in this study register a similar signal. However, the high-resolution, temporally-focused nature of the data generated here, when also combined with a high-quality chronology (see Chapter 6), will provide an opportunity to examine the timing and nature of the climatic changes during this period more comprehensively than any previous study.

To the author's knowledge, this study will also be the first to generate testate amoebae, humification and plant macrofossil data alongside multiple stable isotope analysis from an ombrotrophic raised bog. Previous studies have employed the stable isotopic analysis of peat material alongside multi-proxy palaeoecological records, but have restricted their examinations to single isotopes (e.g. $\delta^{13}C$, Lamentowicz *et al*., 2008, Loisel *et al*., 2010, Markel *et al*., 2010; $\delta^{18}O$, Daley *et al*., 2010). Only one peatland study has employed both $\delta^{13}C$ and $\delta^{18}O$ analysis alongside multi-proxy palaeoecological techniques (van der Knaap *et al*., 2011), but it could be argued that this Swiss sub-Alpine mire lacks the climatic sensitivity afforded to raised bog sites on the North Atlantic seaboard (see Chapter 4).

Many have hypothesised that $\delta^{13}C$ in *Sphagnum* is largely a factor of carbon isotope fractionation as influenced by BSW (Ménot and Burns, 2001; Loader *et al*., 2007; Moschen *et al*., 2009), and a number of subsequent studies have demonstrated a relationship between $\delta^{13}C$ and BSW records (e.g. Lamentowicz *et al*., 2008; Loisel *et al*., 2009, 2010; van der Knaap *et al*., 2011). Therefore the $\delta^{13}C$ record developed within this study should be expect to reflect changes in BSW/WTD as recorded in the other palaeoecological proxies, providing a further palaeohydrological record against which this data can be compared.

Furthermore, Daley *et al*. (2009) have demonstrated the ability the *Sphagnum* α-cellulose $\delta^{18}O$ record to characterise the nature and timing of abrupt climate change events. Their $\delta^{18}O$ record, developed from a site in the Labrador region of Canada, found strong evidence for the 8.2 kyr event, which is thought to have occurred following freshwater inundation of the North Atlantic, due to a catastrophic discharge of glacial meltwater from glacial Lake Agassiz (Barber *et al.*, 1999; Teller *et al.*, 2002; Clarke *et al.*, 2004; Daley *et al.*, 2011). Although not discussed by the authors, a notable excursion, centred around a period approximately 4000 years ago is also evident. Similarly, the $\delta^{18}O$ record produced by Daley *et al.* (2010) from Walton Moss, northern England, exhibits a number of large magnitude changes, characterised by isotopic minima, one of which occurs c. 4200 cal yr BP. This technique, therefore, has clear application to the aims and objectives of this thesis, in as much as it should be expected to record any major changes in atmospheric circulation during the mid-Holocene, hypothesised to be the most likely cause of the 4.2 kyr event (see Section 2.7).

Finally, as discussed earlier, the isotopic techniques described here are still in their scientific infancy. Employing them alongside more established palaeoecological proxies may, consequently, improve our understanding of the mechanisms and signals involved. For example, Markel *et al*. (2010) and Loisel *et al*. (2010) compared $\delta^{13}C$ records with testate amoebae inferred WTD reconstructions in an effort to formulate more informed conclusions as to the true nature of the $\delta^{13}C$ signal.

In conclusion, it is hoped that the multi-proxy, multi-isotope approach described here will provide the best opportunity to characterise the nature and timing of mid-Holocene climatic changes, including the 4.2 kyr event, as reflected in the peatlands of Great Britain and Ireland. This high-quality dataset will also provide an excellent opportunity to evaluate the use of the discussed techniques for the study of Holocene abrupt climatic events in peatland archives.

Chapter 4 - Sites

4.1 Introduction

This chapter describes the rationale behind study site selection, in relation to the aims and objectives of this research project, as outlined in Chapter 1. A detailed description of each site is also included, with the past palaeoecological and palaeoenvironmental research conducted at each site.

4.2 Site selection

A principal aim of this study is to assess whether a coherent 4.2 kyr event signal can be characterised in the Great Britain and Ireland region. In Section 2.8 of this study, four criteria were presented for records employed for the examination of potential abrupt climate change events. These include an ability to: i) represent a climate dominated signal; ii) be high-resolution; iii) have a high-quality chronology; and iv) produce a semi-quantitative reconstruction. In addition, another principal aim is to assess the potential contribution of peat-based stable isotopic analysis to our understanding of the nature and/or cause of the 4.2 kyr event. In light of this, site selection must be governed by an additional set of site selection criteria, put in place to ensure that the subsequent records are suitable.

4.2.1 Site selection criteria

Sites selected for inclusion in the study must meet the set of criteria, outlined and justified as follows:

i) *Ombrotrophic raised bogs*

As initially suggested by Barber (1981), peat-based palaeohydrological reconstructions should employ ombrotrophic raised bogs due to the direct coupling which exists between climate and changes in BSW, ensuring the production of a climate dominated signal.

ii) *Past palaeoenvironmental research*

The sites must have been subject to previous palaeoenvironmental research, thus ensuring the following:

a. *Good chronological potential*

Each site must have a history of robust radiocarbon dating, together with consistent, continuous accumulation at a rate of c. 10 – 20 years/cm.

b. *Presence of Hekla 4 tephra*

Not only does the presence of mid- to late-Holocene Icelandic tephra horizons improve the chronological potential of a sequence, the presence of the Hekla 4 horizon, specifically, will allow the rapid identification of the period of interest within the sequence (see Chapter 6).

c. *Sphagnum-rich*

Each site must possess a *Sphagnum*-rich record throughout the mid-Holocene, so as to enable the stable isotopic analysis of α-cellulose, as extracted from *Sphagnum* stems. Subsequently, previous site stratigraphies must demonstrate a dominance of *Sphagnum* in the bog's down-core species assemblages.

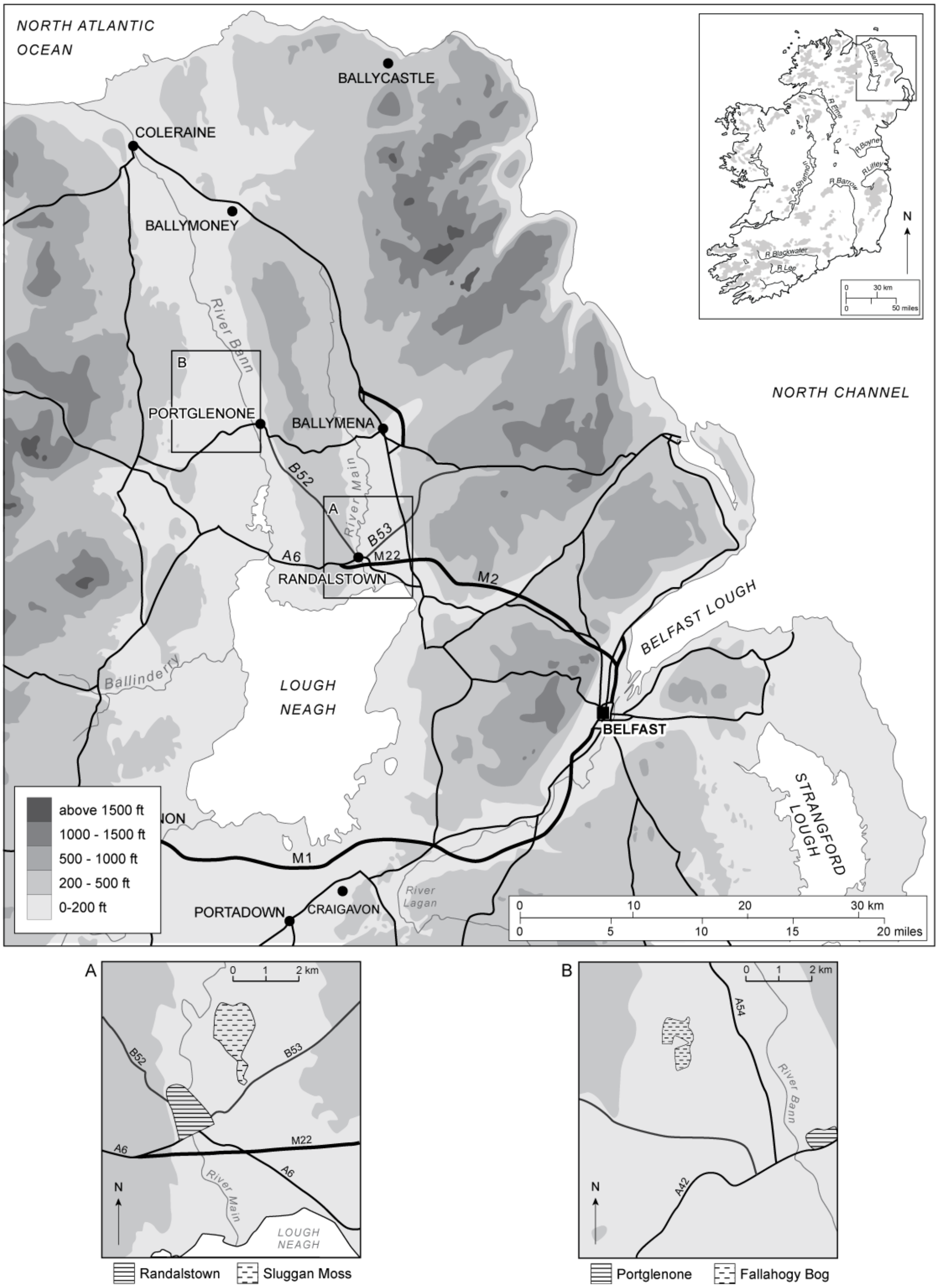
4.2.2 Ireland as a region of focus

When the hypothesised causal factors of the 4.2 kyr event are considered, the significance of the North Atlantic becomes clear (see Section 2.7.1). On the western fringe of Europe, Ireland is located on the eastern seaboard of the North Atlantic; a region absent of the influence of large continents or ice masses throughout the Holocene. Its maritime climate is, therefore, dominated by prevailing westerly airflows and as a result is particularly sensitive to latitudinal variation of the westerlies, influence of the NAO and THC, and other climatic systems of the North Atlantic (McDermott *et al.*, 2001; Turney *et al.*, 2005, Swindles *et al.*, 2010a). Furthermore, Baldini *et al.* (2010) demonstrated that the $\delta^{18}O$ of precipitation in Ireland is heavily influenced by air mass history and atmospheric circulation in the North Atlantic. It has also been shown that ocean-atmospheric processes in the North Atlantic have held influence during both Ireland's glacial and post-glacial environments (e.g. McCabe and Clark, 1998; Anderson *et al.*, 2004). As a result, Ireland can be identified as an appropriate location in which to focus a study of the 4.2 kyr event.

Ireland contains a large number of relatively intact peat sequences spanning, in some instances, the entire Holocene (Double, 1954; Smith and Goddard, 1991). Many of these sequences, particularly those in the North of Ireland, contain tephra layers from Icelandic volcanic sources, which have been used to provide exceptional chronological constraint for, and also facilitate the correlation of, palaeoclimatic and palaeoecological records (Pilcher *et al.*, 1995a, 1996; Hall and Pilcher, 2002; Lowe *et al.*, 2004; Plunkett *et al.*, 2004; Swindles *et al.*, 2007a, b, 2008, 2010a; Walker *et al.*, 2011). Ireland also has an immensely rich archaeological record spanning some 9000 years (Woodman, 1985), therefore presenting an excellent opportunity to study past changes in climate within the context of a continuous Irish society.

The lower valleys of the Rivers Main and Bann form an extensive lowland area, overlying tertiary basalts between the Antrim Hills and Sperrin Mountains. This region is home to some of the most extensive peat sequences in Ireland, both in temporal and spatial terms, and which meet all of the criteria outlined in Section 4.2.1. Two ombrotrophic raised bog sites, Sluggan Moss (e.g. Smith and Goddard, 1991; N.B. the work for this paper was actually undertaken in the 1960s, but not published until much later) and Fallahogy Bog (e.g. Smith, 1958) were the focus of the first modern palaeoecological studies in the region, and have been employed together many times since (e.g. Pilcher and Hall, 1992; Hall *et al.*, 1993, 1994a; Pilcher *et al.*, 1995b, 1996; Hall, 2000, 2003; O'Connell and Molloy, 2001).

Figure 4.1 Map showing the location of study sites: a) Sluggan Moss, County Antrim; b) Fallahogy Bog, County Derry. Map adapted, with permission, from Walker *et al*. (2011).



Peat-based proxy climate records (i.e. those that produce BSW/WTD data) from the North of Ireland are surprisingly relatively few in number (Holmes, 1998; Plunkett, 1999, 2006; Barber *et al.*, 2000; Swindles, 2006; Swindles *et al.*, 2007a, b, 2010a; Amesbury, 2008; Plunkett and Swindles, 2008; Langdon *et al.*, 2012) given the immense potential of the peat record there. Furthermore, in spite of the famously long records at certain sites, including Sluggan Moss and Fallahogy Bog (Smith and Goddard, 1991), most studied peat sequences only extend back as far as 4500 cal yr BP at most (e.g. Swindles *et al.*, 2010a). There is subsequently a significant paucity of early- to mid-Holocene age peat records on this climatically important region. Such records could potentially be used to compliment other records from climatically sensitive peatland sites from both seaboards of the North Atlantic realm, within which a number of key palaeoclimatic events have been shown to occur (e.g. Hughes *et al.*, 2000; Hughes *et al.*, 2006; Daley, 2007; Daley *et al.*, 2009). It is clear, that the quality, location and chronological potential of these peatlands makes them an ideal 'testing ground' for ideas regarding the changing nature of North Atlantic climate-forcing mechanisms, and further work must be done (Blundell *et al.* 2008).

4.3 Sluggan Moss, County Antrim

4.3.1 Site description, location and past research

Sluggan Moss, c. 1.5 miles (c. 2.5 km) north-east of Randalstown, County Antrim, is part of an extensive (600 ha) lowland raised bog complex, c. 4 miles (c. 6 km) north of Lough Neagh, and located between the Antrim Hills and the Sperrin Mountains in northern Ireland. The site has an altitude of 42m and is located in the lower reaches of the River Main valley.

Table 4.1 Coordinate, altitude and grid reference information for Sluggan Moss, County Antrim.

Latitude	54°45'56N (54.76556)
Longitude	6°17'37"W (-6. 29361)
Altitude	42m (138 ft)
Irish grid reference (tetrad)	J 09819 92191 (J09W)

The peatbog complex has been largely destroyed by land reclamation and peat exploitation, but Sluggan Moss remains as an intact remnant (108 ha) of the original ombrotrophic raised bog. As a result, Erdtman's (1928) inaugural investigations found the peats to extend back to the 'oldest postarctic period'. This assertion has since been upheld by a number of more recent publications, which demonstrate that the record at Sluggan Moss extends through the Holocene to the Lateglacial and is exceptional in the completeness of its palaeoenvironmental record (Smith and Goddard, 1991; Lowe *et al.*, 2004; Walker *et al.*, 2011). Sluggan's palynological record has proved to be unique in Ireland in terms of its length, its pollen zone scheme has provided the foundation for understanding late-Quaternary vegetation history in the area (Smith and Goddard, 1991) and it is currently considered the regional palynological reference site for northern Ireland (Mitchell *et al.*, 1996). Importantly, the site has also been the source of numerous bog oaks that contributed to the Holocene Irish subfossil pine and oak record (Pilcher *et al.*, 1995b; Turney *et al.*, 2005; Swindles and Plunkett, 2009, 2010; Moir *et al.*, 2010). Published records from the site have demonstrated a relatively fast accumulation rate (c. 20 years/cm; Plunkett, 2006, 2009) and considerable potential for exceptional chronological control. The site is noted for its internal chronological consistency, and provides excellent opportunities for the construction reliable and extensive radiocarbon-based age-depth models (Smith and Goddard, 1991; Lowe *et al.*, 2004) Sluggan Moss has subsequently been used as a model for the development of new chronological modeling techniques (Haslett and Parnell, 2008). Smith and Goddard (1991) demonstrate continuous and steady accumulation at the site, with no significant stratigraphic changes across the mid-Holocene part of the sequence. In addition, Hughes and Barber (2003) suggest the stratigraphy presented by Smith and Goddard (1991) demonstrates initiation of raised bog development began c. 7550 cal yr BP. As a result, the sequence is extremely *Sphagnum* rich to depths that precede this study's period of focus. Of additional chronological significance, an excellent record of time-parallel marker horizons in the form of Icelandic tephra preserved at this site (Pilcher and Hall, 1992, 1995a, 1996; Hall *et al.*, 1993, 1994a, 1994b; Hall and Pilcher, 2002; Hall, 2003; Lowe *et* al., 2004; Plunkett, 2006). This includes Hekla 4, which has been high-precision radiocarbon dated at Sluggan Moss to 4229 – 4345 cal yr BP (Pilcher *et al.*, 1995a, 1996, see Section 6.4), and was therefore deposited sometime around the 4.2 kyr event.

Palynological reconstructions have been used extensively at the site to examine vegetation history and human-landscape interactions over a wide range of timescales, ranging from the Neolithic to the last half millennia (Smith, 1975; Hall *et al.*, 1993, 1994a, 1994b; Hall, 1994, 2000, 2003; O'Connell and Molloy, 2001; Plunkett, 2009). Furthermore, the examination of sub-fossil beetles has been used at Sluggan Moss to reconstruct past landscape changes (Whitehouse, 2006, 2007). Published palaeohydrological reconstructions of BSW, or WTD, are however severely lacking at Sluggan Moss. Plunkett (2006) presents the only such record from the site, in the form of a 1000 yr humification record spanning the c. 2800 cal BP solar minimum. There is, subsequently, significant scope for development of such studies at the site.

Upon inspection in July 2010, the author found surface vegetation to reflect dry bog surface conditions, with *Calluna vulgaris* dominating. It is likely that this is a reflection of the extensive peat cutting that has taken place at the site. *Eriophorum vaginatum* and *Trichophorum cespitosum* were present in some areas, and where small amounts of *Sphagnum* were present, they were typically of the section *Acutifolia*. As a result of the marginal peat cutting, a test core was taken and subject to rudimentary analysis to ascertain the site's suitability for palaeoclimatic reconstruction. The lower sections of the sequence were found to be in good condition and maintained stratigraphic integrity and so the site was accepted for further analyses (see Section 5.2.1).

4.4 Fallahogy Bog, County Derry

4.4.1 Site description, location and past research

Fallahogy Bog, c. 2.5 miles (c. 4 km) northwest of Portglenone, County Antrim, is part of the Wolf Island Special Area of Conservation, c. 11.5 miles (c. 18.5 km) north of Lough Neagh. The site has an altitude of 43m and is located in the lower reaches of the River Bann valley.

Table 4.2 Coordinate, altitude and grid reference information for Fallahogy Bog, County Derry.

Latitude	54°54'42N (54.91155)
Longitude	6°33'40"W (-6.56107)
Altitude	43m (141 ft)
Irish grid reference (tetrad)	C 92274 08053 (C90J)

When compared to Sluggan Moss, Fallahogy Bog can be considered a mostly intact, large lowland raised bog, and has subsequently be designated an Area of Special Scientific Interest (Hall, 2003). Scientific research began at the site with Smith's (1958) palynological study, and was developed by Smith and Willis (1962), who found the peat sequence to extend back to the earliest Holocene period, the pre-Boreal, with the initiation of raised bog development beginning in the Boreal period. The length and nature of the Fallahogy Bog sequence is therefore comparable to that of Sluggan Moss, with the sequence also extremely *Sphagnum* rich to depths preceding this study's period of focus. Smith (1958) and Walker and Walker (1961) have both demonstrated the sequence's potential for palaeoclimatic research, observing dynamic transitions from pool to hummock vegetation, with *Sphagnum cuspidatum* banding and minimal Ericaceous remains.

As at Sluggan Moss, the site has played a central in the construction of regional tephrochronologies and tephra-based studies (Pilcher and Hall, 1992; Pilcher *et al*., 1995a, 1996; Hall, 2003) and dendrochronologies (Pilcher *et al*., 1995b), and Hekla 4 has been identified here. Palynologically focused reconstructions of landscape development and modification are, again, numerous at this site (Hall *et al*., 1993; Hall *et al*., 1994a, Hall, 2000; O'Connell and Molloy, 2001). Palaeoclimatic studies are, again, limited at the site. Barber *et al*. (2000) who employed plant macrofossil and peat humification analyses to reconstruct changes in palaeohydrology over the last 2000 years, registering noticeable wet and/or cool shifts during the Little Ice Age and c. 1400 cal yr BP. Amesbury (2008) employed the site in part of a wider examination of the potential of fine-resolution peat-based analysis through Holocene climatic events, finding evidence for wet and/or cool shifts during two periods of analytical focus, c. 1400 cal yr BP and 3500 cal yr BP, based on plant macrofossil and humification data, thus demonstrating its potential climatic sensitivity.

Upon inspection in July 2010, the author found surface vegetation to, again, reflect relatively dry conditions, with *Calluna vulgaris*, *Eriophorum vaginatum* and *Trichophorum cespitosum* all dominant. *Sphagnum* was present, however, typically with section *Acutifolia* species forming hummocks. *Eriophorum angustifolium* and *Erica tetralix* were sporadically present, and were generally associated with wetter sections of the bog surface topography.

4.5 Modern climate data

A final suggestion, outlined in Section 2.8, was that multiple records from the same region should be prepared using comparative sampling, chronological and analytical strategies, in order to be able to ascertain the degree to which site-specific noise or analytical error was masking the climatic signal (cf. Payne and Blackford, 2008). Coherence in pattern and timing of key palaeohydrological changes demonstrate coherence between hydrological separate peatlands, would suggest a common climatic forcing (Nilssen and Vorren, 1991; Blackford, 1993; Barber *et al.*, 1998; Mauquoy *et al.*, 2002b; Charman *et al.*, 2006; Payne and Blackford, 2008). As a result, this study chose to employ two sites in close proximity to one another, which have thus been subjected to similar climatic and topographic conditions. Owing to their close proximity, modern climate data for Sluggan Moss and Fallahogy Bog are extremely similar. High-resolution gridded datasets provide information about each sites annual precipitation and temperature variation (Tab. 4.3 and Figs. 4.2

and 4.3), demonstrating these similarities. Fallahogy Bog is marginally wetter, by approximately 50 mm/year, with this difference mostly distributed through the autumn and winter months. Sluggan Moss is marginally warmer, by 0.4°C on average through the year, with most of this difference occurring in the summer months. For the purposes of this study however, and the likely effect of these climatic conditions of bog hydrology, these sites are currently subject to identical climatic parameters.

As previously discussed, changes in bog hydrology over time can be considered functions of summer water deficit, where negative effective precipitation is calculated (i.e. precipitation minus potential evaporation) (Charman, 2007; Charman *et al.*, 2009). Data required to calculate potential evaporation is available only in a gridded dataset of slightly lower resolution (0.5° x 0.5°; New *et al.*, 1999) than used for individual site modern climate data. As a result Figure 4.4 presents effective precipitation data for the grid square encompassing both sites, demonstrating that a small deficit is likely to occur at both sites in May, June and July. In oceanic locations, such as Ireland, summer precipitation is likely to be the primary driver of summer water deficit, with summer temperature more influential in continental settings (Charman, 2007). As a result, it has been suggested that oceanic margin sites, such as those chosen here, should be considered more sensitive to palaeohydrological variation (Haslam, 1987; Blackford and Chambers, 1995), such as those witnessed in many records during the largely hydroclimatic 4.2 kyr event. Conrad's Index of Continentality (CIC; Conrad, 1946) provides a quantitative estimate of continentality, with both sites in this study possessing similarly low values (Tab. 4.3) further supporting their similarity.

The marked similarities between the climatic regimes of the two study sites ensures that differences in the palaeohydrological reconstructions from each site can be attributed to locally-specific factors, or autogenic variations in the response of the individual bog. As a result, palaeoecological and palaeoclimatic proxy records developed here can be compared to one another and assessed for the presence of a regional climatic signal (cf. Payne and Blackford, 2008), thus presenting an excellent opportunity to assess whether a coherent 4.2 kyr event signal can be characterised in the region. In addition, stable isotope data from these sites can be compared in an effort to evaluate the regional isotopic signal, and its subsequent potential for contribution to the palaeoclimatic reconstruction.

Table 4.3 Modern climate data (1961-1990) for Sluggan Moss and Fallahogy Bog, northern Ireland (CRU CL 2.0, 10' x 10' resolution; New *et al.*, 2002).

	Sluggan Moss	**Fallahogy Bog**
Average annual precipitation	941 mm	997 mm
Driest month	April (57.6 mm)	April (59 mm)
Wettest month	October (100.6 mm)	October (107.6 mm)
Average annual temperature	8.59°C	8.55°C
Warmest month	July (14.3°C)	July (14.1°C)
Coolest month	January (3.7°C)	January (3.7°C)
CIC	5.586	5.217

Figure 4.2 Average annual precipitation data for Sluggan Moss (dark blue) and Fallahogy Bog (light blue), northern Ireland (CRU CL 2.0, 10’ x 10’ resolution; New *et al*., 2002).

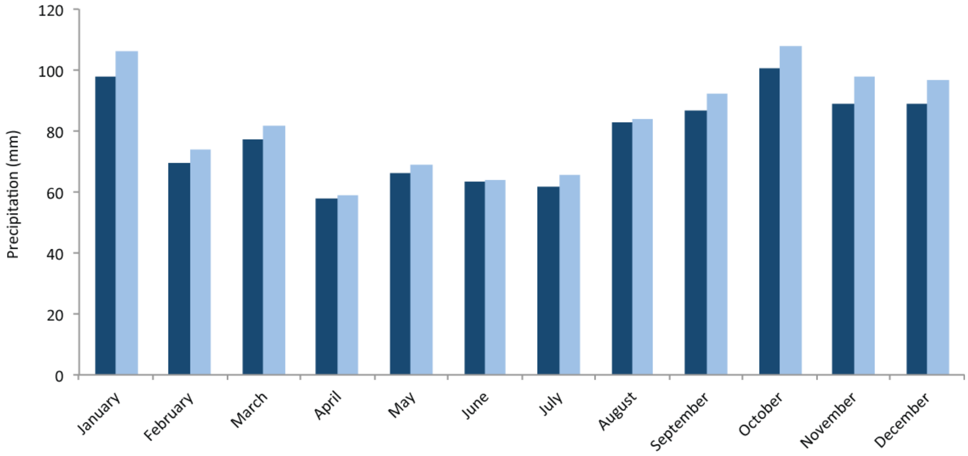


Figure 4.3 Average annual precipitation data for Sluggan Moss (dark red) and Fallahogy Bog (light red), northern Ireland (CRU CL 2.0, 10’ x 10’ resolution; New *et al*., 2002).

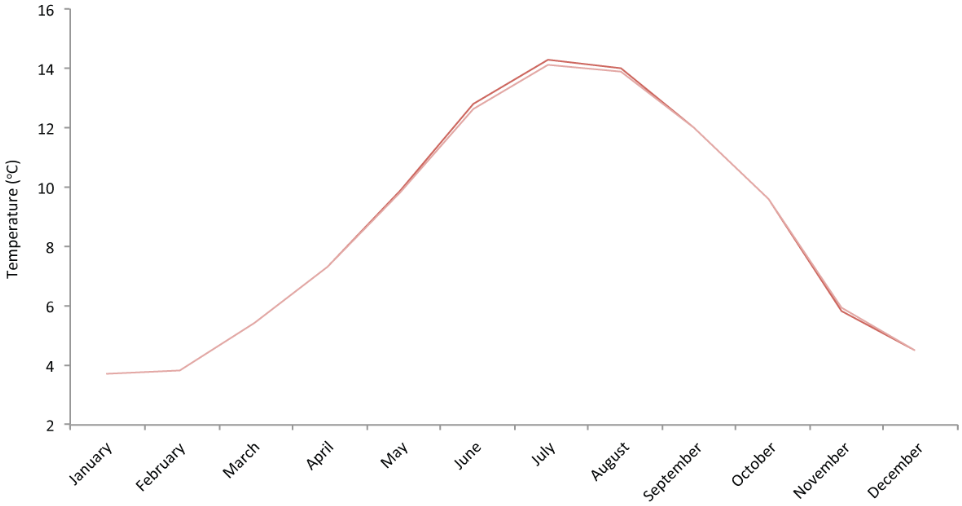
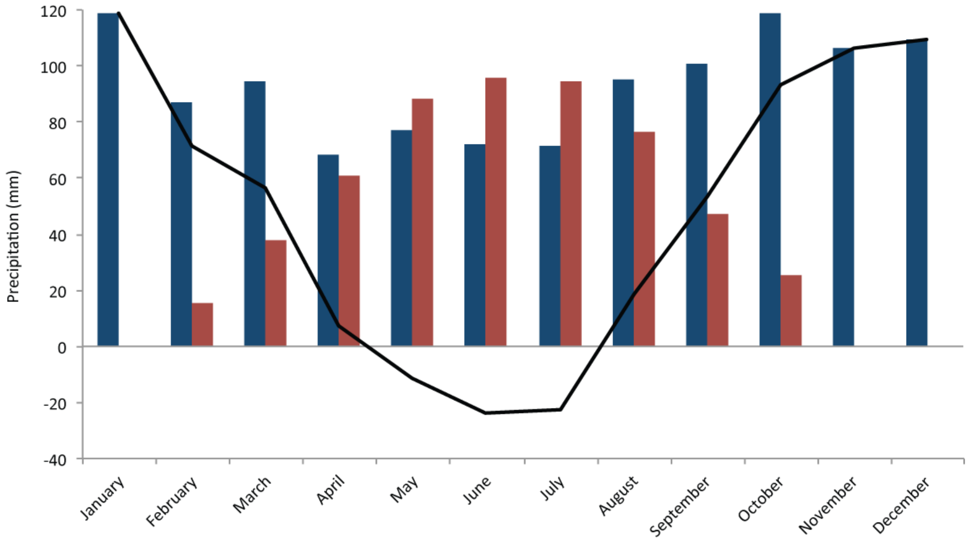


Figure 4.4 Average annual precipitation, potential evaporation and effective precipitation data for Sluggan Moss and Fallahogy Bog (light blue), northern Ireland (CRU CL 1.0, 0.5° x 0.5° resolution; New *et al*., 1999).



Chapter 5 – Methods

5.1 Introduction

This chapter describes the field, laboratory and statistical methods employed in this study. A detailed discussion and rationale for the selection of the various palaeoecological and isotopic techniques can be found in the previous chapter. Chronological methodologies can be found in Chapter 6.

5.2 Field methods

5.2.1 Field sampling techniques

At Sluggan Moss, field stratigraphy was assessed by eye, alongside Ordinance Survey data and aerial photography, to help determine an optimal coring location. A test core was taken using a 100 x 5 cm Russian pattern corer and inspected in the laboratory. Suitability for palaeoclimatic analysis was confirmed by the presence of evidence for potential climatic shifts (e.g. the presence of pool layers indicating shifts to cooler and/or wetter conditions) and an abundance of *Sphagnum* moss (Barber *et al.*, 1998). In addition, the Hekla 4 tephra layer was identified. Consequently, a master core was taken adjacent to the test core using a 50 x 10 cm Russian pattern corer at a later date. Core overlaps of 10 cm were enabled through the use of two adjacent bore holes, preventing the need to sample from potentially disturbed extremities of each core, and also ensuring that the entire depth of the peat cored is available for analysis from a non-overlapping section, in accordance with the ACCROTELM project coring protocol (Amesbury, 2008). At Fallahogy Bog, fieldwork methods endeavoured to follow the optimal procedure for peat bog coring, as outlined by Barber *et al.* (1998). Here, field stratigraphy cores were taken along two perpendicular transects using a 100 x 4 cm Russian pattern corer. Again, pool layers and an abundance of *Sphagnum* moss indicated the coring location's suitability for palaeoclimatic analysis. The master core at Fallahogy Bog was taken using a 30 x 11 cm Russian pattern corer with a 15 cm overlaps. All fieldwork at Fallahogy Bog was undertaken by Dr Matthew Amesbury (Amesbury, 2008). Corers of 10 and 11 cm bore sizes were employed at Sluggan Moss and Fallahogy bog, respectively, to maximise the amount of peat available for multi-proxy analysis. Subsequently, the cores presented c.40 and 48 cm^3, respectively, for analysis per 1 cm slice.

5.2.2 Sample storage and sampling procedure

Upon collection, cores were placed in labelled plastic guttering, wrapped in cling-film and/or sealable carbon-stable plastic bags and sealed using electric tape. Care was taken to ensure that cores were transported securely and horizontally to the laboratory. Cores were then stored in a cold store at a temperature of c. 4°C to minimise biological activity.

The sampling procedure followed that which is conventional for peat-based palaeoclimatic research. Non-contiguous samples of 1 cm stratigraphic depth were manually removed for plant macrofossil, testate amoebae and stable isotope analysis using a scalpel and spatula. Scissors were used, where necessary, to cut through root and other material. For humification, contiguous samples of 1 cm stratigraphic depth were taken. For tephra analysis, contiguous, bulked samples of 5cm stratigraphic depth were initially taken. Contiguous samples of 0.5 cm stratigraphic depth were then taken across any 5 cm in which a potential tephra horizon was identified. All cutting tools were thoroughly cleaned between samples to prevent contamination. The remaining material was wrapped in cling-film to prevent water loss over time, and archived in the cold store for future work.

5.3 Palaeohydrological methods

The techniques described here have been chosen to achieve the latter two aims of this study, namely: i) characterising the 4.2 kyr event signal the 4.2 kyr event in Great Britain and Ireland; and, ii) evaluating the contribution of peat-based stable isotopic analysis to an examination of the nature and/or cause of the 4.2 kyr event. Three techniques are routinely employed in peat-based palaeohydrological studies, plant macrofossil, humification and testate amoebae analysis, and a rationale for their inclusion here is outlined in Sections 3.4 and 3.5. The importance of the multi-proxy approach both to peat-based palaeohydrological studies generally, and also to this study specifically, is outlined in Section 3.5.

5.3.1 Plant macrofossil analysis

The Quadrat and Leaf Count (QLC) technique (Barber *et al.*, 2003), adopted here, traditionally involves sub-sampling of the core to 4cm^3 volumes. However, Amesbury *et al.* (2011) have shown that reducing this sample size from 4 to 2cm^3 does not affect the interpretation of the results. As this study also requires sub-sampling of the material for stable isotopic analysis of *Sphagnum* α-cellulose, a technique that requires relatively large amounts of material, this smaller sample size was adopted. Subsequently, sub-samples of 2cm^3 were extracted from the core using individually cleaned stainless steel sampling tools, and care was taken to snip through roots and twigs before sample removal.

Samples were washed through a 125μm sieve under a constant jet of water to disaggregate the plant remains and remove fine detritus. Care was taken when rinsing, to minimise damage to plant macrofossils and charcoal, by keeping the sieve just below water surface (Mauquoy *et al.*, 2010). A standard 2.5l of water was used for sieving, in order to allow comparison of the quantity of unidentified organic matter (UOM) present, by minimising bias in the amount of UOM that passes through the sieve. Sample residues were then stored in distilled water in Sterelin tubes and labelled accordingly. Samples were stored, ready for analysis, in a cold store at c. 4°C to minimise further decomposition.

The material was then gently stirred, partially decanted into a petri dish, and microscopically inspected using a low powered stereo-zoom microscope at x10 magnification, with a 10x10 square grid graticule in one of the eyepieces. This process was repeated until all the remains had been scanned. The petri dishes were moved randomly to a total of 15 different views, and plant macrofossil remains were estimated as percentages of each view within the graticule. Percentage empty space was also recorded, before the results were averaged to represent the whole sample (Mauquoy *et al.*, 2010). Identifications were made using a range of reference material (Grosse-Brauckmann, 1972; Daniels and Eddy, 1990; Smith, 2004; Mauquoy and van Geel, 2006), as well as type specimens and photos provided by Dr Matthew Amesbury, and drawings provided by Dr Paul Hughes.

Sub-samples which contained well preserved epidermal tissues of monocotyledon species were temporarily mounted on microscope slides, using water, and examined at higher (x100 – x400) magnification where necessary, to refine identification to species level. Occasionally, certain monocotyledon species (e.g. *Eriophorum vaginatum*) remains could be identified without additional magnification (Mauquoy *et al.*, 2010). Where several components of a single species were present (e.g. leaves, stems, roots, seeds etc.), each was recorded separately to aid future interpretation of the plant macrofossil record. Roots, for example, often penetrate older sediments, whereas seeds can be distributed more widely from their parent plant. Any fruits, seeds and/or macroscopic charcoal fragments were recorded as total number (n) present in the sample.

Upon completion of these 15 views, a random selection of 100 *Sphagnum* leaves (avoiding bias towards larger and/or less degraded leaves) were mounted on a slide, using aquamount, examined at x400 magnification, and identified to species level, where possible, at x400 magnification. Where *Sphagnum* leaves were exceptionally low in concentration or preservation quality, the maximum number of leaves feasible was identified. *Sphagnum* leaves were identified using Daniels and Eddy (1990), Smith (2004) and Mauquoy and van Geel (2006), as well as reference material and photos, provided by Dr Matthew Amesbury, and drawings provided by Dr Paul Hughes. Results were expressed as percentages of total identifiable *Sphagnum*.

5.3.2 Peat humification analysis

The procedure adopted in this study is an adaptation of the standard colorimetric technique (Blackford and Chambers, 1993). Amesbury *et al.* (2011) found that reducing the standard sample volume from 0.2g to 0.1g dried peat did not affect the interpretation of results, and therefore this reduction is employed here. As a result, all other quantities could be reduced by half from the standard. This is especially beneficial given the multi-proxy nature of this study, and in particular, the amount of material required for stable isotopic analysis (see Section 5.3.4).

Samples of approximately 1-2cm^3 were dried in porcelain crucibles overnight at 105˚C. The dried samples were then ground to a fine powder using a pestle and mortar. Of this powder, precisely 0.1g (± 0.0001g) was weighed out into a 100ml volumetric beaker. 50ml of freshly mixed 8% NaOH extractant was added and the samples were brought to the boil and left to simmer gently for one hour to extract the humic acids, whilst minimising evaporation. Upon cooling, the samples were topped up to 100ml using distilled water and mixed thoroughly before being filtered through Fisherbrand QL1000 qualitative filter paper (equivalent to Whatman Qualitative 1 filter paper). Next, 25ml of this filtered solution was diluted to 50ml using distilled water and mixed well.

Percentage light transmission was then measured using a Jenway 6315 spectrophotometer. The spectrophotometer was set at 540 nm and calibrated to 100% light transmission using distilled water between each reading. Each reading was taken three times, and an average calculated and taken as the final value. Care was taken to ensure that all readings were taken four hours after initial mixing with NaOH to ensure consistency of results.

Results were recorded as percentage light transmission, with higher concentrations of humic acids darkening the extracts, and indicating well humified peat from which a drier bog surface can be inferred. Conversely, lighter extracts with lower concentrations of humic acids suggest less humified peat, as a result of a wetter bog surface (Anderson, 1998).

5.3.3 Testate amoebae analysis

Preparation for testate amoebae analysis in this study followed the standard technique outlined by Hendon and Charman (1997) and Charman *et al.* (2000), with minor modifications. Again, Amesbury *et al.* (2011) found that reducing sample size from 2 to 1cm^3 made no significant difference to the assemblages recorded. Consequently, in the interests of preserving material for stable isotopic analysis, the lower sample size was adopted here. Subsequently, 1cm^3 samples, as measured by water displacement, were placed in a 250ml glass beaker with 100ml of distilled water. In order to enable the calculation of testate amoebae concentration a known quantity *Lycopodium* spore tablet was added. The samples were then boiled for ten minutes, with occasional stirring, to encourage disaggregation of the peat and spore tablet before being washed through 300 and 10μm sieves using distilled water. The residue on the 10μm sieve was then washed into 15ml centrifuge tubes and centrifuged at 3000 rpm for 5 minutes. The supernatant was discarded and the centrifuge process repeated. The resulting concentrate was stored in distilled water in stoppered glass vials, labelled accordingly and retained for analysis. Water was chosen as a mountant, ahead of glycerol (cf. Hendon and Charman, 1997), as it provides better visibility for taxonomic identification from test morphology (Charman *et al.*, 2007). Prior to microscope analysis the concentrate was mixed gently and a drop mounted on the microscope slide together with two drops of distilled water to maximise microscopic clarity and definition (Barber and Langdon, 2001). Testate amoebae were examined at between x100 and x400 magnification for morphology, composition, size and colour. Identifications were made using the standard dichotomous key of Charman *et al.* (2000), as well as various diagrammatic and photographic identification guides (Ogden and Hedley, 1980; Lee *et al.*, 2001; Booth *et al.*, 2010). A morpho-species approach was used in certain circumstances, where a designation that includes other species has been designated as a ‘type’ (cf. Charman *et al.*, 2000). Taxonomy followed Charman *et al.* (2000), except for the addition of *Centropyxis ecornis* (after Booth, 2008).

At least 100 individual tests were counted for each level, as this is deemed adequate to represent the main faunal components of each assemblage (Payne and Mitchell, 2009). Although more extensive counts may occasionally reveal rarer taxa, it is unlikely that this would add significantly to the environmental interpretation (Charman, 1999). Indeed, counts of as low as 50 may be sufficient in some circumstances (Payne and Mitchell, 2009) and may still be accepted for statistical analysis (Swindles *et al.*, 2007b), counts of this size were avoided in all but a very small number of levels where testate concentration and/or preservation was exceptionally poor.

5.3.4 Stable isotopic analysis

Preparation of *Sphagnum* α-cellulose for stable isotopic analysis consists of three stages: i) isolation of *Sphagnum* from the peat matrix; ii) extraction of α-cellulose; and iii) analysis of α-cellulose by mass spectrometry.

5.3.4.1 Isolation of *Sphagnum*

In an extensive study of the use of *Sphagnum* in stable isotopic analysis, Daley (2007) trialled a number of methods for isolating *Sphagnum* remains including sieving, centrifugation, density separation and swilling. A combined approach of sieving and swilling proved the most successful technique. However, Daley (2007) was attempting to isolate *Sphagnum* leaves, as opposed to the stem material chosen for analysis in this study (See Section 3.4.4.8). As a result, a combination of sieving and micromanipulation was found to be most effective.

During stacked sieving of bulk peat samples Daley (2007) found that *Sphagnum* section *Sphagnum* leaves largely collected on a 500μm sieve, whereas *Sphagnum* section *Acutifolia* and *Cuspidat*a collected on 250μm sieves. *Sphagnum* stem and branch material of all *Sphagnum* sections, often with leaves still attached, generally collected in the 1000μm sieve. As a result, bulk peat samples of approximately 4cm^3 were gently washed through a 1000μm sieve with distilled water. The residual material was then washed into a petri dish where it could be examined microscopically using a low powered stereo-zoom microscope at x10 magnification. Stem material was then stripped of branches, leaves, extraneous material and potential contaminants. Approximately 200 mg of *Sphagnum* stem material was collected per sample to ensure sufficient α-cellulose could be extracted for stable isotopic analysis.

5.3.4.2 Extraction of α-cellulose

The advantages associated with the use of α-cellulose in stable isotopic analysis have been discussed in Section 3.4.4.7. The procedure used in this study for the extraction of α-cellulose from *Sphagnum* stems follows a procedure developed by Daley (2007), albeit with slight modifications. Daley's protocol was itself based on a procedure intended for use with wood slivers (Green, 1963; Loader *et al.*, 1997; Rinne *et al.*, 2005).

Daley (2007) found that filtration of material was problematic in some stages of the extraction, due to the formation of a gel-like substance following chemical treatment. This occurred as a result of the simple cellular *Sphagnum* leaves breaking down into very fine-grained material. This problem was not encountered in this study as cellulose extracted from the *Sphagnum* stems maintained its polysaccharide structure, and subsequently filtration was relatively straightforward. The maintenance of structure did, however, have implications for the ultrasonic homogenisation of the α-cellulose, with this part of the procedure taking somewhat longer than the 30 seconds practised by Daley (2007). Tillman *et al.* (2010) suggest crushing and freeze drying stems before homogenisation to aid the process, but this was found to be unnecessary in this study.

Raw samples of isolated and cleaned *Sphagnum* stem material were placed in coarse sintered glass extraction tubes (Soxhlet vessels) as designed and commissioned by Daley (2007) specifically for *Sphagnum*-based cellulose extraction. The sinters in these vessels are placed closer to the mouth of the vessel, thus providing greater surface area than the equivalent vessels used in wood-based cellulose extraction (Loader *et al.*, 1997), consequently improving filtration rates. The extraction tubes were then suspended in a water bath heated to 80°C, and acidified sodium chlorite bleach ($NaClO_2$) was added to the samples until the stem material was completely covered. The samples were then allowed to oxidise for 50 minutes before being actively drained using a water jet vacuum pump. This bleaching process was then repeated a further four times, to complete five bleach treatments in total. Upon completion of the fifth bleach treatment, the samples were washed through with boiling distilled water and actively drained *in vacuo*, followed by a further four room temperature distilled water washes. This ensured the removal of bleaching treatment agents from the residue. The drained samples were then placed back in the water bath, this time at 70°C, and 10% conc. sodium hydroxide (NaOH) was added to the samples until stems were submerged. The samples were left for 45 minutes before being actively drained and washed through a further three times with distilled water. The water bath was returned to 80°C and samples were subjected to a further two bleach treatments, as described previously. Finally, the samples were washed through with distilled water and actively drained a further five times to ensure the removal of all extraneous chemicals before the α-cellulose could be prepared for ultrasonic homogenising.

Samples were then transferred to 2 ml plastic Eppendorf tubes using a metal spatula and filled with 1ml of 4°C distilled water. An ultrasonic probe was then used to homogenise the material. This took between 1 and 5 minutes dependent on the sample, but was continued until all visible polysaccharide strands had been broken down and the mixture was cloudy, indicating that the α-cellulose was fully homogenised. Samples were then freeze dried for two days to complete the α-cellulose extraction and leave samples ready for analysis by mass spectrometry.

5.3.4.3 Analysis of α-cellulose by mass spectrometry

Traditionally, the stable isotopes of carbon and oxygen are measured separately, using different combustion and 'low' temperature (i.e. 1000°C) pyrolysis techniques, respectively (see Young *et al.*, 2011). However, a desire to develop faster methods for reliable stable isotope analysis has encouraged researchers to investigate the potential for simultaneous measurement of these isotopes, through the 'high' temperature (i.e. 1400°C) pyrolysis of cellulose, with several initial studies reporting positive results (e.g. Saurer *et al.*, 1998; Knöller *et al.*, 2005). Following extensive investigative work into the use of pyrolysis techniques to obtain $\delta^{13}C$ results by Young *et al.* (2011), Woodley *et al.* (2011) presented a novel technique facilitating the simultaneous measurement of the stable isotopes of carbon and oxygen using high temperature pyrolysis. This study adopted this new technique on the grounds of it thereby reducing both the amount of sample required for analyses, alongside the analytical costs associated with separate measurements. The benefits of this adoption were deemed to outweigh the detriment of not obtaining δD data. As Section 3.4.4.5 describes, of the three stable isotopes in question, the least is known about the biomechanics associated with δD uptake. As a result, the considerable financial and temporal commitment of employing a dual measurement approach (i.e. combustion for $\delta^{13}C$; pyrolysis for $\delta^{18}O$ and δD) was deemed to be unwarranted.

Between 0.3 to 0.35 mg of dry α-cellulose was weighed into silver capsules using a Sartorius Supermicro fine balance. The capsules were then crimped to prevent sample loss, and freeze-dried again alongside standards to nullify the affects of atmospheric moisture, before being placed in an equilibration chamber and subjected to a continuous flow of 'Meerwasser' water vapour of a known isotopic composition (+1 ‰ vs. VSMOW) at 110°C at a rate of 7.8 μL/min, delivered by a helium carrier gas stream at 40 mL/min. After 9 minutes the water supply was stopped and the sample transferred into an AS128 autosampler, purged with helium gas to, again, nullify the effects atmospheric moisture, before being moved to the pyrolysis reactor after a few seconds. Pyrolytic conversion took place in a Thermo-Scientific FLASH HT Elemental Analyser set to 1400°C, over a glassy carbon granulate on a 2mm layer of silver wool and a 20mm bed of quartz wool. Pyrolysis products were then separated using a 1m GC column, with a 5 Å molecular sieve, and transported via a ConFlow IV open split unit to a ThermoFinnigan Delta V isotope ratio mass spectrometer.

Stable isotopic ratio variance in precipitation (i.e. $\delta^{18}O$) is very small (typically in the range of the third to fifth decimal place) due to its low natural relative abundance in the water molecule and is therefore expressed as *per mil* (‰) (N.B. not to be confused with 'ppm', which is commonly pronounced 'parts per mil', but which actually refers to 'parts per million'). In water, stable isotope ratio variance is typically expressed using delta notation (i.e. δ per mil deviation) relative to a standard (normally Vienna standard mean ocean water (V-SMOW), which represents the mean isotopic composition of ocean water) (Craig, 1961a, b; Coplen, 1994, 1995a, b; Pendall *et al*., 2001). Variations in carbon isotope ratios ($\delta^{13}C$) are measured against the Vienna Pee Dee Belemnite (VPDB) standard (Coplen *et al*., 2006; Tillman *et al*., 2010). All isotopic ratio measurements are expressed in this study using standard delta notation relative to V-SMOW and VPDB, respectively, and accordingly.

5.4 Statistical analysis of the proxy data

Results produced by each palaeoecological technique were subjected to varying statistical analyses or transformation to enable effective interpretation as well as inter-comparison between proxies, and records from other sites. Palaeoecological diagrams throughout this study were produced using *Tilia* and *TGView* (Grimm, 1993; 2004). Stable isotope data were not subjected to statistical analyses or transformation of this sort.

5.4.1 Plant macrofossil analysis

For ease of interpretation, it has become common practice for multivariate plant macrofossil data to be transformed to follow a univariate distribution. Any given assemblage down-core can therefore be condensed to relate to a particular point on a gradient of hydrological conditions. Plant macrofossil datasets from both sites were subjected to two primary methods of transformation: i) Dupont hydrological index (DHI), and ii) detrended correspondence analysis (DCA), in an effort to quantitatively summarise past fluctuations in BSW.

Both the above techniques possess associated problems that can limit their effectiveness in reconstructing past palaeohydrological changes, as summarised in the following sections. Whilst DCA may be more frequently utilised (see examples in Section 5.4.1.1), both the discussed techniques have been successfully employed together in past plant macrofossil studies (e.g. Barber *et al*., 2003; Swindles *et al*., 2007b; Daley and Barber, 2012), with replicable palaeohydrological curves developed using both DHI and DCA methodologies. As a result, both techniques will be applied in this study in order to achieve a comprehensive analysis of the plant macrofossil data.

5.4.1.1 Detrended correspondence analysis

DCA is an ordination technique that can be used to identify latent environmental gradients within ecological data using an objective approach (Birks, 1995). The technique has been employed extensively in peat-based palaeoclimate studies to identify proxy-climate indices, typically BSW from plant macrofossil data (e.g. Barber *et al*., 1994, 2000, 2003, 2004a, Charman *et al*., 1999; Chiverrell, 2001; Hughes *et al*., 2000, 2006; Langdon *et al*., 2003; Langdon and Barber, 2005; Mauquoy and Barber, 1999a; Swindles *et al*., 2007b; Amesbury *et al*., 2011, 2012; van der Knaap *et al*., 2011; Väliranta *et al*., 2012; Daley and Barber, 2012). This section serves to discuss only the basic premises of DCA in the context of palaeoecology, as an extensive discussion of the concepts and mathematical algorithms associated with DCA can be found elsewhere (ter Braak and Prentice, 1988; Kent and Coker, 1992; Kovach, 1995; ter Braak, 1995).

In essence, ordination techniques order data in any number of dimensions to approximate patterns of response. Subsequently, in an ecological context, ordination is used to examine the relationship between species composition and underlying environmental gradients (Digby and Kempton, 1997). DCA is classed as an indirect ordination technique as, unlike canonical correspondence analysis (CCA), it operates with no *a priori* knowledge of the environmental variables influencing species distribution. DCA evolved from correspondence analysis (CA) after the negative effects of the ‘arch effect’ were documented, whereby a mathematical artefact linked the two axes and species and samples were distributed not linearly, but instead adopted an arch shape, resulting in erroneous interpretations (Kovach, 1995; ter Braak, 1995).

DCA adopts a two-way weighted-averaging algorithm, assuming a unimodal species response to environmental variables (ter Braak, 1995). As a result, species response alone is modelled, with the underlying environmental variable left open to ecological interpretation. In the study of plant macrofossils in ombrotrophic peatlands, DCA is used to check that the main macrofossils are distributed along a hydrological gradient and then to assign a relative wetness value for the fossil assemblages down-core (see above references).

DCA operates by extracting ordination axes, or eigenvectors, that explain as much of the underlying variance as possible. Each axis is subsequently assigned an eigenvalue, which places a quantitative value on its contribution to the explanation of total variance within the data (Kent and Coker, 1992) relative to the maximised dispersion of species cores along the length of the axis (ter Braak, 1995). The first axis, with the largest eigenvalue, accounts for the most variance in the dataset, with subsequent axes explaining gradually less of the variance. Generally, eigenvalues of close to or greater than 0.5 are considered to represent a good separation of species along each axis (ter Braak, 1995) and interpretation of their potential ecological representation can begin. Low eigenvalues can be interpreted as a failure of DCA to identify a significant underlying, driving variable. Therefore, the interpretive power of DCA relies not only upon the ability of an axis to account for a satisfactory portion of total variance within the data, but also the confidence with which it can be realistically assigned to an environmental variable.

The principal limitation of DCA occurs when crossed gradients are experienced within a dataset, which can limit the interpretive power of the technique (Daley and Barber, 2012). Crossed gradients occur when the dominant driving variable is replaced by an alternative driver, and are often characterised by one or more samples possessing significantly different characteristics from the rest of the dataset. These significant changes in species assemblage are typically driven by non-climatic factors, and can potentially skew the eigenvectors, making other variables appear less significant. The best example of this in relevant studies is the localised 'extinction' of *Sphagnum austinii* in Great Britain and Ireland and its subsequent replacement by *Sphagnum magellanicum* and *papillosum*. The precise drivers of this change are still debated (see Section 3.4.1.2). However, two species, confidently believed to possess the same ecological niche, can be combined for the purposes of ordination to strengthen the eigenvalue of axis one (e.g. Barber *et al.*, 1994). Crossed gradients can also be caused by ecological succession. Daley and Barber (2012) encountered this problem in the form of a bipartite division at Walton Moss, Cumbria, when early- to mid-Holocene *Eriophorum*/*Calluna* peat was replaced by *Sphagnum*-dominated assemblages in the mid- to late-Holocene. It is likely that this transition owed more to bog ontology than climate (Hughes and Barber, 2004), thus significantly distorting the DCA analyses. A related limitation of DCA is its susceptibility to outlying or occasionally dominant taxa. However, any skewing of eigenvectors can be rectified by removing problematic taxa from the analyses (Kent and Coker, 1992).

The effectiveness of DCA can also be influenced by species diversity and the often large hydrological tolerances of some species. Such issues can result in all other taxa appearing as relative outliers regardless of their own hydrological tolerances. Issues such as these can only be identified during the course of the analysis, and often prevent axis one being interpreted as a hydrological gradient (e.g. Blundell and Barber, 2005). However, providing these issues are taken into consideration during analysis and interpretation, DCA can produce useful environmental reconstructions in peat-based palaeoclimatic studies.

For the purposes of DCA, different components originating from the same species are typically combined to form a single taxon. For example, in this study *Eriophorum vaginatum* leaves, stems, roots, nodes and spindles were combined and collectively classed as *E. vaginatum*. If this was not undertaken, it could result in different components of the same plant species being located at varying points on the DCA axes, which is clearly ecologically unreasonable.

DCA was undertaken in this study using CANOCO for Windows (v. 4.52) (ter Braak, 1987; ter Braak and Šmilauer, 1998). Initially, the results of DCA are presented as bi-plots of axes one and two species and sample scores. Species, which are frequently found to coexist in the dataset, will occupy similar positions on axis one, whilst those, which are rarely associated, remain separate. This subsequently enables the assessment of potential environmental gradients by eye, as taxa responding similarly to the variable assigned to axis one will naturally group and separate from those which respond differently. The same principle applies for sample scores, with assemblages spaced along axis one based on their compositional similarities (ter Braak, 1995). It should then be possible to relate this gradient to a likely environmental factor, which is most often, in this context, species tolerance to WTD, based on a prior understanding of each species' ecological preferences. Should an environmental variable be confidently assigned to this gradient, axis one scores can be plotted against down-core age and depth to produce a proxy-climate index (cf. Barber *et al.*, 1994).

Down-weighting of rare species is often adopted in instances where the sporadic occurrence of a particularly rare species could skew DCA results, thus placing artificial or unrealistic importance on the presence of such a taxon (ter Braak and Šmilauer, 1998). However, at both sites, down-weighting of rare species did not achieve higher axis one eigen values and was therefore deemed unnecessary. Similarly, the threshold for a species' inclusion in DCA is sometimes set at 2 or 5 % total abundance (Hill and Gauch, 1980). However, DCA performs optimally with the presence of more species (Barber *et al.*, 1994) and this practice was not found to improve axis one eigenvalues, and so all species were included in the analyses.

5.4.1.2 Dupont hydrological index

The DHI is a method of estimating relative hydrological conditions based on the weighted averaging of plant macrofossil data. The technique, developed by Dupont (1986), has since been successfully employed in a number of studies (e.g. Mauquoy, 1997; Barber *et al.*, 2003; Daley, 2007; Swindles *et al.*, 2007a, b; Amesbury, 2008; Daley and Barber, 2012).

Each taxon is weighted, based on ecological knowledge of the hydrological conditions under which it naturally grows, and assigned an indicator value on a scale of 1-8. For example, species that are generally associated with drier conditions, such as members of the Ericaceae family and hummock-forming sections of the genus *Sphagnum*, are assigned lower values. Conversely, Sphagna associated with pool formation are assigned higher values. Dupont's (1986) original hydrological index weightings were revised and expanded by Mauquoy (1997) and Daley and Barber (2012) to include species commonly found in ombrotrophic bog ecosystems (Table 5.1).

Table 5.1 Dupont hydrological index species weightings.

Taxon/Plant macrofossil type	Dupont weighting
Ericaceae undiff.	8
Calluna vulgaris	8
Erica tetralix	8
UOM	8
Monocotyledonae undiff.	7
Eriophorum vaginatum	7
Trichophorum cespitosum	6
Brown moss undiff.	6
Sphagnum section *Acutifolia*	5
Sphagnum austinii	4
Rhynchospora alba	3
Eriophorum angustifolium	2
Sphagnum section *Cuspidata*	1

Weighted average scores for each sample are calculated using the following equation:

$$x_0 = \frac{\sum_k Y_k u_k}{\sum_k Y_k}$$

where: x = Dupont hydrological index; x_0 = value of x for a given sample; Y_k = abundance of the k th species; u_k = indicator value of the k th species (following ter Braak and Barendregt, 1986). Effectively, each taxon's indicator value is multiplied by its percentage abundance in a given assemblage, before being added together and divided by the total abundance of all species, thus deriving an average wetness value for the sample in question. Whilst employed with success, as previously discussed, concerns have been expressed regarding the lack of field data incorporated when assigning species indicator values in the DHI, potentially leading to a degree of subjectivity in the analyses (Barber *et al.*, 2000, 2004). It would be possible to argue for the logic of giving greater credence to species that possess narrow ecological niches, when compared to those with a wide tolerance of WTDs, capable of growing across the microtopographical gradient (ter Braak, 1995). Despite this, the DHI assumes that each species is an equally good indicator of WTD.

Whilst DCA has been more broadly employed in these studies, DHI does, in fact, possess a number of advantages over DCA. For example, DCA frequently requires data to be omitted to achieve functional ordinations. Conversely, DHI can routinely incorporate all data into the analyses without distortion. Also, as discussed above, there is a possibility that axis one in DCA can bear no clear relation to an environmental variable (e.g. Blundell and Barber, 2005). DHI, however, is based upon well-established and informed ecological knowledge of species response to BSW and, therefore, despite the degree of subjectivity associated, it is always likely to produce meaningful results (Daley and Barber, 2012), thus justifying its inclusion in this study.

5.4.2 Peat humification analysis

Peat humification results are often displayed as detrended residuals, following linear regression of the raw data, to remove the down-core tendency towards a higher degree of humification (e.g. Blackford and Chambers, 1995; Langdon *et al*., 2003; Blundell and Barber, 2005; Borgmark and Wastegård, 2008; see Section 3.4.2.2). This trend is caused by increased proportions of humic acids present in peat subjected to continued decomposition through the catotelm (Clymo, 1970, 1984).

5.4.3 Testate amoebae analysis

5.4.3.1 Detrended correspondence analysis

Whilst its application to plant macrofossil data is far more established, as discussed in detail in Section 5.4.1.1, DCA can also be applied successfully to testate amoebae data (e.g. Chiverrell, 2001; Blundell and Barber, 2005). Other ordination techniques have also been applied to testate amoebae data (e.g. CCA: Tolonen *et al*., 1994; Woodland *et al*., 1998), but are less widely employed. DCA was undertaken in identical fashion to that outlined in Section 5.4.1.1. The bi-plots were examined for evidence of a water table tolerance gradient by inspecting species distribution along axis 1 in conjunction with hydrological indicator values and ecological preference information outlined in Woodland *et al*. (1998) and Charman *et al*. (2000). Testate amoebae species with low abundances or those that occur only rarely were omitted, where appropriate, to avoid the distortion of analyses caused by outliers, thus improving axis 1 eigenvalues (Blundell and Barber, 2005).

5.4.3.2 Transfer functions

Whilst applied to a variety of palaeoecological assemblage data to reconstruct a range of environmental variables, of most interest to this study are the transfer functions that have been developed for the analysis of testate amoebae, to reconstruct changes in WTD in ombrotrophic peatlands. As discussed earlier, such transfer functions have been developed for a number of peatland regions (see Section 3.4.3.1).

Transfer functions facilitate the reconstruction of an environmental variable (X_0; e.g. WTD) over time, based on palaeoecological data (Y_0; e.g. testate amoebae). In the development of the transfer function, a ‘training set’ of modern data is created to enable the responses of analogous modern taxa from surface samples (Y) to variation in the environmental variable (X) to be modelled. This modelled response can subsequently be applied to Y_0 data, to quantitatively estimate X_0, via the transfer function (Birks, 1995, 2003).

Whilst the development of transfer functions has raised the profile of testate amoebae for palaeoenvironmental reconstruction, the technique is not without its associated problems. The precision, accuracy and confidence associated with transfer functions relies heavily on the development of a 'robust' training set. In principle, this training set must conform to a number of criteria in order to provide a reliable basis on which to model species response. In effect, the developers of transfer functions strive to include as many samples from across the environmental gradient, and from as many sites, as possible. However, Booth (2008) outlines a primary concern in that most training sets are based on single-point measurements of WTD, taken at the time of sampling, despite the fact that surface samples of testate amoebae incorporate several years of assemblage accumulation. It has also been shown that testate amoebae assemblages exhibit seasonal variability, presumably in response to changing hydrological conditions through the year (Warner *et al*., 2007; Sullivan and Booth, 2011). As a result, comparison of species-environment relationships between studies that took place at varying times of year can be problematic. However, whilst the inclusion of annually averaged water table data has been shown to provide more meaningful reconstructions (Woodland *et al*., 1998), Booth (2008) acknowledges that long term hydrological data are rarely available at relevant spatial and temporal densities, and are often prohibitively expensive to obtain.

Another problem, highlighted by Booth (2008), is that many training sets possess relatively small geographic sampling ranges. Despite the fact that many taxa can be considered 'cosmopolitan' in distribution and occupy similar ecological niches in different regions (e.g. Booth and Zygmunt, 2005), Charman *et al*. (2006) found that the application of a transfer function developed in one region, to a dataset collected from another, was often problematic. Such geographically limited sampling may also account for the lack of good modern analogues for certain fossil assemblages (Charman, 2001, see Section 3.4.3.2). Booth (2008) suggests that larger sampling networks and associated longer-term environmental data would improve our understanding of testate amoebae ecology and biogeography, thus allowing more accurate and precise transfer function-based palaeoenvironmental inferences to be made. Furthermore, Telford and Birks (2005, 2009) maintain that training sets produced from sites geographically close to one another, and in which species assemblages and environmental conditions are extremely similar, creates a lack of statistical independence as a result of spatial autocorrelation. It is claimed that this can lead to inappropriate model choice and misleading, over-optimistic estimates of a transfer function's efficacy.

The widely applied transfer function for Great Britain and Ireland (e.g. Charman *et al.*, 1999; Chiverrell, 2001; Blundell and Barber, 2005; Langdon and Barber, 2005; Daley, 2007; Amesbury *et al.*, 2008; Blundell *et al.*, 2008) developed by Woodland *et al.* (1998), has, however, recently been replaced by a pan-European transfer function, employing sites from across Europe, and developed as part of the ACCROTELM project (Charman *et al*., 2007). This study employed the ACCROTELM transfer function, as well as a more localised model for the North of Ireland (Swindles *et al.*, 2009). For the ACCROTELM transfer function, Charman *et al*. (2007) developed a training set in which all samples from the eight well-dispersed sites were taken during a single season, thus minimising any interpretive problems presented by seasonal variation in WTD, as discussed by Booth (2008).

Likewise, Swindles *et al*. (2009) took all water table measurements in the early summer, but maintain that smaller regional training sets, such as the one developed here, can suffice for the development of transfer functions, providing no-analogue situations are not encountered.

Increasing development and application of transfer functions has been facilitated by wider understanding of the statistical techniques involved in their formation (ter Braak and Juggins, 1993; Birks, 1995; Charman *et al*., 2007), together with the development of relevant computer programmes, such as C^2 (Juggins, 2007).

Within the transfer function development process, a number of statistical models are available, which follow either linear (e.g. partial least squares regression, PLS) or unimodal (e.g. weighted average regression, WA; weighted average partial least squares regression, WA-PLS; weighted average tolerance downweighted, WA-Tol; and the modern analogue technique, MAT) response models to environmental factors. The performance of each response model can be evaluated, during development, using the r^2 and root mean squared error of prediction (RMSEP) values produced. RMSEP measures the overall predictive abilities of the relevant training set, as derived from bootstrapping. A more extensive discussion and explanation of bootstrapping can be found in Birks *et al*. (1990) and Birks (1995).

Telford and Birks (2011a) state that as many transfer function training sets are 'samples-of-convenience' and, consequently, do not represent a uniform distribution of the environmental gradient in question, it would be intuitive to suggest that such uneven sampling would reduce the efficacy of any subsequent reconstructions. However, Ginn *et al*. (2007) found that RMSEP values remained very similar between two training sets, one of which possessed an evenly spaced sampling regime along the environmental gradient, and the other with certain portions of the sampled gradient from the dataset. Telford and Birks (2011a), therefore, propose a 'segment-wise' approach to RMSEP, whereby the environmental gradient in question is divided into segments and the RMSEP for each segment is calculated. If the root mean square of the segment-wise RMSEPs are taken, then a more accurate RMSEP can be presented, representing any uneven sampling along an environmental gradient, and compared to the original RMSEP. However, this technique was developed after the transfer functions that will be used in this study (Charman *et al*., 2007; Swindles *et al*., 2009) were developed and therefore, these have not been subjected to such tests. Amesbury *et al*. (submitted) are, to the author's knowledge, the first to statistically validate a transfer function training set, by submitting it to the tests proposed by Telford and Birks' (2005, 2009, 2011a, b) recent critiques.

In addition, maximum bias, which quantifies the tendency of each model to over- or underestimate at some point along the gradient, can be used to assess the efficacy of transfer functions (ter Braak and Juggins, 1993). It is calculated by dividing the environmental gradient into a number of segments, and then calculating the mean residual for each segment. The largest value is taken as the maximum bias. Average bias, another measure of transfer function efficacy, can be taken as an average of these segments.

Of the models chosen for use in this study: during the development of the ACCROTELM transfer function, Charman *et al*. (2007) evaluated the performance of five response models, finding that WA-PLS performed most effectively (RMSEP = 7.64cm; r^2 = 0.6); and, in the development of the North of Ireland transfer function, Swindles *et al*. (2009) favoured WA-Tol (RMSEP = 4.99cm; r^2 = 0.83). Recently developed transfer functions have favoured a variety of response models, including WA (Woodland *et al*., 1998), WA-Tol (Wilmshurst *et al*., 2003), PLS (Ooms *et al*., 2011), MAT (performed well but not adopted by Charman *et al*., 2007), as well as WA-PLS (Charman, 1997).

Conversion of testate amoebae data to depth to water table estimates was conducted by Professor Dan Charman at the University of Exeter, using the ACCROTELM transfer function (Charman *et al*., 2007), and by Dr Graeme Swindles at the University of Leeds, using the North of Ireland transfer function (Swindles *et al*., 2009). In both instances, sample-specific errors of prediction (reconstruction errors) were generated through 1000 bootstrap cycles (Birks *et al*., 1990; Line *et al*., 1994).

Swindles *et al*. (2009) found that the North of Ireland transfer function performed and predicted in a similar fashion to the ACCROTELM model. It is subsequently argued that this serves to reinforce conclusions made about the ACCROTELM model and the quantitative technique itself. As a result, Swindles *et al*. (2009) adopt the ACCROTELM reconstructions, as this enables robust comparisons with recent and future records, which use this model. Given the location of the sites in this study, an opportunity presents itself whereby both transfer functions can be applied to the data in an effort to test the North of Ireland model against the ACCROTELM model with new region-specific datasets. It is expected, however, that ACCROTELM-based reconstructions will be used for future comparison with other European records.

5.4.3.3 Dupont hydrological index

A version of the DHI technique, as applied to plant macrofossil data in this study (see Section 5.4.1.2), was modified for application to testate amoebae data. The potential benefits of the DHI technique over DCA have already been outlined (see Section 5.4.1.2). A criticism of DHI in plant macrofossil analysis is that it does not accurately represent the relative variation in WTD; for example, all dry indicator species are generally confined to 7 and 8 on the index. In this modification, sample scores were calculated through the weighted averaging of species values, assigned using an estimation of their average associated WTD (Tab. 5.2), as presented by Charman *et al*. (2007), rather than on a ordinal scale. By adopting average WTD it was hoped that relative differences in WTD would be more accurately represented. Three species, *Arcella artocrea*, *Nebela collaris* and *Centropyxis ecornis* were not included in the Charman *et al*. (2007) study, and owing to their low relative concentrations in this study, could be excluded from DHI analysis without any significant repercussions for the results. Analysis followed the same procedure outlined in Section 5.4.1.2.

Table 5.2 Modified Dupont hydrological index species weightings for testate amoebae.

Testate amoebae taxon	**Weighting (average depth to water table; cm)**
Difflugia lucida	2
Nebela carinata	2.5
Amphitrema wrightianum	3
Arcella discoides	4
Centropyxis aculeata type	4.5
Heleopera petricola	6.5
Heleopera sphagni	6.5
Amphitrema flavum	7.5
Difflugia pulex	8.5
Cyclopyxis arcelloides	10.5
Hyalosphenia papilio	10.5
Nebela flabellum	11.5
Difflugia pristis type	12
Corythion-Trinema type	17
Heleopera sylvatica	17
Hyalosphenia subflava	17.5
Nebela militaris	17.5
Assulina muscorum	18.5
Assulina seminulum	18.5
Nebela tincta	19.5
Bullinularia indica	21
Trigonopyxis arcula	23.5
Arcella catinus	27
Arcella artocrea	Excluded
Nebela collaris	Excluded
Centropyxis ecornis	Excluded

5.4.4 Normalisation of records in multi-proxy studies

Intercomparison and critical assessment of the proxy and isotopic records in relation to one another can be achieved by normalising all curves and placing them on the same axis (Charman *et al.*, 1999). The following equation was used to normalise the data:

$$z = \frac{X - \mu}{\sigma}$$

where: z = normalised value, X = raw value, μ = mean of all X, σ = standard deviation of X.

Chapter 6 - Chronology

6.1 Introduction

This chapter provides a review of the chronological techniques employed in this study, a rationale and strategy for chronological construction, an outline of the laboratory and statistical methodologies associated with radiocarbon dating and tephra analysis, a presentation of chronological results, and details of the development of age-depth models

6.2 Chronological strategy

Chronological control on sedimentary sequences intended for palaeoclimatic analysis is vital. The construction of robust age-depth models, and the identification of time-parallel marker horizons, allows the timing of any variability to be more accurately identified, described and interpreted. Furthermore, this practice facilitates the correlation of palaeoclimatic records from one site, or one archive, to another. Within this study, chronological strategy also aims to provide high-resolution chronological control over the time period of interest, encompassing any possible palaeoclimatic signal manifested by the 4.2 kyr event.

Smart and Frances (1991) outline a series of criteria, which should be considered during the selection of a dating technique, namely: applicability of material for dating, time range, and the precision and accuracy of the technique. Assuming the first criterion can be met, the latter two must be considered in terms of whether the research objectives of the project itself, and whether they can be addressed using the technique in question. Consequently, radiocarbon dating was selected as the primary method of providing chronological control to the records in this study. In addition, tephrochronology was also used to provide additional chronological control, as well as an explicit time marker, with which records could be correlated.

6.3 Radiocarbon dating

6.3.1 Basic principles

Atmospheric radiocarbon (^{14}C) is absorbed into the tissue of all living organisms. Upon death, this absorption halts and ^{14}C begins to decay, with a half-life of 5568 ± 30 years (the 'Libby half-life'; internationally accepted) or 5730 ± 40 years (the 'Cambridge half-life'; best estimate; Godwin, 1962). Whilst the latter may be more accurate, radiocarbon laboratories continue to use the Libby half-life to maintain comparability with earlier studies. However, calibration of radiocarbon dates effectively cancels out the half-life, providing the same value is used throughout, and thus negates any problems presented by the inaccuracy of the Libby half-life. The fundamentals of radiocarbon dating are discussed in detail by Olsson (1986), Pilcher (1991; 2003), Lowe and Walker (1997) and Bronk Ramsey (2008a).

6.3.2 Radiometric and accelerator mass spectrometry dating

Historically, peat-based studies have employed radiometric radiocarbon dating, whereby a sample's radioactive emissions are measured over a set period of time. Because of the amount of material required for this analysis (c. 5-10 g carbon), the standard sampling approach was to use 'bulk dates', which incorporated anywhere between 1 to 8 cm bulked peat material. Clearly, this reduces the potential dating resolution for a given sequence significantly, as well as reducing the amount of material available for palaeoecological or palaeoenvironmental analysis.

More recent peat-based studies have, instead, employed the accelerator mass spectrometry (AMS) technique, whereby the actual number of ^{14}C atoms in a sample is measured. As a result, the AMS technique requires much smaller sample sizes (c. 1-10 mg carbon), enabling much higher dating resolution, and significantly reducing the likely errors in chronologies.

An additional advantage of the reduction in sample size is that it is now feasible to isolate and date individual plant remains (e.g. seeds, wood fragments) and specific structural components of a plant (e.g. *Sphagnum* stems, *Calluna vulgaris* stems), reducing the chance of contamination. The latter is particularly useful as it guarantees the dating of aboveground material (i.e. stems), which died *in situ*, and is therefore representative of the depth sampled (Killian *et al.*, 1995, 2000). Nilsson *et al.* (2001) confirmed that ^{14}C ages from mosses were most accurate and reliable, and that no significant difference existed in the average ^{14}C content of various *Sphagnum* species.

Killian *et al.* (1995, 2000) identified a 'reservoir effect', whereby an uptake of old methane is hypothesised to have caused the artificial aging of *Sphagnum* samples by 100-150 years. Blaauw *et al.* (2004b), however, find no evidence of this effect in either isolated macrofossil or bulked samples. Blaauw *et al.* (2004a) also note that a similar effect is experienced when *Sphagnum* remains are not thoroughly cleaned of ericaceous root materials and fungal remains (cf. Killian *et al.*, 1995, 2000; Speranza *et al.*, 2000), suggesting that this may be the cause of the 'reservoir effect'. Whilst the AMS technique may reduce statistical precision slightly, the advantages of the modern method, in terms of sample size required and the subsequent potential dating resolution, allow it to remain preferential to the conventional radiometric method (Törnqvist *et al.*, 1992; Oldfield *et al.*, 1997; Mauquoy *et al.*, 2004b).

6.3.3 Radiocarbon age calibration

As atmospheric ^{14}C has not remained constant over time, ^{14}C years are not equal to calendar years (van der Plicht, 2004). Dating of known age dendrochronological samples (e.g. Pearson *et al.*, 1986) demonstrated that variations in atmospheric ^{14}C are such that ^{14}C ages are typically younger than their corresponding, correct calendar age. To rectify this, ^{14}C ages must be calibrated using calibration curves, such as the IntCal curve series (Stuiver *et al.*, 1998; Reimer *et al.*, 2002; Reimer *et al.*, 2009) and the Fairbanks curve (Fairbanks *et al.*, 2005) developed from annually resolved records of atmospheric ^{14}C, including tree-rings, corals and laminated marine sediments. The development of calibration software (e.g. CALIB v. 6.1.1, Stuiver *et al.*, 2005; *OxCal* v.4.1, Bronk

Ramsey, 2009) has enabled ^{14}C ages to be converted to calendar ages quickly and easily. Bartlein *et al*. (1995) emphasise the importance of calibrating radiocarbon dates in order to aid the formulation meaningful and useful age-depth models.

There are, however, issues associated with the calibration process, which must be considered during the construction of age-models. Firstly, the calibration of ^{14}C ages produces non-Gaussian, and often polymodal, probability distributions on the calendar age scale. Errors associated with the ^{14}C age, derived during sampling, preparation and measurement, combined with variations in the profile of the calibration curve, caused by past variations in atmospheric ^{14}C, can result in the calibrated probability distributions spanning several hundred years.

Particular difficulties arise when dealing with certain areas of the calibration curve, where a horizontal flattening of the curve occurs, often known as a 'radiocarbon plateau', resulting in notably large, and often complicated, calibrated probability distributions (Fig. 6.1a). Here, errors attached to calibrated dates can subsequently overlap, leading to difficulty when interpreting or distinguishing the calibrated age estimates. Baillie (1991) describes how this can sometimes lead authors to 'suck in and smear' events together, potentially leading to false conclusions of event synchronicity. Attempts are, however, being made to address this problem statistically (e.g. Blaauw *et al*., 2007a). Conversely, near vertical periods in the calibration curve can produce extremely precise calibrated probability distributions (Fig. 6.1b). However, it is precisely these sections of variation, which are of greatest benefit to the practice of WMD (e.g. van Geel *et al*., 1998 Mauquoy *et al*., 2004b; see Section 6.3.5.2). Figure 6.2 also demonstrates that no significant radiocarbon plateaus occur during the period encompassing the 4.2 kyr event, apart from a short plateau 4400 – 4300 cal yr BP.

6.3.4 Mid-points

An uncalibrated radiocarbon age possesses errors, which are normally distributed around a mean. However, after calibration, errors can be highly non-normal and multimodal, occasionally with non-continuous probability fields due to the shape of the calibration curve. It is worth noting that the 'true' calibration curve is actually a smooth continuous function (Gómez Portugal Aguilar *et al*., 2002). However, the estimated curves (e.g. Stuiver *et al*., 1998, Reimer *et al*., 2002; Reimer *et al*., 2009) with which ^{14}C ages are calibrated is has a more irregular and saw-toothed shape. This complex probability distribution prevents the generation of a single, summary value. The discussion and development of chronologies in palaeoenvironmental studies often necessitates the calculation of a point estimate. There are a number of ways to generate point estimates of a calibrated radiocarbon age, of which the intercept method is the most widely adopted (Telford *et al*., 2006).

Telford *et al*. (2006) found that intercepts and modal averages were sensitive to small changes in ^{14}C age and subsequently possessed higher median differences between adjacent ^{14}C ages. Conversely, weighted and median averages exhibited more stable behaviour and could therefore be considered more robust estimates, based on the assertion that a point estimate should approximately mimic the change seen in the ^{14}C age. Two hypothetical ^{14}C ages (4530 ± 50 and 4540 ± 50 ^{14}C years) were used to demonstrate that a shift of just 10 ^{14}C years resulted in a difference of 138 cal years, when using the intercept method to calculate a single point estimate, but just 8 cal years when the weighted averaging method was adopted.

Figure 6.1 Variation in radiocarbon calibration precision resulting from differences in the calibration curve between a) plateaus and b) steep sections. Calibrations conducted in *OxCal* v.4.1 (Bronk Ramsey, 2009), using the IntCal09 calibration curve (Reimer *et al.*, 2009).

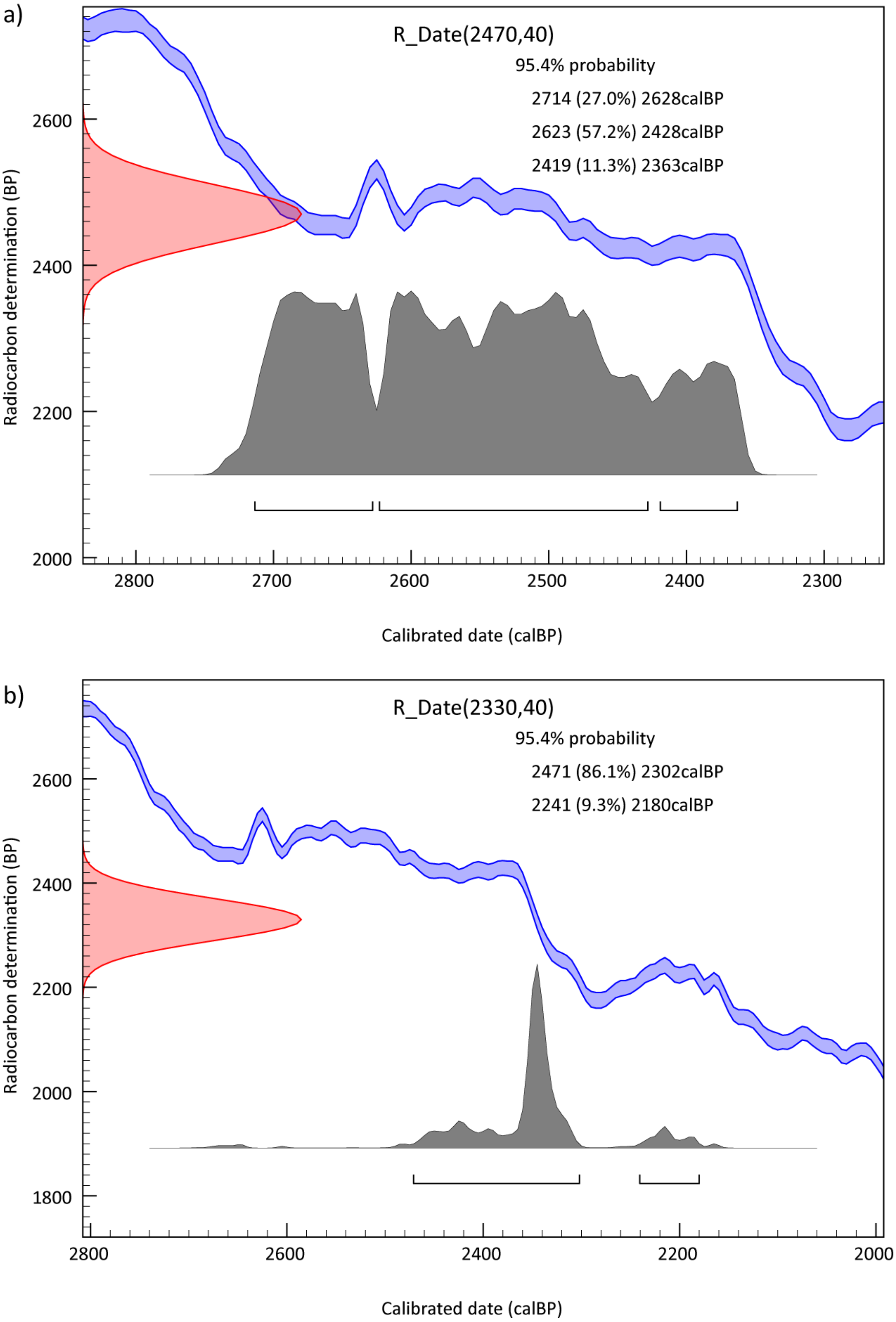
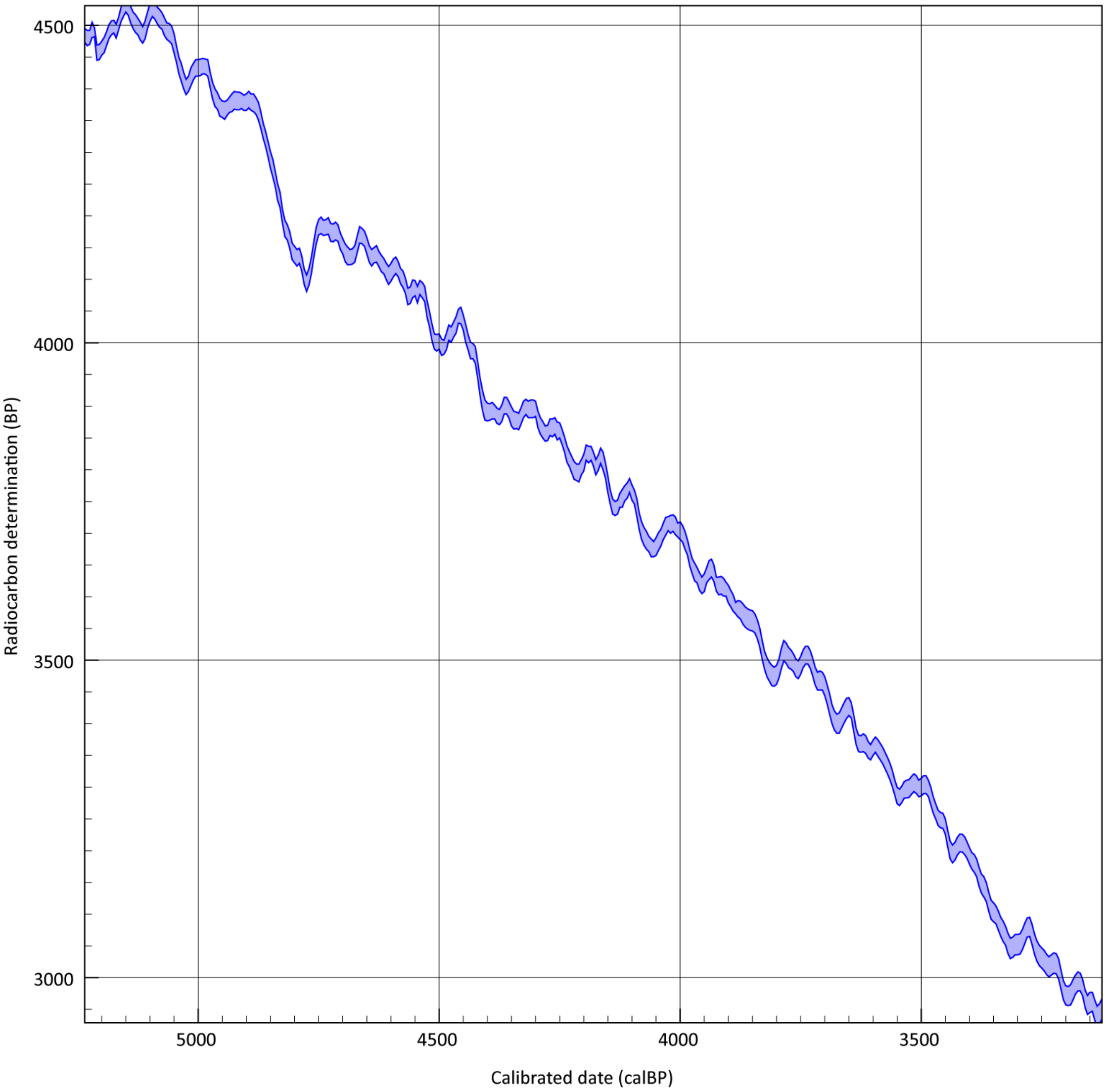


Figure 6.2 The IntCal09 ^{14}C calibration curve c. 5200 – 3200 cal yr BP.



Michczyński (2007) compared the various methods of producing a single point estimate with the true value of the calendar age, using computer simulations, finding that no method may be considered as a good estimate. Telford *et al.* (2004) suggest that whilst no single function can adequately express the probability distribution of a calibrated radiocarbon date, the weighted average or median should be adopted where a single point estimate is required. Where possible, however, it is emphasised that a full distribution should be favoured. This is especially true during the development of age depth models (Blaauw, 2010).

6.3.5 Age-depth models

Parnell *et al.* (2011, p. 2949) succinctly describe the overarching aim constructing chronologies, via the development of age-depth models, within the context of palaeoenvironmental studies, stating that the task is to "*use scientific dating and depth information to produce a chronology which can be used to estimate the sedimentation rate and thus offer an estimate for the date at any given depth in the core along with a carefully quantified statement of uncertainty*".

The modelling of accumulation of palaeoenvironmental sequences through time has, subsequently, been attempted with a range of statistical techniques. In an analysis of the literature, Blaauw (2010) found that linear interpolation models were most frequently employed (31 times), followed by linear regression (13), Bayesian techniques (11) and 6 other model types. Model choices were not specified by 18 studies. Bennett and Fuller (2002) emphasise the importance of choosing the right type of model, as they are likely represent the single greatest source of error when determining the age of a given point in the sequence.

6.3.5.1 'Classical' age-depth models

Bennett (1994) and Telford *et al*. (2004) review the various methods, which could be considered 'traditional', or 'classical'. Linear regression, whereby a line of global fit is calculated and applied to the chronology, is inherently problematic, given that a change at either end of the core can affect the entire chronology (Telford *et al*., 2004). However, linear regression can be used to quickly, if a little crudely, to assess approximate accumulation records in sequences, which demonstrates broad uniformity in their rate of accumulation (e.g. a raised peat bog).

The use of linear interpolation, whereby a line of best fit is forced to pass through each radiocarbon age, with straight lines connecting each of the mid-points, also presents a number of problems. Firstly, it assumes abrupt changes in the rate of deposition at each dated depth. Whilst this assumption is often unrealistic, the technique is often capable of producing what appear to be conceivable age-depth models (Bennett, 1994). However, linear interpolation models do not take into account the errors associated with radiocarbon dates, with the gradient simply running through the mid-point of each radiocarbon age (Bennett, 1994). Parnell *et al*. (2011) observe that such models do not allow for flexibility of behaviour, or error within the model, as well as providing no strategy for the assessment and identification of 'outliers', or ages that do not conform to the monotonic model of ages increasing with depth.

Spline interpolation, whereby each of the radiocarbon ages are connected by a polynomial, usually cubic, fitted between pairs of points to created a spline, can avoid abrupt changes in accumulation rate at each age due to its employment of non-local information (i.e. consideration of the other dates in the sequences), thus resulting in a smoother curve overall. Again, however, this technique does not take into consideration the associated errors of each radiocarbon age (Bennett, 1994). Furthermore, as both linear and spline interpolation force the age-depth model to pass through each radiocarbon age, they inherently introduce noise to the model, as each point estimate is unlikely to be the true age of the sample dated (Telford *et al*., 2004).

Alternatively, the fitting of polynomial curves to a series of dates aims to find a line, which is a reasonable model of those ages and, conversely, is not required to pass through the age points. This acknowledges that radiocarbon ages are only statistical estimates of the true age of sample (Bennett, 1994). Polynomials should be fitted, which have the least terms possible, whilst maintaining a good level of fit. However, Bennett and Fuller (2002) describe the difficulties involved, whereby polynomials with too few terms can perform too rigidly, whilst polynomials with too many terms can lack constraint.

Various conclusions have been reached as to the most effective age-model to use, but much depends on the nature of the individual sequence involved, and the number of radiocarbon ages available. For example, Telford *et al*. (2004) found that, when large numbers of dates were available, a cubic spline model performed best when compared to a known age sequence. It is acknowledged, however, that the number of dates required (≥ 24 over a Holocene length record) were rarely available to 'typical' studies. Alternatively, Bennett and Fuller (2002) found that linear interpolation models, whilst never 'correct', were often the most reliable.

However, all the 'classical' techniques discussed here are underpinned by a number of fundamental shortcomings. Firstly, key papers that discuss the development of age-depth models using such techniques, including Bennett (1994) and Bennett and Fuller (2002), base their analyses on uncalibrated radiocarbon dates. The importance of calibrating radiocarbon dates, as first stated by Bartlein *et al*. (1995), has already been discussed. Whilst it is true, as Bennett (1994) claims, that calibrating radiocarbon dates adds additional complexity to the development of age-depth models, due to the subsequent non-Gaussian probability distributions, it remains vital that calibration is performed. Telford *et al*. (2004) state that models based on uncalibrated dates make the implicit and implausible assumption that variations in sedimentation rate are capable of cancelling out the variations in the calibration curve. Furthermore, software programmes, which have often been used in the past to develop 'classical' age-depth models, such as *Tilia* (Grimm, 1990) and *Psimpoll* (Bennett, 2009), generally do not use calibrated dates or take into account the associated errors.

Secondly, Telford *et al*. (2004), although calibrating radiocarbon ages first, still use point estimates for the model. Since the above techniques do not take into consideration the errors associated with radiocarbon ages, they must, inevitably, adopt a single point estimate summary of each age. Telford *et al*. (2004) acknowledge that whilst the use of a central point estimate is common practice, it is not entirely appropriate as the asymmetric, multimodal probability distribution produced by calibration cannot be meaningfully characterised by a single value (see Section 6.3.4, Telford *et al*., 2006). Blaauw (2010) agrees, by indicating that error ranges of all point estimates (as outlined by Telford *et al*., 2006) will be asymmetric, and that therefore 'classical' techniques which depend on symmetric error distributions, such as those described above, should not be applied to ^{14}C age-depth modelling. The mid-point is an exception, but cannot robustly be employed as it assumes a symmetrical probability distribution, which is unrealistic (Nilsson *et al*., 2001). Rather, Blaauw (2010) proposes a new approach to 'classical' age-depth modelling, whereby the entire calibrated distribution of each date is employed in model creation (see Section 6.3.5.4), thus addressing the primary criticisms of these models (Telford *et al*., 2004, 2006; Parnell *et al*., 2011).

6.3.5.2 Wiggle-matched age-depth models

The WMD approach to the construction of age-depth models also employs the full probability distribution of radiocarbon dates. The technique, first used by Pearson (1986) and van Geel and

Mook (1989), matches a series of closely spaced ^{14}C ages to the variations in the profile of the calibration curve (i.e. 'wiggles') using a linear regression model. The linear regression model will, subsequently, give an estimated calibrated date for every depth in the core, including those with radiocarbon estimations. These can then be overlain on top of the calibration curve to get an estimate of the strength of fit of the proposed model. A more satisfactory chronology can be achieved by optimising the linear regression model, providing a closer fit between calibrated dates and the calibration curve (Blaauw *et al.*, 2003; Parnell *et al.*, 2011). Put simply, the dates are adjusted on the calendar timescale until a position of best fit on the calibration curve is found.

Blaauw *et al.* (2003) extended the technique to use separate linear regressions for different parts of the core where changes in sedimentation rate are hypothesised. Whilst providing more chronological flexibility, this technique is criticised by Parnell *et al.* (2011) for not yielding sufficient variable uncertainty, and not matching uncertainty to the shape of the calibrated radiocarbon date distributions. Parnell *et al.* (2011) also warn against making assumptions of dramatic chronological variation over a short depth period, without strong *a priori* evidence.

Ideally, when using WMD the time period involved must be long enough to encompass more than one wiggle, and preferably include a number of pronounced wiggles or steep gradients. Precision obtained during such periods can be extremely high (Blaauw *et al.*, 2004), and have subsequently been the focus of a number of high-resolution WMD peat-based studies (e.g. Kilian *et al.*, 1995, 2000; Speranza *et al.*, 2000; Mauquoy *et al.*, 2002b). The advent of the AMS radiocarbon dating technique, and the small sample sizes it requires has made the dating resolution required by WMD more achievable, in terms of available material.

The inherent assumption of WMD, that of a linear accumulation rate within peat sequences, has been acknowledged as a potential oversimplification by Blaauw *et al.* (2004), whilst also suggesting that other growth models also make use of their own assumptions, and that more 'sophisticated' growth models are not necessary. This is especially true over short intervals, where the assumption of linear accumulation has resulted in highly satisfactory WMD age-depth models (Blaauw *et al.*, 2004).

Despite the obvious benefits of WMD, it remains an extremely time consuming and expensive technique (Mauquoy *et al.*, 2004b). As a result, many WMD studies focus on relatively short periods of time (e.g. Little Ice Age, Mauquoy *et al.*, 2002a, b; 2800 yr BP 'event', van Geel *et al.*, 1998; Speranza *et al.*, 2000). Furthermore, the number of dates required for effective WMD is so large that it is prohibitive for most 'typical' peat-based studies, as Telford *et al.* (2004a) acknowledge, and sequences with coarsely spaced ^{14}C ages should be modelled by investigating a range of other methods (Yeloff *et al.*, 2006).

6.3.5.3 Bayesian age-depth models

It is possible to reduce the possible calendar age intervals by applying Bayesian statistics (Buck *et al.*, 1996; Blaauw *et al.*, 2003). This technique applies a framework that systematically combines data and prior information. In most raised bog environments, for example, a considerable amount of prior information is known, for example: the surface of the bog is equal to or older than the year of sampling; the bog has accumulated continuously over time (often with an accumulation rate c. 10 – 20 years/cm); the bog is unlikely to be older than the Holocene. Subsequently, in Bayesian frameworks, this information is systematically combined with the ^{14}C data to construct age-depth models, employing sophisticated iterative sampling methods, such as Monte Carlo, or bootstrap sampling, with formal measures of their fit (Yeloff *et al*., 2006). A number of programmes have been developed for the construction of Bayesian age-depth modelling including: *BChron* (Haslett and Parnell, 2008), *Bpeat* (Blaauw and Christen, 2005), *BCal* (Buck *et al*., 1999), and *OxCal* (Bronk-Ramsey, 2009); and are reviewed in detail by Parnell *et al*. (2011) and Yeloff *et al*. (2006). In addition, an excellent review of the role of Bayesian statistics in the development of radiocarbon calibration curves is offered by Blackwell and Buck (2008).

The application of Bayesian statistics clearly has immense potential for the development of high-resolution, precise age-depth models in peat-based studies (e.g. Blaauw and Christen, 2005; Bronk Ramsey, 2008; Haslett and Parnell, 2008), and their use should be considered the preferred option for the development of future age-depth models (Blockley *et al.*, 2007). However, as Blaauw (2010) observes, existing Bayesian age-depth modelling methods are unlikely to add much extra value to sequences from sites with lower dating resolution.

6.3.5.4 Age-depth model development methods

The sheer number of dates required by the Bayesian and wiggle-matched approaches, to yield meaningful and useful results, or to provide chronological constraint greater than that typically associated with 'classical' age depth modelling, is beyond the means of this project (Yeloff *et al*., 2006; Blaauw, 2010). In fact, for many studies, after consideration of the time available for analyses and the likely funding available to each project, WMD and Bayesian techniques can often be deemed unsuitable (cf. Telford *et al*., 2004).

Blaauw (2010) presents new software, which allows non-Bayesian, 'classical' age-depth models, employing the entire calibrated distribution of each date, to be produced quickly, and easily. The '*Clam*' (i.e. CLassical Age Modelling) software for 'classical', non-Bayesian age-depth modelling, coded by Blaauw (2010) for the 'R' statistical package (R Development Core Team, 2011), provides an opportunity to explore the impact of different modelling choices in a transparent and systematic manner, addressing many of the criticisms of traditional or 'classical' age-depth models (see Section 6.3.5.1) and, consequently, presenting a viable alternative to age-depth modeling using Bayesian statistics.

Put simply, *Clam* calibrates ^{14}C dates individually by calculating their calibrated distributions for every calendar year. Age-depth models are subsequently constructed using these calibrated distributions, as opposed to the inappropriately assigned single point estimates (Telford *et al*., 2006). Models can be developed using linear interpolation, linear or higher order polynomial regression or cubic, smoothed or locally weighted splines, at the user's discretion. Through a process of repeated sampling of the calibrated distributions, each time calculating a new age-depth model, an estimate of the age-depth relationship is consequently built up, which is based on the dated depths, their associated uncertainties, and the chosen type of age-model.

Any ages considered outliers can be removed manually and disregarded from model development. Hiatuses can also be included in model development, if identified by either stratigraphy and/or initial estimations of age-depth relationship based on ^{14}C dates.

Based on the discussion presented here, *Clam* was used to develop the age-depth models at both sites in this study. The transparency and added robustness afforded by *Clam* to 'classical' age-depth model construction, was deemed to be of more chronological benefit to the study than the adoption of Bayesian techniques, owing to the uncertainty surrounding their efficacy for sequences with relatively fewer dates.

6.3.6 Radiocarbon dating strategy and methods

A total of twenty ^{14}C AMS radiocarbon dates were applied to the two cores in this study. At Sluggan Moss, two initial dates were processed by Beta Analytic Inc., Miami. Subsequently, six dates were processed at the NERC Radiocarbon Laboratory, East Kilbride (NERC RCF Allocation no: 1523.0910) and four dates were processed at the ^{14}CHRONO centre, Queen's University Belfast. One ^{14}C date was later excluded from age-depth model development due to a coring error. At Fallahogy Bog, three of the dates were processed at the NERC Radiocarbon Laboratory (Amesbury, 2008). Six further dates from Fallahogy were processed at the ^{14}CHRONO centre.

Bennett (1994) suggests that radiocarbon ages should be placed evenly through the sequence, in an effort to minimise errors associated with the dating of specific horizons. It is suggested that the bracketing of a horizon with radiocarbon ages can obtain smaller confidence intervals. Telford *et al*. (2004), however, claim that this approach may not be the optimal strategy, if distinct changes in sedimentation can be identified. Based on sequence stratigraphies at both sites, no significant changes in accumulation rate or hiatuses were identified, and therefore dates were evenly spread through the sequences. At Sluggan Moss, two rangefinder dates were used to clarify the age of the sediments in question, before the additional dates were added at regular intervals. At Fallahogy Bog, the existing dates from Amesbury (2008) removed the need for rangefinders. As a result, dates were evenly spaced along the sequence to encompass the period of interest. As a limited number of dates were likely to be available to the project, no effort was made to take the predicted slope of the calibration curve into consideration (cf. Reimer *et al*. 2002) when placing the dates.

Samples for AMS ^{14}C dating were prepared in a clean environment, with the utmost care taken to ensure no contaminants came in to contact with the sample (e.g. clothing, blue roll). A small amount of peat was washed through a 125 µm sieve using distilled water and the residue retained for microscopic inspection using a low powered stereo-zoom microscope at x10 magnification. The sample was examined for aboveground macrofossils, as previously discussed in Section 5.3.1. Where possible, *Sphagnum* moss was isolated due to its relative abundance, weight and ease of preparation. When *Sphagnum* was not available, aboveground ericaceous material was selected instead. In this instance, *Calluna vulgaris* wood, identified by its distinctive pattern of petioles, was isolated from the peat matrix. All ^{14}C samples were cleaned of any contaminators (e.g. ericaceous rootlets, UOM) before being stored in distilled water in stoppered glass vials and sent for analysis.

Calibration of raw ^{14}C ages was undertaken using CALIB v. 6.1.0 (Stuiver *et al.*, 2005) and the IntCal09 calibration curve (Reimer *et al.*, 2009). Median point estimates were calculated within the CALIB programme and weighted averages of the date range distribution (2σ) were taken as the midpoint of the range that accounted for the greatest proportion of relative probability. Full, calibrated probability distributions were used in the construction of age-depth models (see Section 6.3.5.4), however, after the recommendations of Telford *et al.* (2006) and Michczyński (2007).

6.4 Tephrochronology

Chronologies associated with peat-based palaeoclimatic research are dominated by radiocarbon dating. However, tephra analysis of these sequences can provide additional chronological constraint to age-depth models, as well a fixed-age correlative points that enable comparison between sites and sequences, and should be included, where possible, in age-depth model construction (Blaauw, 2010).

6.4.1 Basic principles

In the strictest sense, 'tephra' refers to all unconsolidated pyroclastic material formed as a volcano erupts of varying grain sizes. These include ash (<2 mm), lapillus or lapilli (2–64 mm), and blocks (angular) or bombs (sub-rounded) (>64 mm). For the purposes of this study, the term 'tephra' will be used to describe solely the microscopic ash component. A full discussion of tephra and tephrochronological-related terminology and nomenclature can be found in Lowe (2008; 2011). Furthermore, a detailed review of the use of tephrochronology in peat cores can be found in Gehrels *et al.* (2008) and Swindles *et al.* (2010b).

Despite the relative abundance of dateable tephra layers, and palaeoecological and palaeoclimatic studies in the northwest Europe, tephrochronology was underemployed as a chronological tool (Buckland *et al.*, 1981), until relatively recently. Despite the completion of early work in Scandinavia (e.g. Persson, 1971), it was, perhaps, not until the first discovery of an Icelandic tephra in Great Britain and Ireland (Dugmore, 1989) that the practice of constructing tephrochronologies became more widespread. Currently, detailed regional tephrochronologies exist across northwestern Europe, including: northern Ireland (Pilcher and Hall, 1992; Pilcher *et al.*, 1995a, 1996), northern England (Pilcher and Hall, 1996), northern Germany (van den Bogaard *et al.*, 1994, 2002; van den Bogaard and Schmincke, 2002) and Sweden (Boygle, 2004; Wastegård, 2005). Such tephrochronologies are now frequently employed to constrain and refine age-depth models within palaeoecological and palaeoclimatic studies (e.g. Eiríksson *et al.*, 2000; Langdon and Barber, 2005; Borgmark and Wastegård, 2005; Hall and Mauquoy, 2005; Plunkett, 2006; Vorren *et al.*, 2007; Barber *et al.*, 2008). Ombrotrophic raised bogs often lack a significant mineral component, with only wind-blown dust present and an organic content approaching 100% (van den Bogaard and Schmincke, 2002). These sequences are an ideal medium from which to extract tephra shards.

The application of tephrochronology in a peat-based context relies on assumptions that each tephra horizon is: rapidly dispersed over a wide area and subject to geologically instantaneous deposition; in possession of a distinctive geochemical signature (Hunt and Hill, 1993); not mobile within the peat profile (Langdon and Barber, 2004). If these assumptions are considered reasonable, tephra layers can potentially be used to precisely correlate regional peat sequences. Payne and Gehrels (2010) found that, in laboratory experiments, tephra was mobile within the peat profile, with shards being found up to 15cm below the surface after 6 years simulated rainfall. However, the vast majority of the material remained on the bog surface, and subsequently, should not pose a chronological problem providing the peak of tephra concentration is taken as being representative of the horizon.

Tephra horizons can be employed for correlation and to provide absolute ages in age-depth models. It is possible to date historical tephras to the individual calendar year in which the eruption took place (e.g. Hekla AD 1510, Dugmore *et al.*, 1995a, 1996; Larsen *et al.*, 1999). However, the majority of horizons utilised in Holocene peat-based climate research are prehistoric and dated by radiocarbon at multiple sites (e.g. Pilcher *et al.*, 1995a; Dugmore *et al.*, 1995b; Buck *et al.*, 2003; Plunkett *et al.*, 2004). Studies conducted over longer timescales and in varying sedimentary environments have employed other techniques to date tephra horizons (e.g. thermoluminescence dating, Berger and Busacca, 1995; fission-track dating, Alloway *et al.*, 1993; ^{40}Ar-^{39}Ar dating, Ton-That *et al.*, 2001). More recently, significant progress has been made dating tephra horizons within ice cores, as volcanic shards are extracted from ice layers of a known age (Vinther *et al.*, 2006; Davies *et al.*, 2010, 2012; Abbot *et al.*, 2012).

The peatlands of northwestern Europe have a particularly rich potential for the employment of tephrochronology, owing to their relative proximity to Iceland. The Icelandic volcanic systems are largely dominant throughout the Holocene in northwest Europe, although eruptions from smaller volcanic fields, such as those in Jan Mayen Island (Chambers *et al.*, 2004) and the Eifel region, Germany (van den Bogaard and Schmincke, 1985; Zolitschka *et al.*, 1994) are also evident. Other European volcanic fields exist, but tephra produced by such eruptions tends not to disperse as far as northwestern Europe (e.g. Vesuvius, Campanian volcanic zone, Rolandi *et al.*, 2008; Santorini, Cycladic massif, Watkins *et al.*, 1978, Eastwood *et al.*, 1999).

However, the technique of tephrochronology is not without its limitations. Several factors can reduce the correlative power of a given tephra layer, and also increase the potential for its misattribution. Firstly, not all tephras are distinguishable based on their geochemistry alone, although their relative stratigraphic or chronological context can usually be used to differentiate them (e.g. Hekla 4 and Lairg A, Hall and Pilcher, 2002). Secondly, the area of dispersal and deposition varies between eruptions, even of those that emanate from a similar source (Davies *et al.*, 2010). As a result, there can be no guarantee that a particular tephra will be present at all sites, even on a local scale. Thirdly, Dugmore and Newman (1992) demonstrate that tephra layers do not necessarily form continuous, horizontal and well-defined horizons, especially when deposited over a short period of time. As a result, it cannot be guaranteed that two cores from the same site will produce the same tephra horizons.

6.4.2 Site tephrochronology

Tephrochronology can provide valuable additional age control to radiocarbon age-depth models, and will, therefore, be included in this study to provide as robust a chronology as possible at each site. Complete tephrochronologies exist for both Sluggan Moss and Fallahogy Bog (Pilcher and Hall, 1992; Pilcher *et al.*, 1995a, 1996), and are summarised in Table 6.1.

Table 6.1 Existing tephrochronologies from Sluggan Moss and Fallahogy Bog, northern Ireland.

Site	Tephra horizon	Age (cay yr BP)	Age (AD/BCE)	Age reference	Site reference
Sluggan Moss	Hekla AD 1510	440	AD 1510	Historical	Pilcher *et al.* (1996)
	Oraefajokull AD 1362	588	AD 1362	Historical	Pilcher *et al.* (1995a, 1996)
	Hekla 1	846	AD 1104	Historical	Pilcher *et al.* (1995a, 1996)
	AD 860	1174 – 1063	AD 776 – 887	Pilcher *et al.* (1996)	Pilcher *et al.* (1995a, 1996)
	Hekla 3	3143 – 2855	905 – 1193 BCE	Dugmore *et al.* (1995a)	Pilcher and Hall (1992)
	Hekla 4	4229 – 4345	2395 – 2279 BCE	Pilcher *et al.* (1996)	Pilcher and Hall (1992); Pilcher *et al.* (1995a, 1996) Hall *et al.* (1994a, 1994b)
	Lairg B	6564 – 6728	4778 – 4614 BCE	Pilcher *et al.* (1996)	Pilcher *et al.* (1996)
	Lairg A	6809 - 6998	5048 – 4859 BCE	Pilcher *et al.* (1996)	Pilcher *et al.* (1996)
Fallahogy Bog	AD 860	1174 – 1063	AD 776 – 887	Pilcher *et al.* (1996)	Barber *et al.* (2000)
	Hekla 3	3143 - 2855	905 – 1193 BCE	Dugmore *et al.* (1995a)	Pilcher and Hall (1992)
	Hekla 4	4229 – 4345	2395 – 2279 BCE	Pilcher *et al.* (1996)	Pilcher and Hall (1992); Pilcher *et al.* (1996)
	Lairg B	6564 – 6728	4778 – 4614 BCE	Pilcher *et al.* (1996)	Pilcher *et al.* (1996); Hall (2003)
	Lairg A	6809 - 6998	5048 – 4859 BCE	Pilcher *et al.* (1996)	Pilcher *et al.* (1996); Hall (2003)

Hekla 4, as well as being the first Icelandic tephra to be discovered (Dugmore, 1989) and geochemically identified (Dugmore *et al.*, 1992) in Great Britain and Ireland, is also one of the most widely distributed, with presence confirmed at sites in Scotland (Dugmore *et al*, 1989, 1992, 1995a; Bennett *et al.*, 1992; Blackford *et al.*, 1992; Dugmore and Newton, 1992; Charman *et al.*, 1995; Langdon and Barber, 2004), Ireland (Pilcher and Hall, 1992; Pilcher *et al.*, 1995a, 1996; Hall *et al.*, 1994a, 1994b; Plunkett *et al.*, 2004) and England (Pilcher and Hall, 1996; Wells *et al.*, 1997; Amesbury, 2008). Consequently, the horizon has been extensively dated. Scott *et al.* (1992) attempted to date a peat sample from below the Hekla 4 tephra layer in Iceland, with upwards of a hundred ^{14}C measurements being taken. However, the dates fell on a short-lived radiocarbon plateau c. 4400 – 4300 cal yr BP (see Section 6.3.3) and so no number of dates could produce a precise age estimate. Subsequently, the most precise age estimate was achieved by exploiting this variation in the calibration curve through wiggle match dating (see Section 6.3.5.2), providing a date of 4287 ± 58 cal yr BP (Pilcher *et al.*, 1996).

The likelihood of identifying Hekla 4 at both sites enabled tephra analysis to serve an additional purpose in this study, as it is well placed to act as time marker horizon around which to focus a palaeohydrological examination into the nature of the 4.2 kyr event, prior to radiocarbon age-depth model development. Hekla 4 horizons found so far have not been contaminated by other tephra layers (Pilcher *et al.*, 1996). Extensive analyses of the horizon's geochemical signature have been published using samples from the type-site in Iceland (Dugmore *et al.*, 1992), as well as sites in Scotland (Dugmore *et al.*, 1992) and Ireland (Pilcher *et al.*, 1996). Although it does possess a similar geochemical composition to Lairg A, chronostratigraphic differences allow the two to be clearly distinguished (Hall and Pilcher, 2002).

Blackford *et al.* (1992) demonstrate a temporal link between Hekla 4 and a decline in *Pinus* pollen in northern Scotland, hypothesising that the demise of the taxa in the climatically marginal region may have been caused by either the direct products of the eruption, or by secondary environmental effects. Hall *et al.* (1994b) refuted this causal link, however, finding that the decline in *Pinus* pollen and the Hekla 4 horizon were clearly separated in time at sites in Ireland, including Sluggan Moss.

Whilst this issue remains contentious (Edwards *et al.*, 1996; Hall *et al.*, 1996), the broad temporal association of the Hekla 4 tephra and the well-documented mid-Holocene pine decline in Great Britain and Ireland is uncontested. As a result, *Pinus* pollen was counted during initial, skeletal testate amoebae analysis in this study, as an additional correlative and chronological tool. The pollen counts were converted to concentrations calculated using known-quantity *Lycopodium* spore tablets added during laboratory preparation, and plotted alongside testate amoebae data. The tephra horizon was clearly associated with a significant decline in pine and subsequently allowed a reasonable assumption to be made, attributing the horizon to the Hekla 4 eruption, and thus allowing palaeoecological analysis to continue, prior to geochemical analysis.

Pilcher and Hall (1992) suggest the discovery of the Hekla 3 tephra horizon at both Sluggan Moss and Fallahogy. However, no subsequent geochemical or geochronological data has been published since, including in the northern Irish tephrochronology (Pilcher *et al.*, 1996), and recent studies in the area have geochemically typed other tephra horizons in similar stratigraphic positions (e.g. Swindles *et al.*, 2010a). Consequently, the 'Hekla 3' horizon is excluded from further discussion.

6.4.3 Tephrochronological strategy and methods

A number of novel core-scanning techniques have been proposed for the detection of cryptotephra layers (i.e. those which cannot be seen by eye) in peats. Caseldine *et al.* (1999) presented a methodology based on the rapid scanning of peat sequences for variations in light reflectance and luminescence, indicating tephra horizons. Whilst results were promising, it was found that, in some instances, luminescence peaks occurred where tephra was absent. A technique that uses magnetisation profiles to detect tephra layers has been used with varying degrees of success (van den Bogaard *et al.*, 1994; van den Bogaard and Schmincke, 2002). Dugmore and Newton (1992) used X-radiography to identify tephra layers, but encountered problems with the detection of finer layers.

The non-destructive nature of these scanning techniques is clearly beneficial to multi-proxy studies in which the availability of peat for analysis is of high importance. In addition, the potential for high-resolution analysis (e.g. 0.05 mm, Caseldine *et al*., 1999), can allow very thin, and often discontinuous, distal tephra layers to be identified. However, the various problems associated with the techniques, as well as a need for replicate cores (Caseldine *et al*., 1999) and expensive specialist equipment meant that these approaches were not adopted in this study. Rather, a more conventional approach was taken, following a process of ashing and microscopic analysis to extract and confirm the presence of tephra, similar to that developed by Pilcher and Hall (1992). This technique allows the presence or absence of tephra to be determined quickly, over relatively longer cores, and without significant expenditure.

Firstly, contiguous samples, measuring 5 x 1 x 1 cm were taken down the entire length of the core, placed in crucibles and ashed in a furnace set to 600°C for four hours. The ash was then suspended in c. 5 ml 10% hydrochloric acid (HCl) to remove any residual soluble inorganics. The solution was then transferred to a centrifuge tube, using an individually labelled pipette for each sample, and the crucible rinsed again using HCl to ensure no ash remained. The samples were then centrifuged for five minutes at 3000 rpm. The supernatant was discarded and the residue washed with distilled water, before being centrifuged again. This process was repeated to ensure that no HCl remained. After two distilled water washes, the supernatant was carefully removed by labelled pipette until c. 0.5 ml distilled water was left. The sample was gently mixed to re-suspend the residue and pipetted on to a labelled slide, which had been placed on top of a warm hotplate. Once all the water on the slide had evaporated, a small amount of Canada Balsam was used to mount the slides before microscopic examination at x100 and x400 magnification. Samples in which tephra was found to be present were then re-examined at higher resolution to pinpoint the exact position and distribution of the horizon. The relevant depths were then re-analysed with 1cm^3 samples taken at 0.5 cm contiguous resolution. The resolution adopted here, finer than described in most other methodologies (e.g. 1cm, Pilcher and Hall, 1992), has been shown to allow for greater stratigraphic precision in the attribution of depths to tephra horizons (Amesbury, 2008).

Microscopically, tephra shards are identified principally by their colour, shape and vesicularity (e.g. pumiceous grains or connected bubble walls) with sizes ranging from a few microns to c. 2mm (Heiken, 1974; Westgate and Gorton, 1981). In addition, tephra can be distinguished from mineral fragments based on optical properties. Typically, tephra has a lower refractive index (c. 1.5) than that of its mountant, Canada Balsam (1.55). As a result, the Becke line moves into the material with the higher index of refraction when the microscope stage is lowered (i.e. into the Canada Balsam). Minerogenic fragments typically have a higher refractive index than the mountant, and subsequently it is possible to distinguish tephra from mineral fragments using this technique (Enache and Cumming, 2006). Finally, the optically isotropic nature of tephra (i.e. the velocity of light is the same in all directions) differs from other silicate minerals, which are generally anisotropic (i.e. the velocity of light varies with the transmission direction). Subsequently, when plane-polarised light is applied to a slide, tephra appears dark in the microscopic field, when compared to other minerogenic material (Hall and Pilcher, 2002; Chambers *et al.*, 2004; Enache and Cumming, 2006).

Tephra shards are typically subjected to geochemical analysis using an electron microprobe as chemical composition can be used to differentiate between material from different eruptions. However, the ashing methods described above involve heating the shards to temperatures high enough (i.e. > 350°C) to alter their chemical composition (Dugmore *et al.*, 1995b). Consequently, another method of isolating the shards has been developed, and is employed in this study. The acid digestion technique was first developed by Persson (1971), developed by Dugmore (1989) and summarised by Newton *et al.* (2007). Dugmore *et al.* (1992) demonstrated that the technique did not alter shard geochemistry, even in the more mobile elements, and was therefore safe for use prior to electron microprobe analysis. Blockley *et al.* (2005), however, suggest that the technique does risk chemical alteration and potential dissolution of tephra shards. Instead, they propose using a density separation technique to extract tephra shards, additionally citing its success in detecting previously unknown tephras of low shard concentration. Some of the more corrosive stages of the acid digestion procedure that Blockley *et al.* (2005) express concern over, are not required in peat-based studies.

Levels found to contain tephra were prepared for microprobe analysis by placing 3 cm^3 of peat in a conical flask. 50 ml of concentrated (98%) sulphuric acid was added and the flasks were carefully shaken, before being left for 30 minutes for the acid to digest the organic material. Next, 3 ml of concentrated nitric acid was added to the solution and the samples left for a further 15 minutes, to allow the reaction to subside. This was repeated twice more, until the reaction ceased, before the samples were boiled until the liquid turned a transparent yellow. After allowing the samples to cool, they were topped up to 100ml with distilled water and transferred into glass beakers. After leaving the samples to settle for an hour, the solution had cooled and become colourless. The supernatant was then carefully removed using a large pipette, leaving c. 10 ml of residue, which was then transferred to a centrifuge tube and centrifuged at 3000 rpm for five minutes. The supernatant was carefully pipetted off and the sample was washed using distilled water and centrifuged again. This process was then repeated five more times, or until the pH of the sample was measured as neutral using litmus paper. Finally, the supernatant was discarded and the samples were transferred to stoppered glass vials with distilled water, where they were stored until electron microprobe slide preparation.

Prior to analysis by electron microprobe, the sample must be mounted in an epoxy resin, then ground and polished to expose tephra shards. Slide preparation in this study was based on the methods outlined in Langdon (1999), Morriss (2001), Blundell (2002), Amesbury (2008) and Swindles *et al.* (2010b). Firstly, the slides were ground using 100 grit carborundum (or 600 µm alumina grit) paper to produce a suitable surface with which the epoxy can bind. Residual carborundum was then removed by cleaning in petroleum ether within an ultrasonic bath for five minutes. Epoxy resin, in this instance Araldite, was mixed with a hardener at a ratio of 9:1 drops, and a few drops of the solution placed on the slide. Tephra samples were dried on slides warmed on a hotplate before being scraped on to the resin using a razor blade. The samples were subsequently mixed with the resin and left to harden for 90 minutes on a hotplate, producing a resin layer approximately 250 µm thick. In order to expose the tephra shards on the surface, the slide was ground using progressively finer grade carborundum papers (400, 800, 1200, 2500, 4000 grit) until a slide thickness of c. 75 µm was achieved. Care was taken not to over-expose or lose tephra grains, and to maintain a flat, even surface when grinding by frequently checking progress under a reflected light microscope. On completion of grinding, the slides were again cleaned in petroleum ether in an ultrasonic bath for five minutes. Scratches on the surface of the slide can lead to absorption of X-rays during electron microprobe analysis and were therefore removed by polishing, using 6 µm and then 1 µm diamond paste. Between polishes, the slides were, again, cleaned in petroleum ether in an ultrasonic bath. The slides were then carbon coated, before a thin strip of carbon tape was added to improve conductivity across the slide. Slides were labelled using a diamond scriber on the unground side and a pencil on the ground side.

Geochemical typing of tephra shards by electron microprobe analysis, and subsequent correlation to particular eruptions based on glass chemistry, is a well-established practice (Westgate and Gorton, 1981; Larsen, 1981) with protocols that have since been refined (Hunt and Hill, 1993, 1996; Potts *et al.*, 2002; Coulter *et al.*, 2010; Swindles *et al.*, 2010b). Wavelength-dispersive electron probe microanalysis aims an electron beam at each shard, generating X-rays of varying wavelengths and intensities, which relate to chemical composition and abundance, respectively (Reed, 1975; Hunt and Hill, 1993). An accelerating voltage of 10-15 kV and a current of 4-10 nA was used with a defocused beam of 5-10 µm, enabling the analysis of small shards but limiting Na migration (Hunt and Hill, 1993, 2001). All elements were collected for 30 seconds, with the exception of Na, which was collected for 10 seconds. Replicate analyses were performed on the reference material (MPI-DING glasses; Jochum *et al.*, 2006) to check calibration and evaluate the analytical accuracy and precision. The PAP algorithm (Pouchou and Pichoir, 1984) was used to correct for back-scattering, absorbance and fluorescence atomic number effects. For the Sluggan Moss samples, analysis was carried out using a Jeol 8600 electron microprobe in the Research Laboratory for Archaeology and the History of Art, Oxford University by Dr Victoria Smith. For the Fallahogy Bog samples, analysis was carried out using a Cameca SX100 electron microprobe at Edinburgh University, by Dr Matthew Amesbury (Amesbury, 2008).

After electron microprobe analysis, normalisation of geochemical data is a common practice, particularly in the Southern Hemisphere and the Americas, to remove the effects of shard hydration (e.g. Pearce *et al.*, 2008). However, it is advised that, where possible, this practice should be avoided to prevent any alteration in the quality of the data (Hunt and Hill, 1993; Pollard *et al.*, 2006; Swindles *et al.*, 2010b).

Whilst it has been suggested data with oxide totals of less than 95% should be omitted from analyses (Hunt and Hill, 1993), oxide totals of ≥95% are not always consistently obtainable, owing frequently to the primary magmatic and/or post-depositional hydration of the tephra. Subsequently, lower minimum oxide totals (e.g. ≥91%, Morriss, 2001; ≥93%, Eiríksson *et al.*, 2000, Bergman *et al.*, 2004) have been imposed. Given the strong intra-population uniformity (i.e. data were well clustered with few or no outliers) of the geochemical data in this study, it was decided that it was appropriate to include all data points possessing oxide totals of ≥91% at Sluggan Moss and ≥94% at Fallahogy Bog. The number of analysed shards considered necessary for confident geochemical identification varies in the literature and is left to the discretion of the researcher, but generally a minimum of twenty is considered suitable (Swindles *et al*., 2010b). Therefore, combined with internal geochemical uniformity, the shard populations, presented in Sections 6.5.1.2 and 6.5.2.2, were deemed acceptable for geochemical identification. Geochemical results were compared with existing datasets available on Tephrabase (Newton *et al*., 2007)

6.5 Chronological results

6.5.1 Sluggan Moss

6.5.1.1 Radiocarbon dating

A well-distributed series of nine ^{14}C ages were processed at Sluggan Moss, centring on the Hekla 4 tephra horizon, encompassing the period c. 6400 - 2300 cal yr BP. Details of the analyses, including the results and calibrations, are summarised in Table 6.2 and Figure 6.3. After calibration, all ages fell in stratigraphical sequence with no age reversals present. Consequently, the ages can currently be considered a reliable record of peat accumulation at the site.

6.5.1.2 Tephra analysis

As described in Section 6.4.3, ashing was undertaken at Sluggan Moss to initially locate any tephra horizons that may be present. As Figure 6.4 demonstrates, a clear and discrete horizon was detected at a depth of 220-225cm. The horizon was further isolated between these depths by ashing at 0.5cm resolution, with the results shown in Figure 6.5, confirming the presence of a distinct and prominent horizon. This examination of tephra distribution, in terms of both quantity and depth, revealed that c. 78% of the total tephra isolated in these analyses was located within 1cm, centred at a depth of 222cm. As a result, the tephra horizon can be confidently attributed to a depth of 222cm.

Prior to geochemical analysis, reasonable assumptions could be made regarding the identity of this tephra horizon, based initially upon the *Pinus* decline (see Section 6.4.2). Figure 6.6 demonstrates the position of the tephra horizon in relation to *Pinus* pollen from the same sequence at Sluggan Moss. The horizon is coincident with a significant decline in *Pinus* pollen and it can, therefore, be reasonably stated to be consistent with the likely position of the Hekla 4 tephra within the sequence. This provided a central focus for palaeoecological and isotopic analysis, given the temporal association between the Hekla 4 tephra and the 4.2 kyr event (see Section 6.4.2), and allowed laboratory preparations and analyses to proceed whilst arrangements for radiocarbon and geochemical analyses were made.

Table 6.2 Radiocarbon age data from Sluggan Moss.

Laboratory code	Depth (cm)	Material	Radiocarbon date (^{14}C years BP)	δ^{13}C	Calibrated age (2σ; cal yr BP)	
					Age range	*WA mid-point*
UBA-19064	96	*Sphagnum* stems and leaves	2281 ± 27	-26	2163 – 2349	2326
SUERC-33593	148	*Sphagnum* stems and leaves	2912 ± 36	-26.3	2952 – 3207	3059
SUERC-33594	168	*Sphagnum* stems and leaves	3246 ± 35	-23.4	3391 – 3559	3475
BETA-284116	188	*Sphagnum* stems and leaves	3450 ± 40	-25.3	3616 – 3834	3725
UBA-19708	204	*Sphagnum* stems and leaves	3990 ± 35	-32.3	4319 – 4568	4470
UBA-19709	238	*Sphagnum* stems and leaves	4184 ± 35	-29.4	4585 – 4839	4690
SUERC-33595	254	*Sphagnum* stems and leaves	4379 ± 36	-22.7	4858 – 5042	4950
SUERC-33596	270	*Sphagnum* stems and leaves	4744 ± 38	-24.9	5326 – 5587	5517
BETA-284117	282	*Sphagnum* stems and leaves	4870 ± 40	-25.4	5483 - 5709	5621
SUERC-33598	296	*Sphagnum* stems and leaves	5058 ± 35	-27.3	5726 – 5906	5816
UBA-19065	333	*Calluna vulgaris* wood	5617 ± 31	-26.3	6311 – 6465	6382

Figure 6.3 Plot of calibrated radiocarbon ages from Sluggan Moss. A simple linear regression line is presented for reference purposes only.

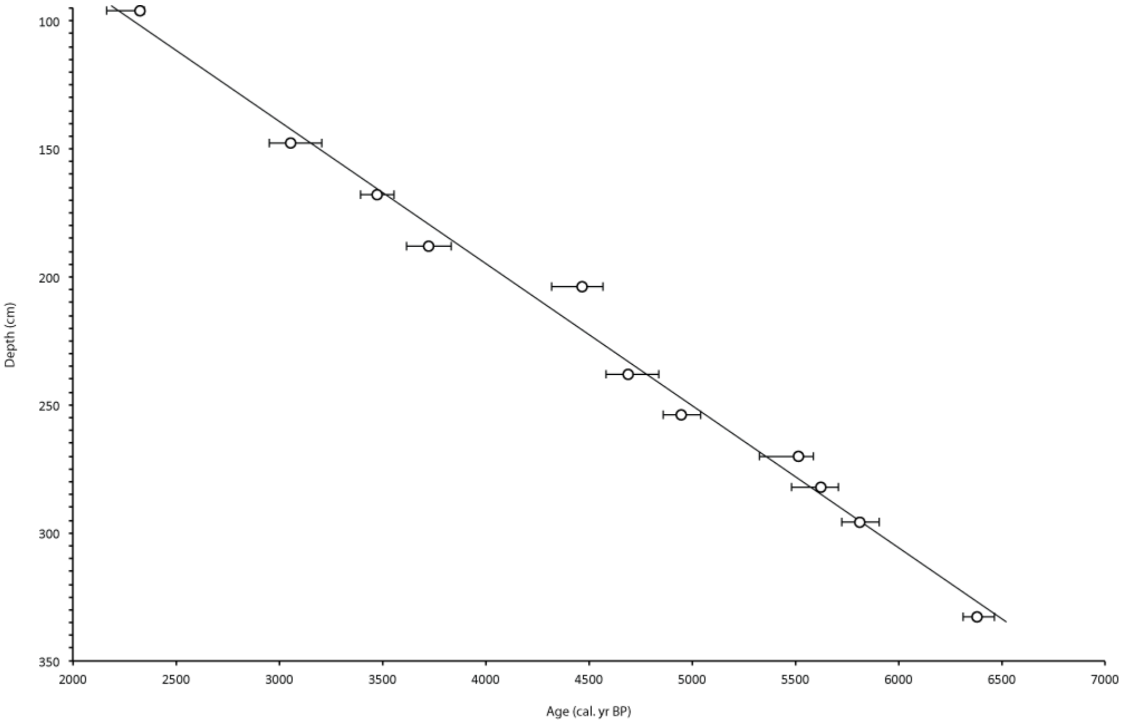


Figure 6.4 Results of 5cm resolution ashing at Sluggan Moss (95-350 cm).

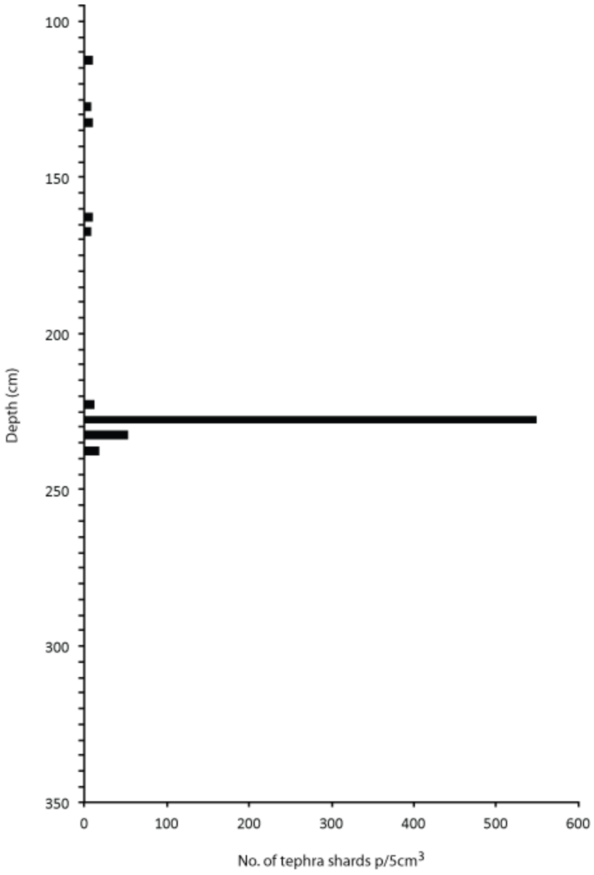


Figure 6.5 Results of 0.5cm resolution ashing at Sluggan Moss (220-225 cm).

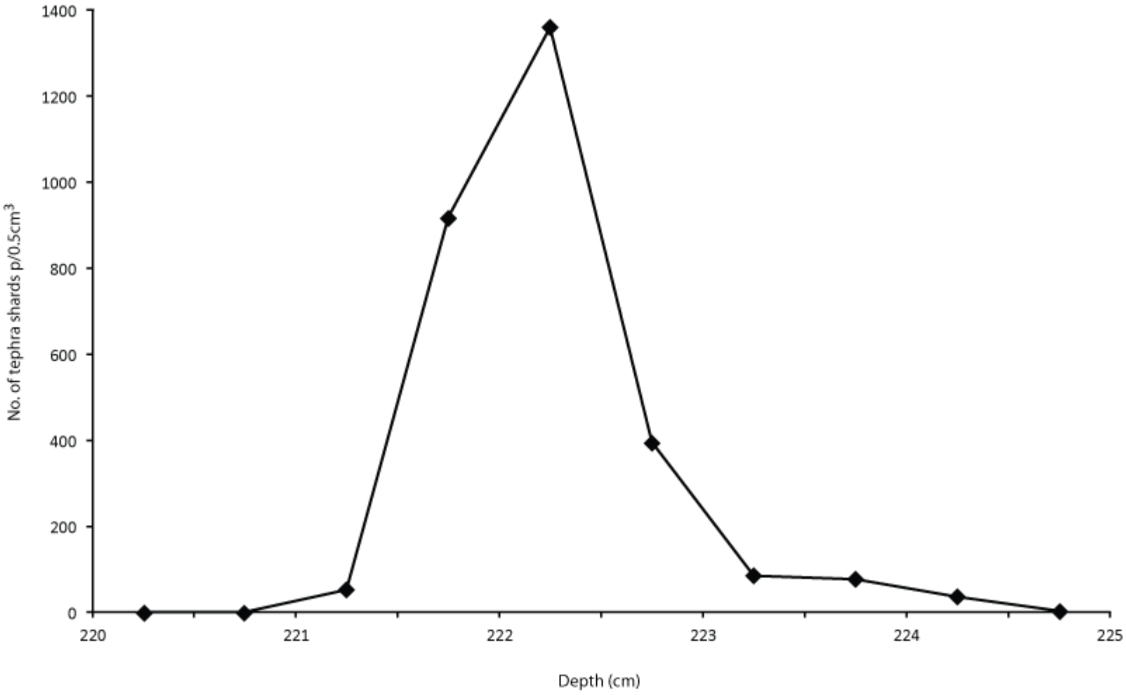
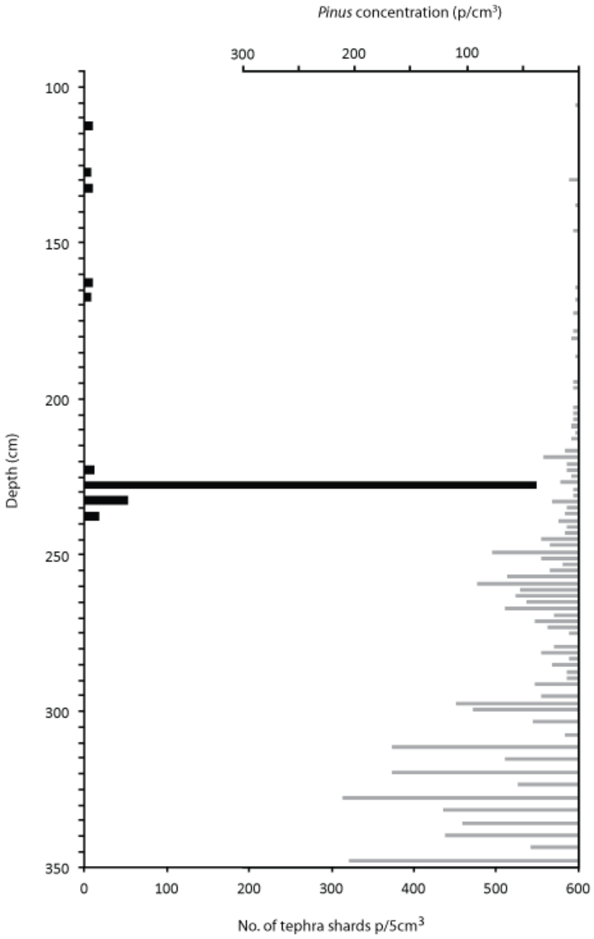


Figure 6.6 Position of tephra horizon against *Pinus* pollen record at Sluggan Moss.



Furthermore, the position of the tephra within the initial radiocarbon stratigraphy confirmed the likelihood of its association with the Hekla 4 eruption. The eruption is arguably one of the best dated in existing northwest European tephrochronologies and, as a result, has been confidently assigned a date range of 4229 – 4345 cal yr BP (Pilcher *et al*., 1996). Tephrochronologies from the northern Irish region demonstrate that the Hekla 4 tephra horizon occurs stratigraphically independent from other major shard concentrations, thus reducing the possibility for confusion with other horizons, and to be chronologically isolated within the period c. 2700 – 6500 cal yr BP (e.g. Swindles *et al*., 2010a). As Table 6.2 and Figure 6.3 demonstrate, radiocarbon dates between 148cm and 333cm show a consistent and coherent period of peat accumulation spanning a period approximately 2900 – 6500 cal yr BP. Therefore, the likely identification of the horizon as the Hekla 4 tephra based on pollen evidence, could be further supported by its position in the preliminary radiocarbon chronology. Despite the confidence of this identification, geochemical analysis of tephra horizons remains vital if they are to be employed in the development of robust, age-depth models, as is the intention of this study.

From the tephra horizon at 222cm at Sluggan Moss, 12 shards were subjected to electron microprobe analysis of which 3 were accepted for geochemical identification, in accordance with the stipulations set out in Section 6.4.3. Geochemical data are presented in Table 6.3. Geochemical data were compared with an existing type-dataset for the Hekla 4 eruption, obtained from Tephrabase, with good agreement between average values for each chemical element, whilst also demonstrating good internal consistency, as indicated by low standard deviations for all elements (Tab. 6.3). However, it is well documented that eruptions from the Hekla volcanic system bear strong chemical resemblance to one another (Van den Bogaard and Schmincke, 2002). Therefore, in order to eliminate any reasonable potential for misidentification, the geochemical data from Sluggan Moss was also compared, alongside the Hekla 4 type-dataset (Tab. 6.4), with type-datasets from two other eruptions from the Hekla volcanic system, Hekla 3 and Hekla-S/Kebister. Whilst stratigraphically distinct (c. 3000 cal yr BP), the Hekla 3 tephra is known to possess geochemical similarity that of the Hekla 4 eruption. The two horizons, Hekla 3 and 4, can be geochemically distinguished by comparing their respective ratios of TiO_2 and FeO (Zillén *et al.*, 2002; Wastegård, 2005; Wastegård *et al.*, 2008). Whilst the regional distribution of the Hekla-S/Kebister tephra is thought to be concentrated northwest of the sites in this study, in the Faroe and Shetland Islands and parts of Scandinavia (Wastegård *et al.*, 2008), it possesses both temporal (c. 3700 cal yr BP) and geochemical similarities with the Hekla 4 horizon. The two horizons, Hekla-S/Kebister and Hekla 4, can be geochemically distinguished by comparing their respective ratios of CaO and MgO (Zillén *et al.*, 2002; Wastegård, 2005; Wastegård *et al.*, 2008). Figures 6.7 and 6.8 demonstrate the horizon found at 222cm at Sluggan Moss is, firstly, geochemically distinct from both the Hekla 3 and Hekla-S/Kebister tephras, and secondly, exhibits strong coherence with the Hekla 4 type-dataset. Whilst the number of shards achieving high totals was lower than desired (cf. Swindles *et al.*, 2010b), shards possessing lower totals (≥71%) plotted in Figures 6.7 and 6.8 also showed good geochemical agreement with the accepted shards. As a result, this tephra horizon can be confidently identified as having derived from the Hekla 4 horizon, therefore, dating between 4229 – 4345 cal yr BP.

Table 6.3 Summary of geochemical data from tephra horizon at Sluggan Moss. Full details of these data can be found in the appendix.

	Na_2O	**MgO**	**Al_2O_3**	**MnO**	**FeO**	**K_2O**	**CaO**	**SiO_2**	**TiO_2**	**Total**
Mean ($\bar{x}$)	4.6776	0.0347	13.5931	0.1024	1.8517	2.7826	1.3402	75.4849	0.1328	94.5158
Std. Dev. (σ)	0.2674	0.0217	0.0784	0.0326	0.1255	0.0458	0.0791	0.2676	0.0290	3.0109

Table 6.4 Average geochemical data from tephra layer at Sluggan Moss, compared with average geochemical data from Hekla 4 type-dataset.

	Na_2O	**MgO**	**Al_2O_3**	**MnO**	**FeO**	**K_2O**	**CaO**	**SiO_2**	**TiO_2**
Sluggan Moss	4.68	0.03	13.59	0.10	1.85	2.78	1.34	75.48	0.13
Type	4.45	0.18	13.24	0.14	3.41	2.56	2.00	71.12	0.24

Figure 6.7 TiO_2 and FeO data for tephra layer at Sluggan Moss, compared with data from Hekla 3 (light grey) and 4 (dark grey) type-datasets. Shards with totals ≥91% (black), ≥71% (red).

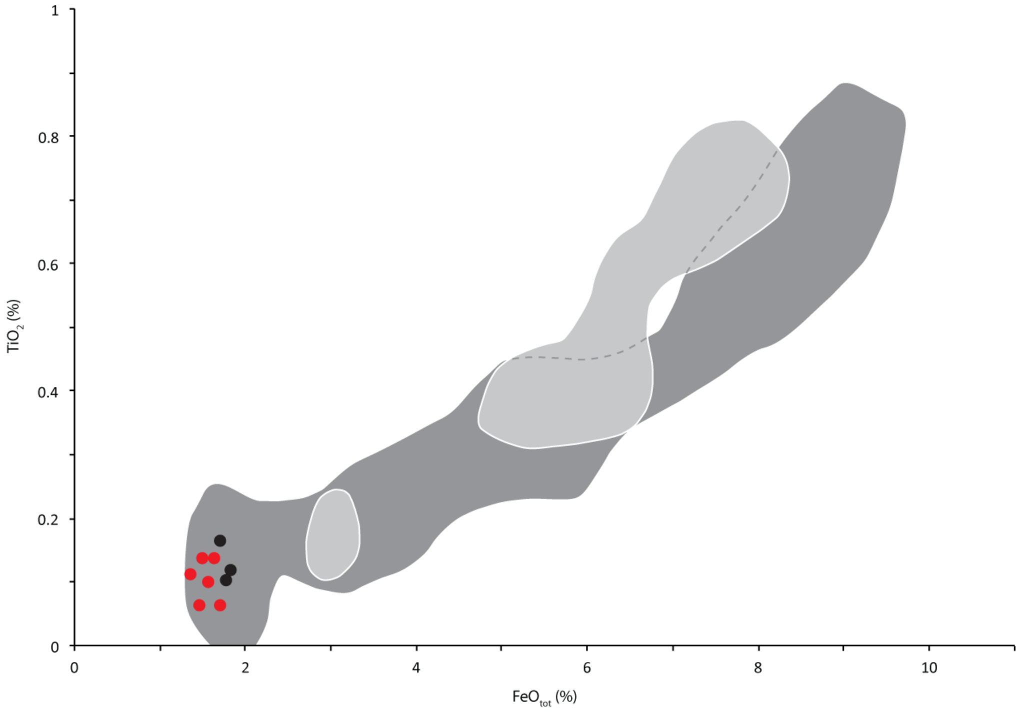
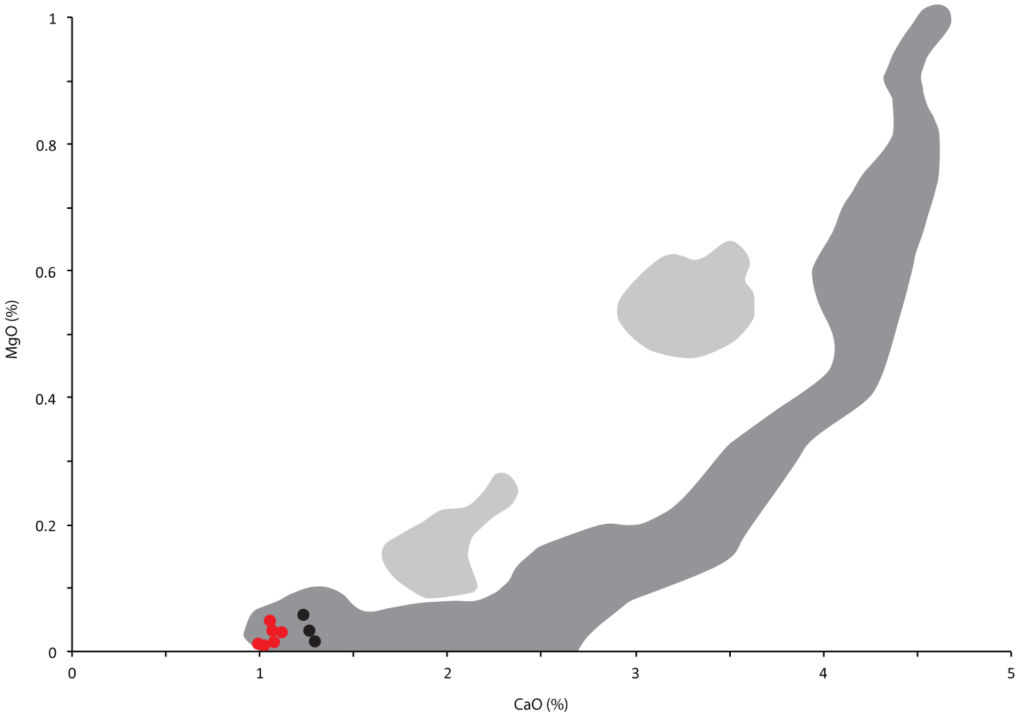


Figure 6.8 CaO and MgO data for tephra layer at Sluggan Moss, compared with data from Hekla S/Kebister (light grey) and 4 (dark grey) type-datasets. Shards with totals ≥91% (black), ≥71% (red).



Other tephrochronological studies based in northern Ireland have detected a greater number of tephra horizons across the period encompassed by this project. A complex of tephra horizons, including BMR-190, OMH-185 (also known, and hereafter referred to, as the Microlite horizon) and GB4-150, summarised by Plunkett *et al*. (2004), are frequently found in close stratigraphic proximity to one another in Irish sites, centred around a period c. 2600 – 2700 cal yr BP.BMR-190 has been found at nine sites across Ireland (Doyle, 1996; Plunkett, 1999; Hall and Pilcher, 2002; J. Holmes, unpublished data; Plunkett *et al*., 2004), but its distribution possesses a western bias, with its absence noted in sites from the northeast of the island. As a result, its absence at Sluggan Moss is neither unexpected nor unusual.

However, both the Microlite and GB4-150 tephra horizons have been identified at sites with close geographical proximity to Sluggan Moss: Garry (c. 40 km northwest), Dead Island (c. 21 km northwest) and Slieveanorra (c. 35 km north) bogs (Plunkett, 2006; Swindles *et al*., 2010a). The Microlite, but not the GB4-150, tephra horizon has also been detected in another core taken from Sluggan Moss (Plunkett, 2006). Therefore, it would be reasonable to expect to find the Microlite tephra within the sequence examined here.

Plunkett *et al*. (2004) assign a date range of 2630 – 2705 cal yr BP to the Microlite tephra horizon. Calibrated radiocarbon results from the site, detailed in Section 6.5.1.1, suggest that, if present, the horizon should occur towards the top of the studied sequence between dates at 96 cm (2163 – 2349 cal yr BP) and 148 cm (2952 – 3207 cal yr BP). As demonstrated in Figure 6.4, a small number of shards were found in this section of the sequence.

The occurrence of small numbers of shards within peat sequences has been noted and discussed in past studies (e.g. Dugmore *et al*., 1995b). Whilst it is possible that small, potentially discrete concentrations of shards, such as these, could be the result of a genuine airfall event, it is also possible that the sediments have been subjected to a certain degree of biomechanical reworking, as plant roots penetrate older sediments. Given the uncertainty surrounding the provenance of such small concentrations, together with the improbability of isolating enough shards for robust geochemical analysis, shard counts of such a low concentration would not normally be subjected to further fine resolution examination. Although in this instance, given the potential for the isolation of a previously confirmed tephra horizon, additional ashing was undertaken. However, it was not possible to identify any further shards.

Another explanation for the absence of the Microlite tephra horizon from Sluggan Moss must therefore be sought. The potential for inter- and intra-site variability in terms of the deposition of any given tephra horizon is great. In a broad examination of tephra horizons in Scotland, Langdon and Barber (2004) found shards from the Glen Garry eruption at Shirgarton Moss, central Scotland, but not at Killorn Moss, just 2 km away. Furthermore, using X-radiography, Dugmore and Newton (1992) demonstrated that the Hekla 4 tephra horizon, in Caithness, northern Scotland, varied significantly in terms of stratigraphic thickness and uniformity of spatial distribution. Using novel light reflectance and luminescence scanning methods, Caseldine *et al*. (1999) present evidence for significant variation in the deposition, concentration and vertical position of tephra on bog surfaces, largely owing to uneven microtopography. It is, therefore, feasible that uneven distribution of the Microlite and GB4-150 tephra horizons on, both, inter- and intra-site scales, could explain their absence in the tephrochronology from this sequence at Sluggan Moss. In addition, Swindles *et al.* (2010a) also found the Microlite and GB4-150 tephra horizons to possess significantly lower shard concentrations, and less discrete distributions in terms of depth when compared to the Hekla 4 eruption, for example, potentially decreasing the likelihood of their detection, even if present in the sequence.

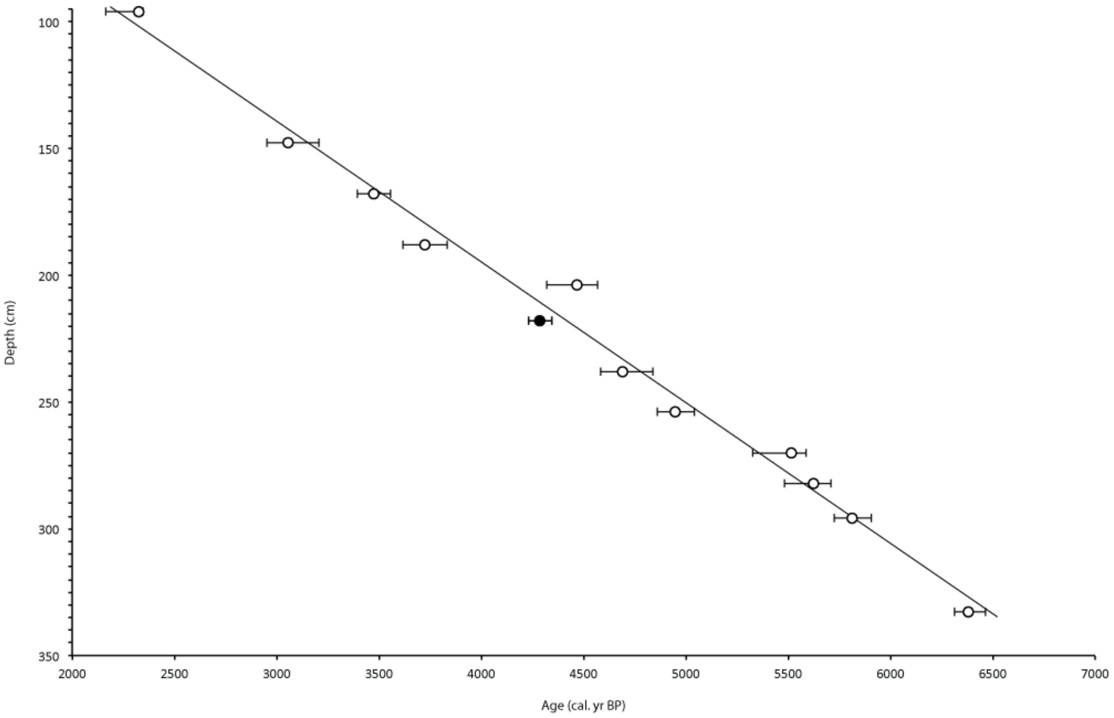
6.5.1.3 Age-depth model development

Initial exploratory age-depth modelling was carried out using the Bayesian *P_Sequence* depositional model in *OxCal* 4.1 (Bronk Ramsey, 1994, 2009). This model assumes that: i) as age increases with depth, radiocarbon dates can be modelled in a specific order, and ii) that fluctuations in the rate of deposition can occur (Bronk Ramsey, 2008b). Therefore, during age-model construction, the *P_Sequence* model utilises depth information for each radiocarbon date, alongside the stratigraphic laws of supposition and succession to constrain posterior age ranges, without assuming a linear sedimentation rate for the sequence (Blockley *et al.*, 2007). The deposition process is modelled as a series of discrete depositional events, in a Poisson process. As a result, a '*k*' factor must be assigned to the model to specify an estimation of the number of events per given depth and the overall stratigraphical process. This factor determines the rigidity under which the model will operate, and is dictated by the nature of the depositional environment. Higher *k* values constrain the data more rigidly and suggest little change in sedimentation rate, as is experienced in relatively simple sedimentary environments. Conversely, lower *k* values permit a greater level of flexibility in age-depth model construction. Consequently, designation of a *k* value must be based upon reasonable, transparent assumption and information derived from the radiocarbon dates themselves (Blockley *et al.*, 2008).

Before a final *P_Sequence* model can be presented, a number of internal tests of the model must be performed. Firstly, it is important to examine the radiocarbon dates that could be considered outliers. Upon inspection of the calibrated radiocarbon ages at Sluggan Moss (Fig. 6.3), they appear to fall in stratigraphical sequence with no age reversals present. Consequently, in this format, the ages can currently be considered a reliable record of peat accumulation at the site. However, when the well dated Hekla 4 tephra horizon is added (Fig. 6.9), radiocarbon date SM204 appears too old. This can be confirmed by statistical analysis.

In *OxCal*, each model, and all the dates within it are assigned an Agreement Index (AI) value, relative to the overall agreement between prior distributions (i.e. calibrated ages) and posterior distributions (i.e. modelled ages). Individual ages with an AI of less than 60% are generally omitted from the analyses. In this instance, the calibrated SM204 age fell below 60% within all models possessing *k* values which enabled them to be considered valid (i.e. AI ≥60%), and can therefore be omitted as an outlier, based on, both, visual and statistical inspection. This date also possessed a markedly lower ^{13}C value when compared with other dates from the sequence.

Figure 6.9 Plot of calibrated radiocarbon ages (hollow) in relation to Hekla 4 tephra horizon (solid) from Sluggan Moss. A simple linear regression line is presented for reference purposes only.

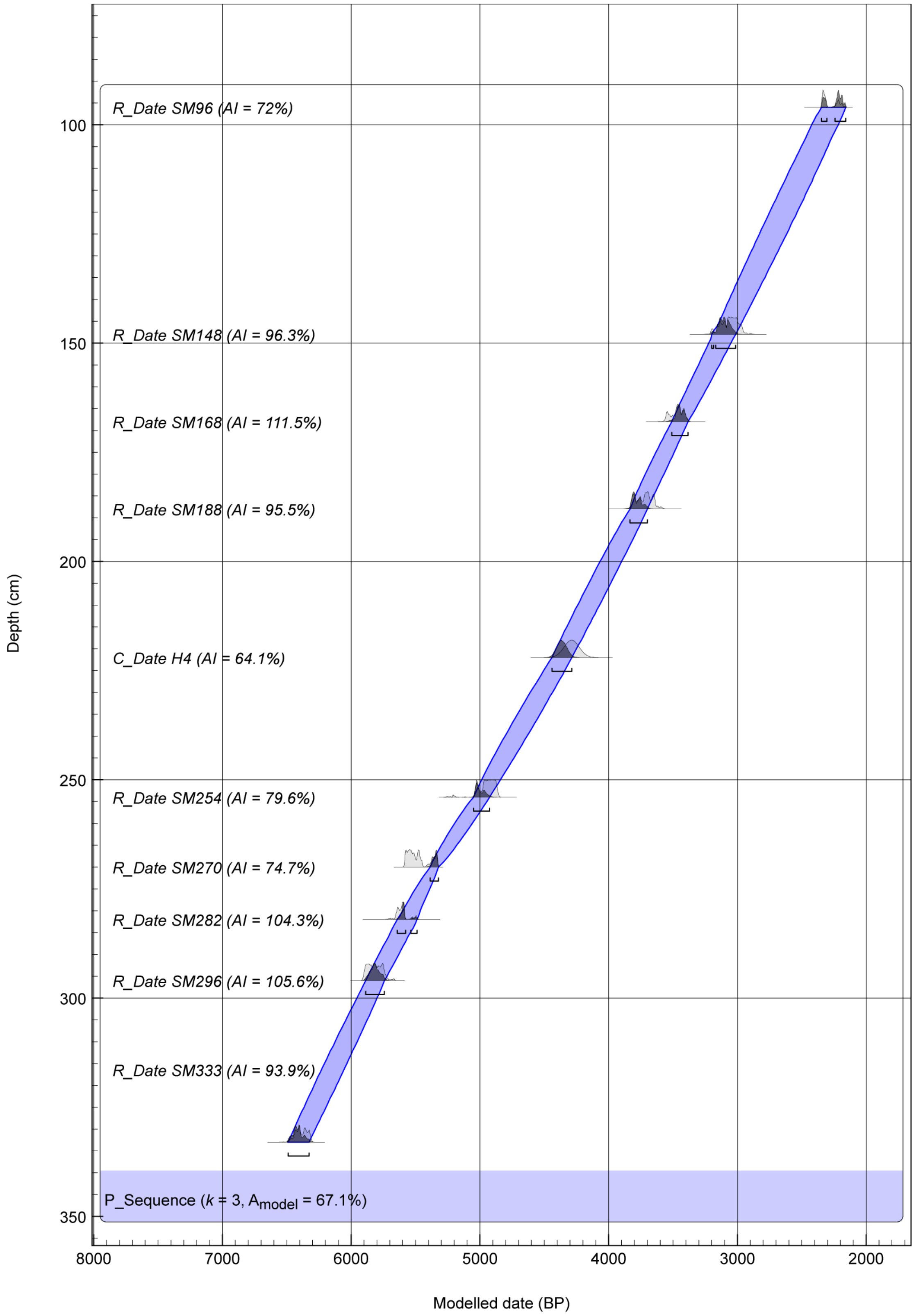


After examination of potential outliers is complete, a number of models must be run to find the optimum *k* value for the *P_Sequence*. This is most effectively achieved by steadily increasing the constraint placed on the data (i.e. increasing from a low starting *k* value) until the model possesses an AI value of close to, but not below, 60%. In order to provide additional constraint, the model has upper and lower 'boundaries', standard mathematical functions, which can be used to delineate between stratigraphic transitions within the sedimentary sequence or mark the start or end of an age model (Bronk Ramsey, 2001). The *P_Sequence* model was consequently assigned a *k* value of 3 cm^{-1} (i.e. 3 depositional events/cm), which is considered to be in the mid-range of model constraint (Bronk Ramsey, 2008b; Blockley *et al.*, 2008). Table 6.5 displays the details of the final *P_Sequence* model at Sluggan Moss, with Figure 6.10 providing a graphical representation.

Table 6.5 Details of *OxCal P_Sequence* (*k* = 3; Model AI = 67.1%) of radiocarbon ages and Hekla 4 tephra horizon from Sluggan Moss. Unmodelled date ranges represent ages calibrated without taking the model into account. Modelled date ranges represent ages calibrated with the model's 'prior information' (i.e. the expectation that dates follow in chronological sequence). Agreement index (AI) shows a measure of how well each age fits within the model's expectations.

	Depth (cm)	**Unmodelled (cal yr BP)**		**Modelled (cal yr BP)**		**Probability (%)**	**AI (%)**
		from	*to*	*from*	*to*		
Boundary				2347	2159		
R_Date SM96	96	2350	2162	2347	2159	95.4	72
R_Date SM148	148	3207	2953	3201	3016	95.4	96.3
R_Date SM168	168	3560	3390	3510	3385	95.4	111.5
R_Date SM188	188	3835	3615	3834	3698	95.4	95.5
C_Date Hekla 4	222	4405	4171	4441	4286	95.4	64.1
R_Date SM254	254	5043	4857	5050	4926	95.4	79.6
R_Date SM270	270	5588	5326	5338	5322	95.4	74.7
R_Date SM282	282	5709	5483	5641	5489	95.4	104.3
R_Date SM296	296	5907	5726	5886	5740	95.4	105.6
R_Date SM333	333	6469	6310	6488	6327	95.4	93.9
Boundary				6488	6327		

Figure 6.10 *OxCal P_Sequence* of radiocarbon ages and Hekla 4 tephra horizon from Sluggan Moss.



The analysis undertaken to construct the *P_Sequence* model in *OxCal* revealed radiocarbon date SM204 to be a statistical outlier. Furthermore, using default settings, *Clam* would not run with the inclusion of this date, as a result of too many age reversals. SM204 was subsequently also excluded from models based in *Clam*. As discussed in Section 6.3.5.4, the 'R'-based statistical package *Clam* has addressed many of the criticisms of traditional or 'classical' age-depth models (e.g. Telford *et al.*, 2004, 2006; Parnell *et al.*, 2011) and, consequently, now presents a viable alternative to age-depth modeling using Bayesian statistics (Blaauw, 2010). Table 6.6 outlines details of the parameters, minimum, maximum and average confidence ranges, and a measure of fit (-log, lower is better) for each model developed in *Clam*, with Figure 6.11 displaying the models graphically. Default settings of note include: a 2σ confidence level (95.4%); 1000 (bootstrap/Monte Carlo) iterations were run per model, removing those that found age-depth reversals; and calendar age point estimates for depths were based on weighted average of all age-depth curves. Custom settings of note include: the omission of the outlier, SM204; small amounts of extrapolation that took place at either end of the model (i.e. 95-96cm and 333-350cm) to encompass all sections of the sequence subjected to palaeoecological analyses. Various polynomial regressions were attempted, of varying orders, with an optimal value of 3 being identified, as orders above this were rejected by *Clam* as they produced too many age reversals.

6.5.1.4 Age-depth model selection

Figure 6.11 and Table 6.6 demonstrate that the four models constructed in *Clam* vary significantly in their confidence ranges and goodness-of-fit. Whilst these values should not be used as directly comparable measures of performance, they can be considered when selecting the optimal age-depth model.

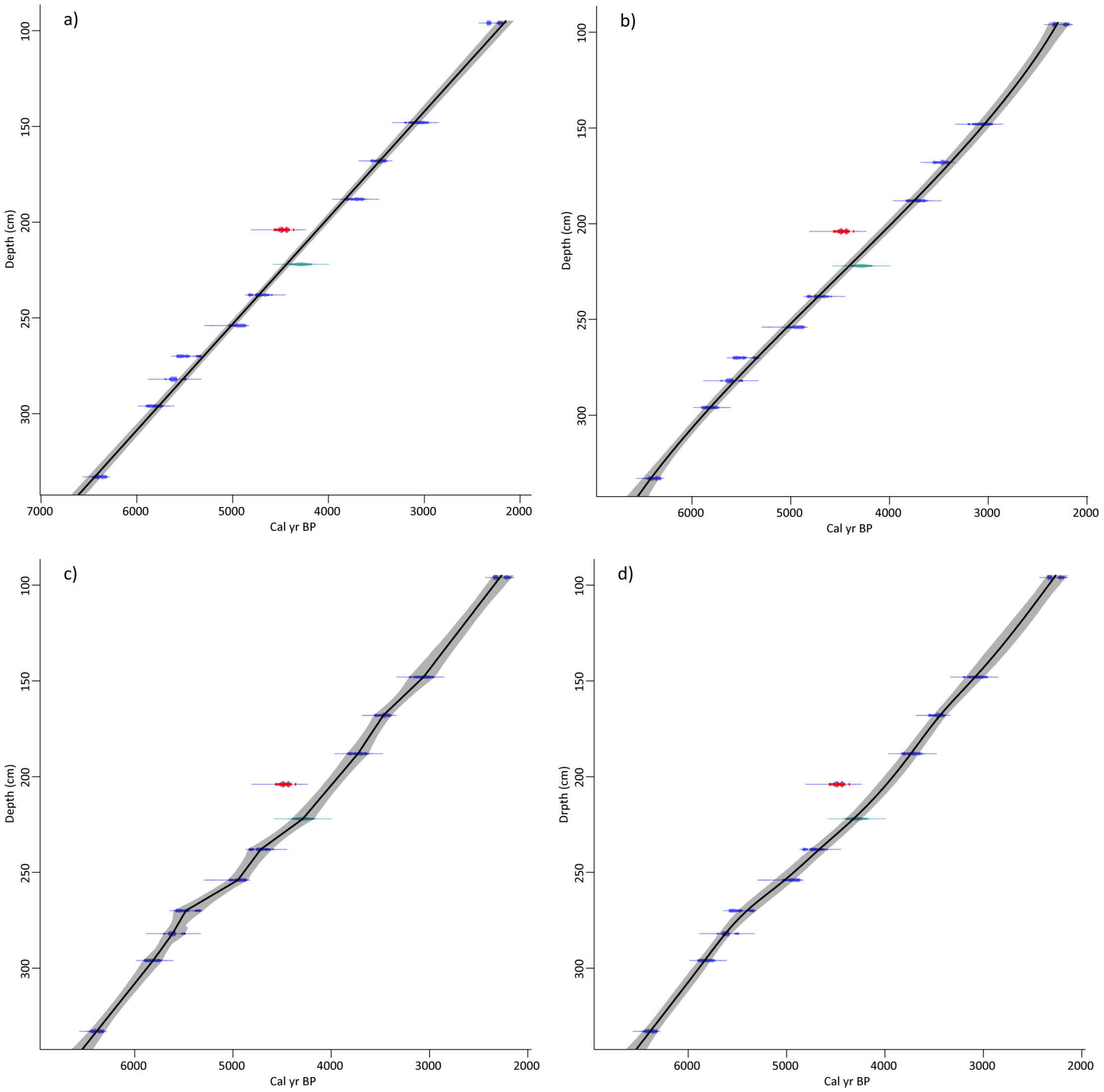
As previously mentioned, linear regression assumes a constant accumulation rate and whilst this is ecologically and stratigraphically unrealistic it can appear to fit the data well (Bennett, 1994). However, the linear regression model presented here (Fig. 6.11a) possesses a poor goodness-of-fit value (Tab. 6.6a), and does not adequately pass through the well-dated Hekla 4 tephra horizon. On this basis, it can be eliminated from the age-depth models.

Bennett and Fuller (2002) have already described the difficulties associated with these models, whereby polynomials with too few terms can perform too rigidly, whilst polynomials with too many terms can lack constraint. Here, the polynomial regression model (Fig. 6.11b) performs poorly in terms of goodness-of-fit (Tab. 6.6b) and, again, fails to adequately pass through the well-dated Hekla 4 tephra horizon, excluding it for this selection process. Higher order polynomial models were attempted for both sequences but spurious flexibility occurred in both instances.

Table 6.6 Details of 'Classical' age depth models developed in *Clam* using radiocarbon ages and Hekla 4 tephra horizon from Sluggan Moss, including parameters, minimum, maximum and average confidence ranges, and a measure of fit (-log, lower is better).

Age-depth model	**a)** Linear regression [type=2]	**b)** Polynomial regression [type=2] of order 3 [smooth]	**c)** Linear interpolation between dated levels [type=1]	**d)** Smooth spline [type=4] with spar = 0.3 [smooth]
Confidence range min.	75	103	127	108
Confidence range max.	150	266	287	253
Confidence range average	102	130	181	161
Goodness-of-fit (-log)	24.85	25.29	16.02	18.62

Figure 6.11 'Classical' age depth models developed in *Clam* using radiocarbon ages and Hekla 4 tephra horizon from Sluggan Moss: a) linear regression; b) 3rd order polynomial regression; c) linear interpolation; d) smooth spline.



Both the linear interpolation (Fig. 6.11c) and smooth spline (Fig. 6.11d) models pass through all non-outlying data points within the model, and subsequently possess relatively high goodness-of-fit values (Tab. 6.6c, d). Again, such models are inherently likely to produce higher goodness-of-fit values than regression models, but provided they are based on sound logic should be accepted where possible. Whilst frequently producing age-depth models that appear plausible, linear interpolation models assume abrupt changes in accumulation rate at the dated depths (Bennett, 1994), and are best employed within a dating strategy that has been able to identify where such changes may have occurred, based on stratigraphic or ecologically important points in the sequence. As no major stratigraphical changes were seen at either site, linear interpolation could be deemed to place inappropriate constraint on the chronological data available. This dating strategy also prevented the practice of wiggle match dating being applied to optimally constrain the radiocarbon chronology (see Section 6.3.5.2).

Blaauw (2010) suggests that sequences that have accumulated within a stable environment are likely to have experienced fewer hiatuses and/or dramatic accumulation changes than those which have accumulated in a more variable environment, and should therefore be modelled using a smooth age-depth model. Blaauw (2010) acknowledges that whilst the 'true' age of a core's dated depths will never be known, an age-depth curve (e.g. smooth spline) can provide a reliable approximation of a sequence's accumulation history. Smooth age-depth models, which can be achieved using polynomial or spline techniques, avoid abrupt changes in accumulation rate. They have previously been criticised for neglecting to consider errors associated with radiocarbon ages (Bennett, 1994), however, this can been rectified through the use of *Clam*.

In spline interpolation, curves are fitted between pairs of points to create a smooth curve, whilst also employing local information (i.e. consideration of the other dates in the sequence). As in linear interpolation, spline interpolation models force the age-depth model through each radiocarbon age, but do not possess the unrealistic assumption that accumulation rates change abruptly. Therefore, given the dating strategy employed in this study, as discussed above, the spline interpolation technique would be favoured over linear interpolation, if possible. Attempts to fit cubic spline models were made, but *Clam* rejected all variations, as a result of too many age reversals. The smooth spline model presented here (Fig. 6.11d) was found to perform well and, as a result, is accepted as the optimal age-depth model produced by *Clam* at Sluggan Moss.

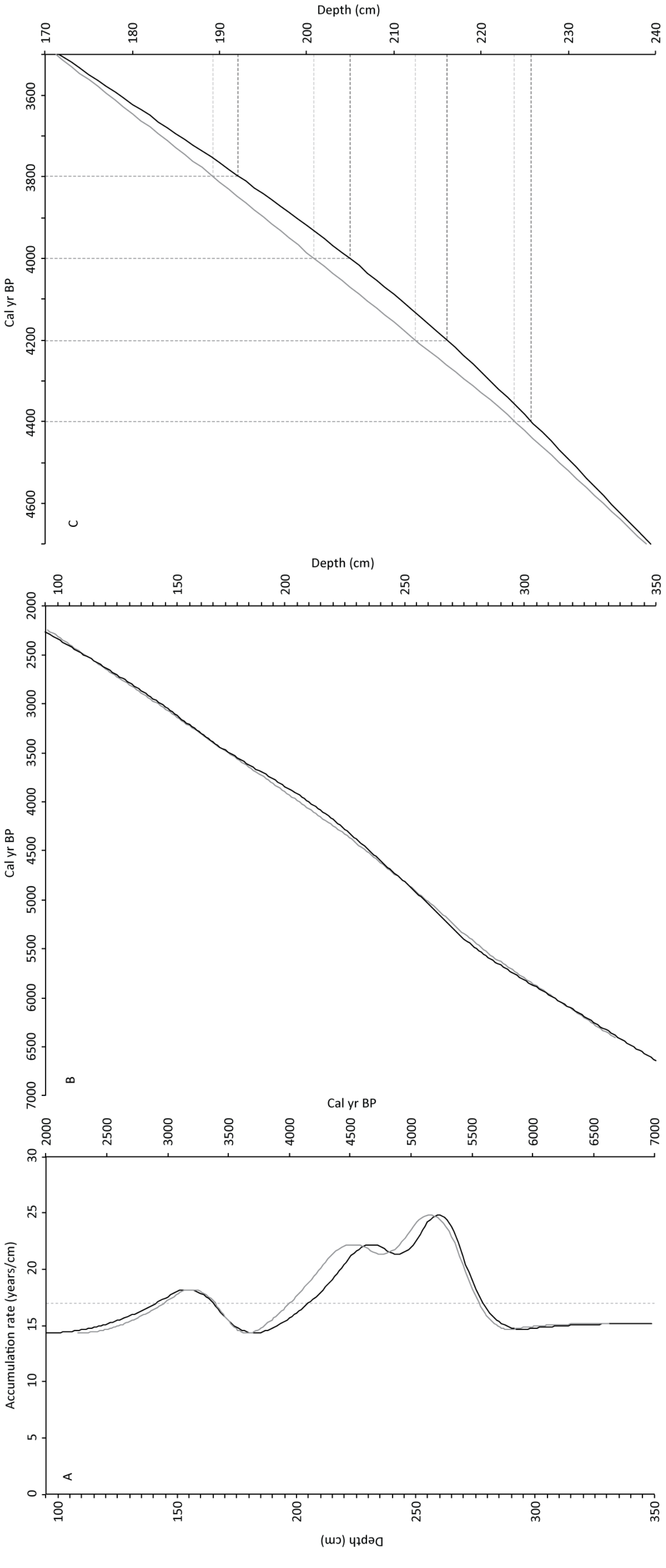
Based on the smooth spline model, accumulation rates at Sluggan Moss for the studied period fluctuate between c. 14 and 25 years/cm, with an average of 17.16 years/cm (Fig. 6.12a). This accumulation rate places the site within desired rate of 10 – 20 years/cm as specified in this study's site selection criteria (see Section 4.2.1), and is therefore suitable for palaeoclimatic analysis of mid-Holocene climatic events. This rate is also slightly faster than previously stated accumulation rates for the site (e.g. 20 years/cm; Plunkett, 2006, 2009).

A comparison between the *OxCal P_Sequence* and *Clam* smooth spline models demonstrates that differences are subtle (Fig. 6.12b), further supporting the use of *Clam* over Bayesian methods. The average difference between the models in this instance is 44.5 years, which is well within the error margins associated with calibrated radiocarbon ages. In addition, comparison of the two models during the period centring on 4200 – 4000 cal yr BP (Fig. 6.12c, Tab. 6.7) shows that only moderate differences exist and are unlikely to have significant implications for an examination of the timing of the '4.2 kyr event', as it may appear in palaeohydrological data in this study.

Table 6.7 Comparison of estimated depths for ages associated with the '4.2 kyr event' from the *OxCal P_Sequence* and *Clam* smooth spline models developed for Sluggan Moss.

Age (cal yr BP)	*Clam* smooth spline estimate (cm)	*OxCal P_Sequence* estimate (cm)	Difference (cm)
3800	197.5	189.5	8
4000	209.5	200.5	9
4200	219.5	212.5	7
4400	229.5	223.5	6
		Average	7.5

Figure 6.12 a) Accumulation rates at Sluggan Moss based on the *Clam* smooth spline model, plotted against depth (black) and age (grey), with average accumulation rate (dashed); b) Comparison of the *OxCal P_Sequence* (grey) and *Clam* smooth spline (black) models developed for Sluggan Moss. Only best point estimates are shown for each model, rather than full age ranges, as it is against these that palaeohydrological data will be plotted (see Section 6.5.3); c) Comparison of the *OxCal P_Sequence* and *Clam* smooth spline models developed for Sluggan Moss for the period 4700 – 3500 cal yr BP.



6.5.2 Fallahogy Bog

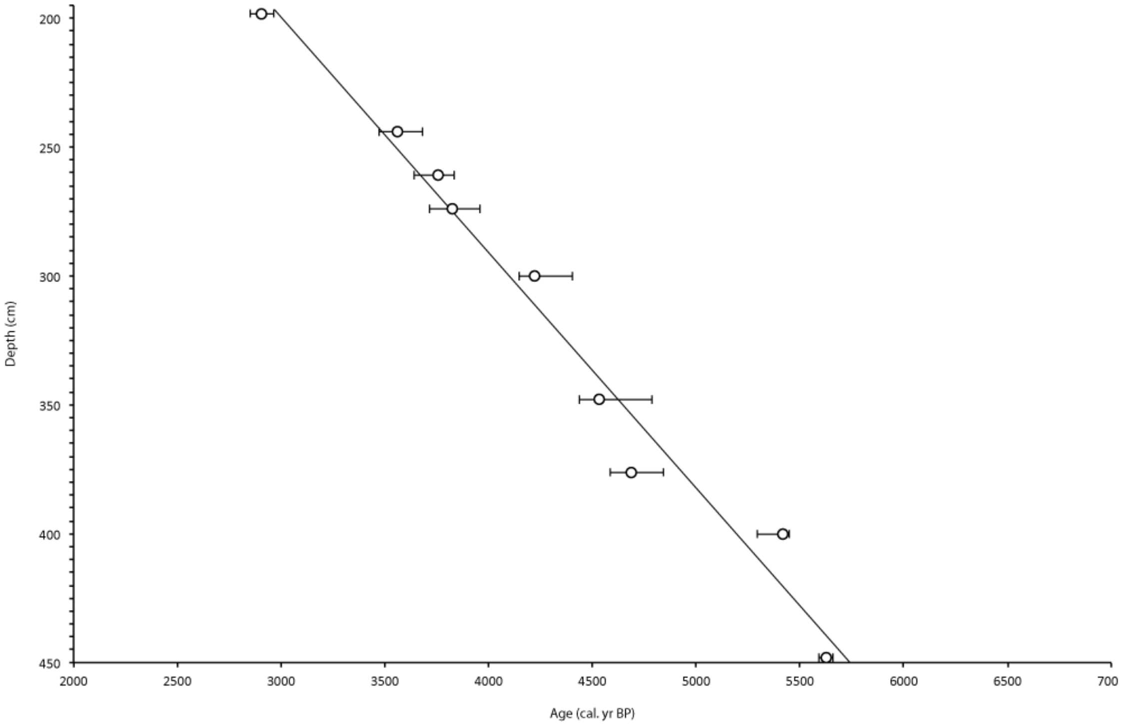
6.5.2.1 Radiocarbon dating

Six additional ^{14}C ages were processed at Fallahogy Bog to complement the three dates processed by Amesbury (2008) and develop a framework of ages. The subsequent chronology encompassed the period c. 5600 - 2900 cal yr BP, and is detailed in Table 6.8 and Figure 6.13. As at Sluggan Moss, Hekla 4 was used as the focus of these ages. Again, no age reversals were present, with all dates falling in stratigraphical sequence, and the ages were considered a reliable record of peat accumulation at the site.

Table 6.8 Radiocarbon age data from Fallahogy Bog. Analyses from SUERC follow Amesbury (2008).

Laboratory code	**Depth (cm)**	**Material**	**Radiocarbon date (^{14}C years BP)**	**δ^{13}C**	**Calibrated age (2σ; cal yr BP)**	
					Age range	*WA mid-point*
UBA-19066	198	*Calluna vulgaris* wood	2806 ± 24	-22.1	2849 – 2965	2907
SUERC-13660	244	*Sphagnum* stems and leaves	3343 ± 37	-28.5	3475 – 3685	3559
SUERC-12999	261	*Calluna vulgaris* wood	3467 ± 35	-29.7	3640 – 3834	3737
SUERC-13000	274	*Sphagnum* stems and leaves	3545 ± 35	-27.4	3716 – 3958	3821
UBA-19607	300	*Sphagnum* stems and leaves	3835 ± 25	-24.2	4150 – 4406	4225
UBA-19068	348	*Sphagnum* stems and leaves	4059 ± 29	-23.8	4435 – 4787	4529
UBA-19710	376	*Sphagnum* stems and leaves	4192 ± 35	-28.9	4588 – 4842	4690
UBA-19069	400	*Sphagnum* stems and leaves	4609 ± 26	-21.8	5296 – 5448	5416
UBA-19070	448	*Sphagnum* stems and leaves	4901 ± 26	-26.7	5589 - 5661	5625

Figure 6.13 Plot of calibrated radiocarbon ages from Fallahogy Bog. A simple linear regression line is presented for reference purposes only.



6.5.2.2 Tephra analysis

As at Sluggan Moss, ashing was undertaken at Fallahogy Bog to initially locate any present tephra horizons. Figure 6.14 demonstrates the occurrence of a clear and discrete horizon, which was then subjected to finer 0.5 cm resolution ashing to isolate the horizon further. Figure 6.15 confirms the presence of a distinct and prominent horizon, with, as at Sluggan Moss, c. 78% of tephra isolated in these analyses located within 1 cm, centred at a depth of 323 cm. As a result, the tephra horizon can be confidently attributed to a depth of 323 cm. In addition to this dominant tephra horizon, there are two levels possessing concentrations of tephra shards, which although small, are noticeable separate to the horizon at 323 cm. Consequently, depths between 360 – 365 cm (10 shards found during initial ashing) and 400 – 405 cm (18 shards) were subjected to fine resolution ashing. No additional shards were found, and it was concluded that these concentrations could most likely be attributed to biomechanical reworking by plant roots, given that they occur below a high-concentration horizon at 323 cm, and did not represent genuine, independent volcanic eruptions. It is also possible that these low-concentration horizons represent smaller eruptions, or eruptions with dispersal patterns did not comprehensively include the region.

As at Sluggan Moss, reasonable assumptions could be made regarding the identity of this tephra horizon, prior to geochemical analysis. Figure 6.16 demonstrates that the position of the tephra horizon in relation to *Pinus* pollen. Again, the horizon is coincident with a significant decline in *Pinus* pollen and is consistent with the likely position of the Hekla 4 tephra, following previous observations (Dugmore *et al*., 1992). Once more, this provided a central focus around which to concentrate palaeoecological and isotopic analyses prior to further radiocarbon and geochemical analyses.

As at Sluggan moss, the development of the radiocarbon stratigraphy at Fallahogy bog provides further support for the attribution of this tephra horizon to the Hekla 4 eruption, owing to its stratigraphic independence and chronological isolation within the period c. 2700 – 6500 cal yr BP in regional tephrochronologies (Pilcher *et al*., 1996; Plunkett, 2006; Swindles *et al*., 2010a). Here, this single tephra horizon is positioned within a period of consistent peat accumulation spanning a period approximately c. 2800 – 5700 cal yr BP (see Table 6.8 and Figure 6.13)

From the tephra horizon at 323 cm at Fallahogy Bog, 35 shards were subjected to electron microprobe analysis of which 24 were accepted for geochemical identification, in accordance with the stipulations set out in Section 6.4.3. Geochemical data are presented in Table 6.9.

Geochemical data was compared with a Hekla 4 type-dataset, demonstrating good agreement between average values for each chemical element (Tab. 6.10). As with the horizon at Sluggan Moss, the geochemical data from Fallahogy was also compared, alongside the Hekla 4 type-dataset, with type-datasets for Hekla 3 and Hekla-S/Kebister, to avoid the possibility of misidentification. Figures 6.17 and 6.18 demonstrate the horizon found at 323 cm at Fallahogy Bog is geochemically distinct from both the Hekla 3 and Hekla-S/Kebister tephras. As a result, this tephra horizon can be confidently identified as having derived from the Hekla 4 eruption and can therefore be a date of between 4229 – 4345 cal yr BP.

Figure 6.14 Results of 5cm resolution ashing at Fallahogy Bog (195-450 cm).

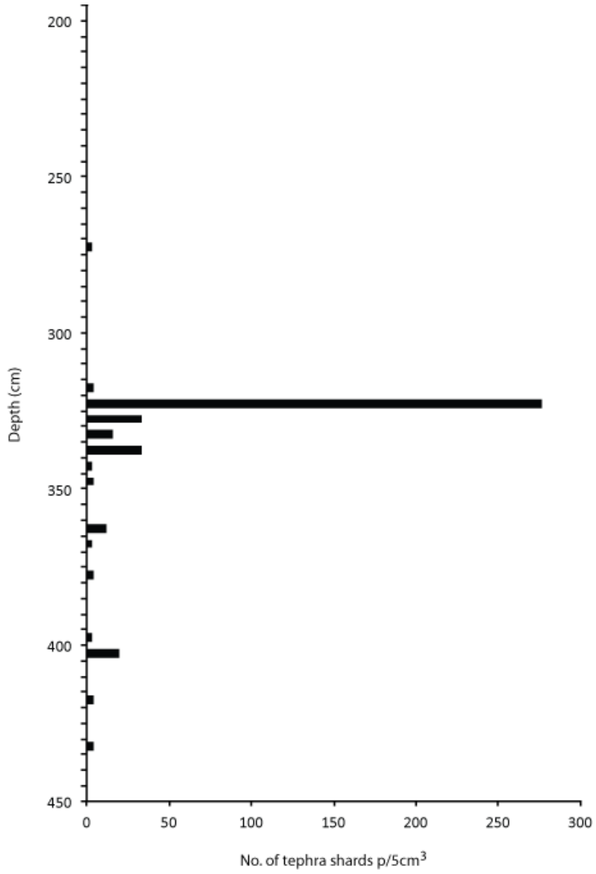


Figure 6.15 Results of 0.5cm resolution ashing between at Fallahogy Bog (320-325 cm).

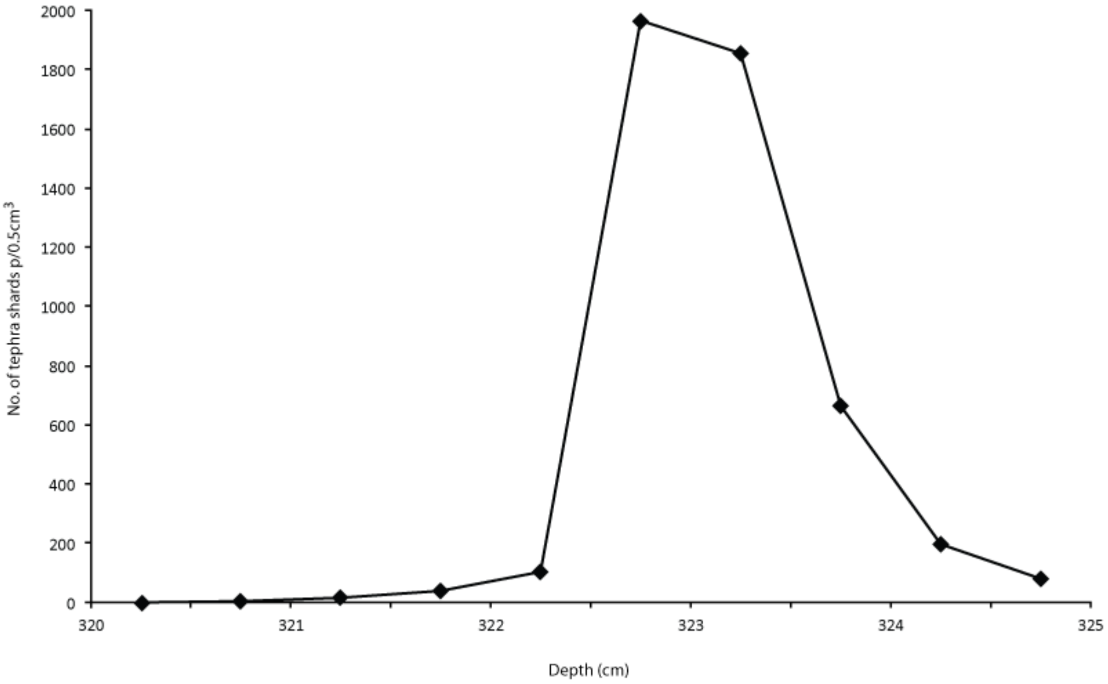


Figure 6.16 Position of tephra horizon against *Pinus* pollen record at Fallahogy Bog.

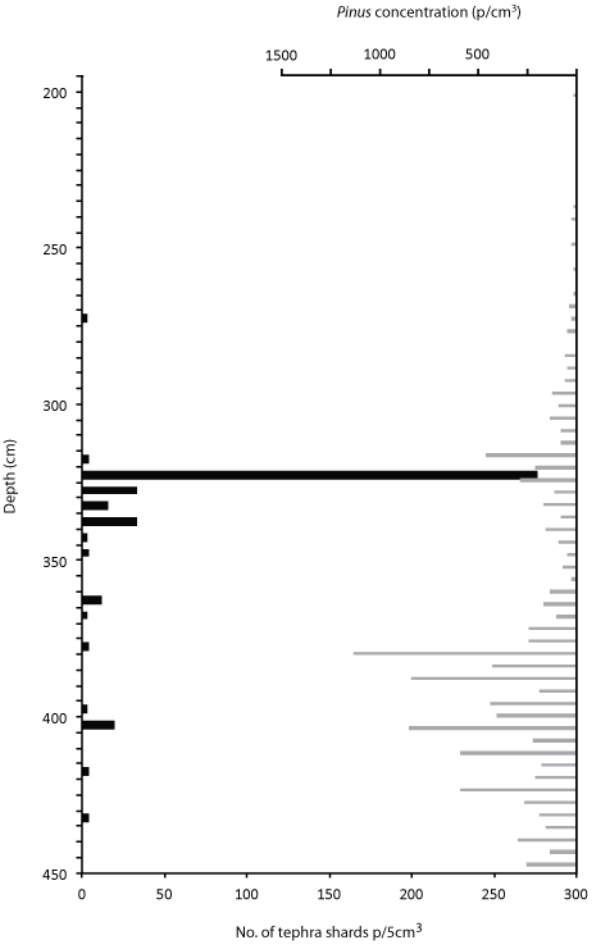


Table 6.9 Summary of geochemical data from tephra horizon at Fallahogy Bog. Full details of these data can be found in the appendix.

	Na_2O	MgO	Al_2O_3	MnO	FeO	K_2O	CaO	SiO_2	TiO_2	P_2O_5	Total
Mean ($\bar{x}$)	4.6434	0.0153	12.404	0.0728	1.9225	2.8256	1.3148	71.7667	0.1029	0.0082	95.118
Std. Dev. (σ)	0.1683	0.0090	0.1913	0.0484	0.1238	0.0667	0.0546	0.6385	0.0399	0.1618	0.9268

Table 6.10 Average geochemical data from tephra layer at Fallahogy Bog, compared with average geochemical data from Hekla 4 type-dataset.

	Na_2O	MgO	Al_2O_3	MnO	FeO	K_2O	CaO	SiO_2	TiO_2
Fallahogy Bog	4.72	0.02	12.49	0.07	1.95	2.84	1.31	72.05	0.11
Type	4.45	0.18	13.24	0.14	3.41	2.56	2.00	71.12	0.24

Figure 6.17 TiO_2 and FeO data for tephra layer at Fallahogy Bog (black dots), compared with data from Hekla 3 (light grey) and 4 (dark grey) type-datasets.

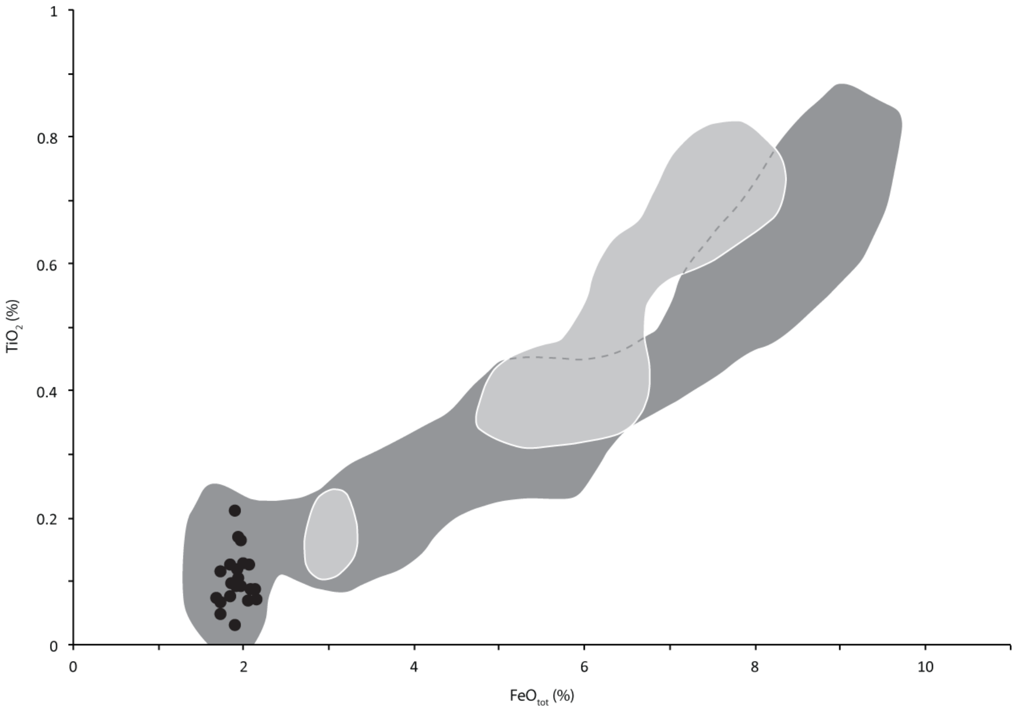
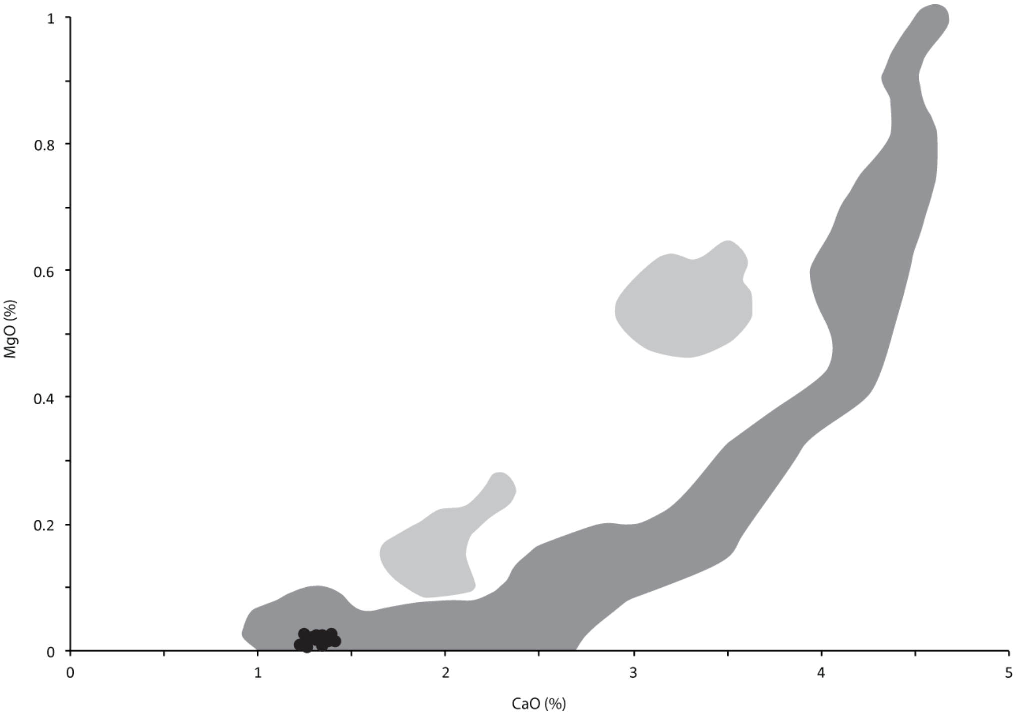
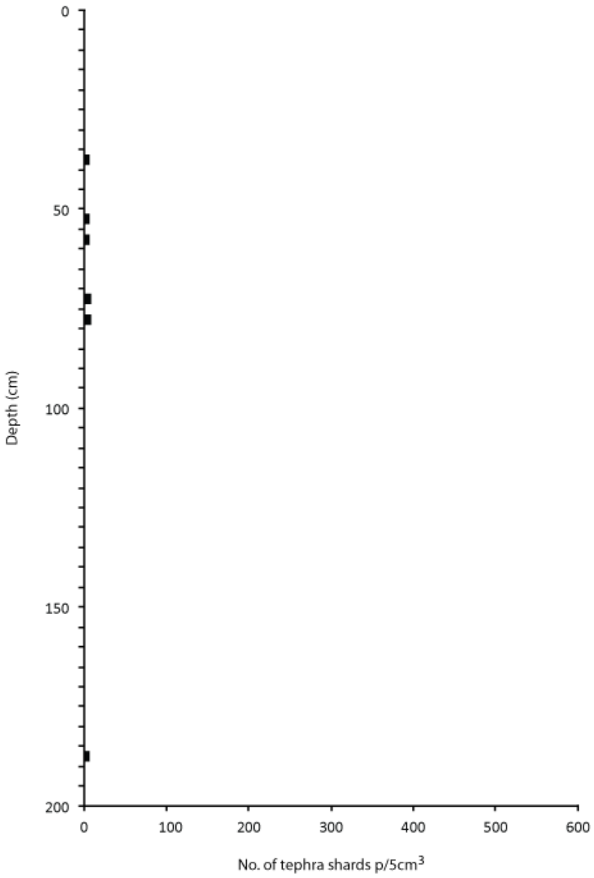


Figure 6.18 CaO and MgO data for tephra layer at Fallahogy Bog (black dots), compared with data from Hekla 3 (light grey) and 4 (dark grey) type-datasets.



As discussed previously, with regard to Sluggan Moss (see Section 6.5.1.2), other tephra horizons have been detected at sites in northern Ireland that encompass the period of interest to this study. Sites in close proximity to Fallahogy Bog, including Garry (c. 22 km north), Garry (c. 2.5 km south) and Slieveanorra (c. 28 km northeast) bogs, possess, both, the Microlite and GB4-150 tephra horizons, neither of which are found here. Based on the calibrated radiocarbon results, outlined in Section 6.5.2.1, these two tephras would be expected to fall just above the top date in this sequence at 198cm (2849 – 2965 cal yr BP), given the approximate date of the Microlite/GB4-150 tephra complex of c. 2600 – 2700 cal yr BP (Swindles *et al.*, 2010a). However, Figure 6.19 shows 5cm resolution ashing of the peat sequence at Fallahogy from the surface to a depth of 200 cm, and demonstrates no evidence for the presence of either tephra horizon, with only a few small concentrations of shards, numbering no more than 2 per/cm^3 sample. As at Sluggan Moss (see Section 6.5.1.2), it is feasible that uneven distribution of the Microlite and GB4-150 tephra horizons on, both, inter- and intra-site scales, could account for their absence in the tephrochronology from this sequence at Fallahogy Bog.

Figure 6.19 Results of 5cm resolution ashing at Fallahogy Bog (0-200cm).



6.5.2.3 Age-depth model development

Upon inspection of the calibrated radiocarbon ages at Fallahogy Bog (Fig. 6.13), they appear to fall in stratigraphical sequence with no age reversals present. The addition of the Hekla 4 tephra horizon to this plot (Fig. 6.20) does not alter this and, consequently, the ages can be considered a reliable record of peat accumulation at the site.

As at Sluggan Moss, initial exploratory age-depth modelling was carried out using the Bayesian *P_Sequence* depositional model in *OxCal* 4.1 (Bronk Ramsey, 1994, 2009). The optimum *k* value was determined to be .25 cm^{-1} (i.e. 0.25 depositional events/cm), which would be considered to be in the lower range of model constraint (Bronk Ramsey, 2008b; Blockley *et al*., 2008). Whilst low, this *k* value is comparable to that of other published studies (e.g. k = 0.25, Swindles *et al*., 2012c), and allows all radiocarbon dates obtained at Fallahogy Bog to be used in the construction of the age-depth model. When models were run with higher *k* values, *OxCal* began to 'identify' outliers, particularly the radiocarbon dates FAL300 and FAL400. Upon visual inspection of these dates, within the wider radiocarbon sequence at Fallahogy Bog, it was not felt that their classification as outliers could be justified. Upper and lower boundaries were also applied to provide additional constraint to the age model (Bronk Ramsey, 2000, 2001). Table 6.11 displays the details of the final *P_Sequence* model at Fallahogy Bog, with Figure 6.21 providing a graphical representation.

It is worth noting that age-depth models developed for these two sites using *OxCal* required notably different '*k*' values (i.e. 0.25 to 3) to function, despite a distinct similarity in depositional environment. Such a lack of inter-site consistency could, therefore, be another justification for disregarding these Bayesian *OxCal* models in this instance (further to those presented in 6.3.5.4).

Figure 6.20 Plot of calibrated radiocarbon ages (hollow) in relation to Hekla 4 tephra horizon (solid) from Fallahogy Bog. A simple linear regression line is presented for reference purposes only.

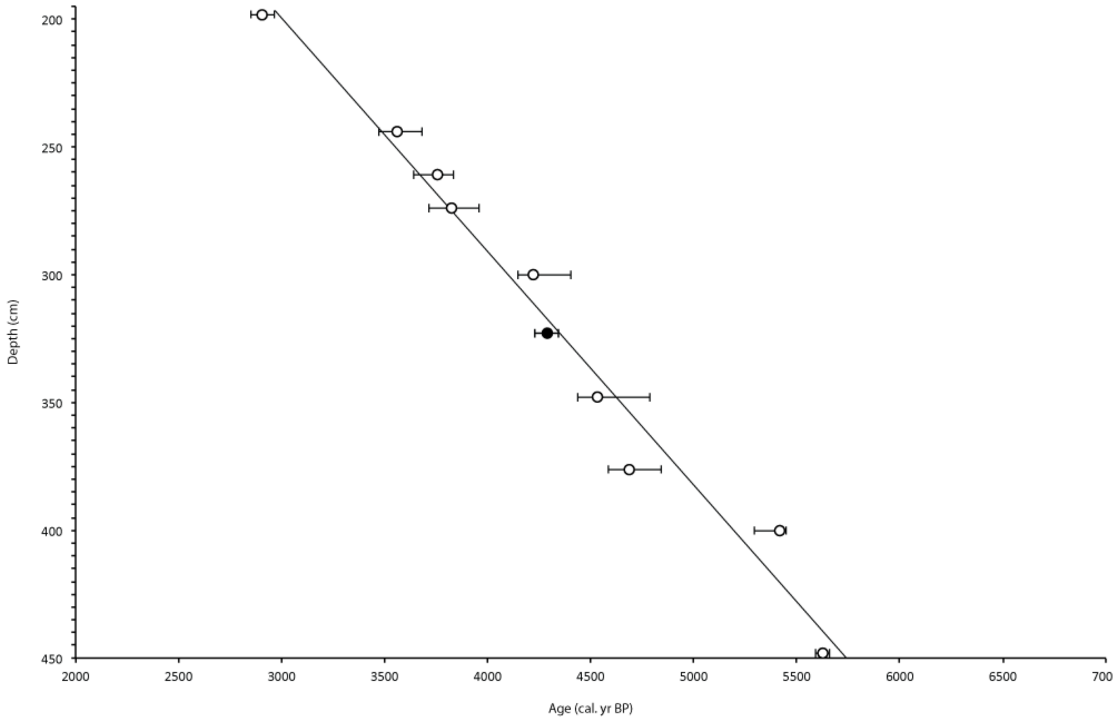
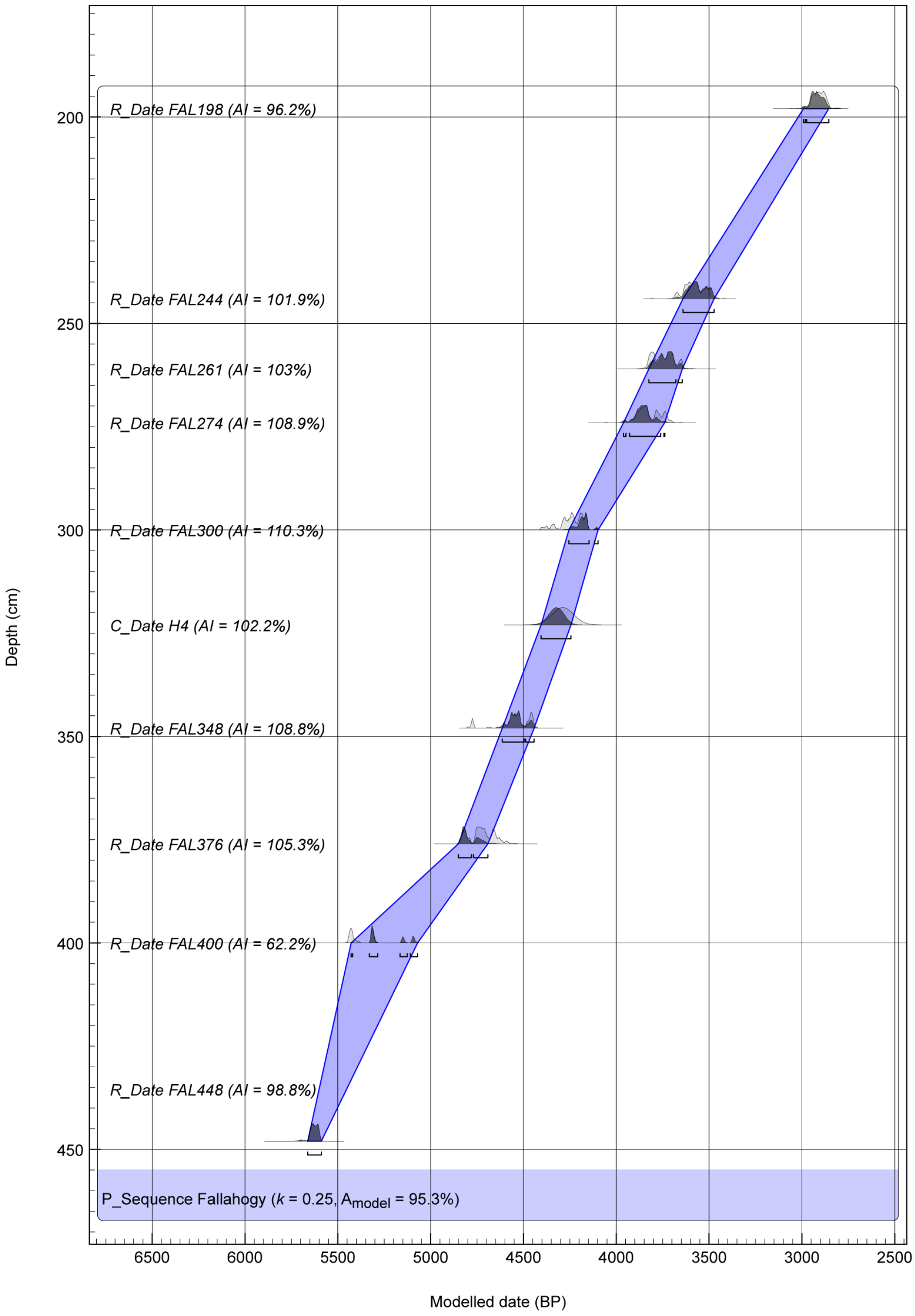


Table 6.11 Details of *OxCal P_Sequence* (*k* = 0.25; Model AI = 95.3%) of radiocarbon ages and Hekla 4 tephra horizon from Fallahogy Bog. See Table 6.5 for explanation of unmodelled and modelled date ranges, and agreement index (AI).

	Depth (cm)	**Unmodelled (cal yr BP)** *from*	*to*	**Modelled (cal yr BP)** *from*	*to*	**Probability (%)**	**AI (%)**
Boundary				2985	2795		
R_Date FAL198	198	2966	2848	2992	2856	95.4	96.2
R_Date FAL244	244	3686	3475	6341	3473	95.4	101.9
R_Date FAL261	261	3835	3640	3825	3644	95.4	103
R_Date FAL274	274	3959	3716	3961	3739	95.4	108.9
R_Date FAL300	300	4407	4150	4255	4098	95.4	110.3
C_Date Hekla 4	323	4405	4171	4405	4244	95.4	102.2
R_Date FAL348	348	4788	4433	4614	4443	95.4	108.8
R_Date FAL376	376	4842	4616	4851	4692	95.4	105.3
R_Date FAL400	400	5450	5295	5428	5070	95.4	62.2
R_Date FAL448	448	5661	5589	5662	5589	95.4	98.8
Boundary				5722	5593		

Figure 6.21 *OxCal P_Sequence* of radiocarbon ages and Hekla 4 tephra horizon from Fallahogy Bog.



As no outliers could be robustly visually and statistically identified, all radiocarbon dates were included in model development in *Clam* (Table 6.12 and Figure 6.22). Default settings were the same as at Sluggan Moss (see Section 6.5.1.3) The only custom settings of note, applied to all models, was the small amount of extrapolation required to extend the age-depth model to encompass all sections of the sequence subjected to palaeoecological analyses (i.e. 195-198cm and 448-450cm). Various polynomial regressions were attempted, of varying orders, with an optimal value of 4 being identified; orders above this were rejected by *Clam* as they produced too many age reversals.

Figure 6.22 'Classical' age depth models developed in *Clam* using radiocarbon ages and Hekla 4 tephra horizon from Fallahogy Bog: a) linear regression; b) 4th order polynomial regression; c) linear interpolation; d) smooth spline.

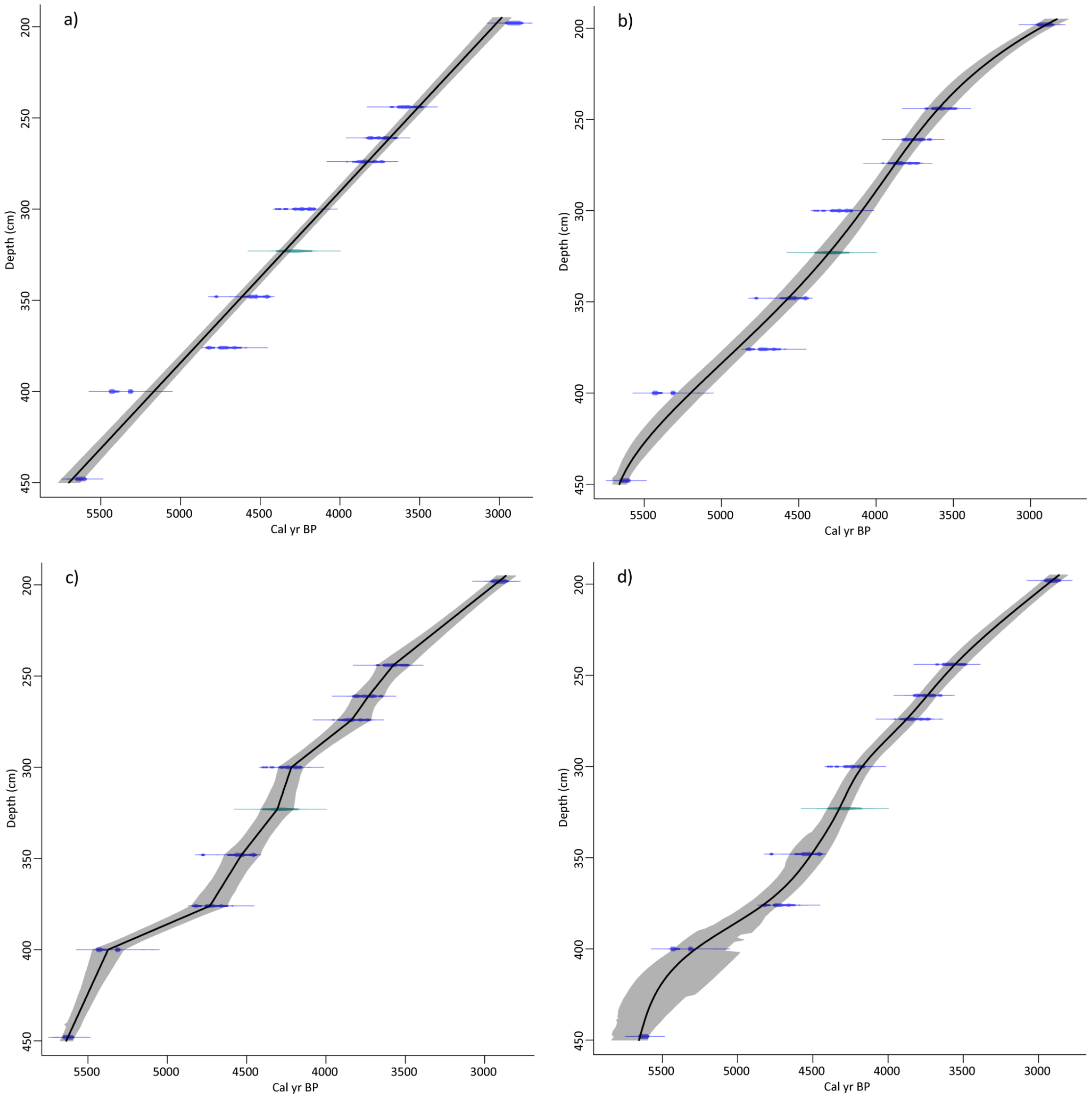


Table 6.12 Details of 'classical' age depth models developed in *Clam* using radiocarbon ages and Hekla 4 tephra horizon from Fallahogy Bog, including parameters, minimum, maximum and average confidence ranges, and a measure of fit (-log, lower is better).

Age-depth model	**a)** Linear regression [type=2]	**b)** Polynomial regression [type=2] of order 4 [smooth]	**c)** Linear interpolation between dated levels [type=1]	**d)** Smooth spline [type=4] with spar = 0.3 [smooth]
Confidence range min.	72	91	77	108
Confidence range max.	134	174	229	502
Confidence range average	91	145	157	211
Goodness-of-fit (-log)	22.57	25.37	8.76	19.17

6.5.2.4 Age-depth model selection

Figure 6.22 and Table 6.12 demonstrate that the four models constructed in *Clam* vary significantly in their confidence ranges and goodness-of-fit. Both sites presented in this study were subject to the same dating strategy and, owing to similarities in their depositional regime, a high degree of similarity exists between the performance of the respective age depth models. Much of the evaluation and discussion of the chronological results presented in the selection of an age-depth model at Sluggan Moss (see Section 6.5.1.4) is applicable here. Both the linear (Fig. 6.22a) and polynomial (Fig. 6.22b) regression models were subsequently excluded from the selection process.

Of the linear interpolation (Fig. 6.22c) and smooth spline (Fig. 6.22d) regression models, the former provides superior goodness-of-fit and confidence range values. However, as at Sluggan Moss, the dating strategy was forced to place dates in a well-spaced framework of increasing resolution as dates became available (Bennett, 1994). Unless dates are strategically placed at point of stratigraphic and/or ecological transition, the abrupt changes in accumulation rates assumed by a linear interpolation model are unrealistic. The problems this can cause are clearly demonstrated in this linear interpolation model. Figure 6.22c displays a near-vertical section of the age-depth model between the Hekla 4 tephra horizon (323 cm) and the radiocarbon date FAL300 (300 cm), suggesting an accumulation rate approaching 1 year/cm. Whilst rapid accumulation of *Sphagnum* peat sections have been documented in northwestern Europe (Barber *et al.*, 2004), accumulation rates as high as this are clearly extremely unlikely.

This provided further justification, alongside that presented in Section 6.5.1.4, for the rejection of the linear interpolation model in favour of the smooth spline regression model. The relatively poor goodness-of-fit and confidence range values associated with the latter model are likely to be heavily skewed by the significant expansion of confidence ranges in the bottom section of the model. Figure 6.22d, demonstrates graphically the expansion in confidence ranges that occurs below the radiocarbon date FAL376. However, if the same smooth spline regression model is run for the depths between radiocarbon dates FAL198 and FAL376 (i.e. excluding the bottom section of the model, see Figure 6.23), then goodness-of-fit and confidence range values are markedly improved (Table 6.13). This suggests that the model performs very well through the period immediately encompassing the 4.2 kyr event and can be accepted as the optimal age-depth model produced by *Clam* at Fallahogy.

Based on the smooth spline model, accumulation rates at Fallahogy Bog for the studied period fluctuate between c. 4 and 20 years/cm, with an average of 10.94 years/cm (Fig. 6.24a). This accumulation rate places the site within desired rate of 10 – 20 years/cm as specified in this study's site selection criteria (see Section 4.2.1), and is therefore suitable for palaeoclimatic analysis of mid-Holocene climatic events. This rate is very similar to previously stated accumulation rates for the site (e.g. c. 10 years/cm; Barber *et al*., 2000). Towards the bottom of the age-model accumulation rates do fluctuate considerably, representing the majority of the range outlined earlier. Accumulation rates of 4 years/cm are not impossible (Barber *et al*., 2004), but are unlikely at depth, more normally being associated with fresh *Sphagnum* growth at the top of a core profile. It is therefore likely that these fluctuations are exaggerated by the expansion in confidence ranges below radiocarbon date FAL376, as previously discussed. Above 376 cm the variation in accumulation rate has a reduced range of c. 6 – 15 years/cm, whilst the average rate remains very similar at 10.91 years/cm.

A comparison between the *OxCal P_Sequence* and *Clam* smooth spline models demonstrates that differences are very subtle (Fig. 6.24b), again supporting the use of *Clam* over Bayesian methods. The average difference between the models in this instance is 28.1 years, which is well within the error margins associated with calibrated radiocarbon ages. In addition, comparison of the two models during the period centring on 4200 – 4000 cal yr BP (Fig. 6.24c, Tab. 6.14) shows that very minor differences exist and would not have significant implications for an examination of the timing of the '4.2 kyr event', as it may appear in the palaeohydrological data of this study.

Figure 6.23 Truncated smooth spline age-depth model developed in *Clam* using radiocarbon ages (FAL376 and above) and Hekla 4 tephra horizon from Fallahogy Bog.

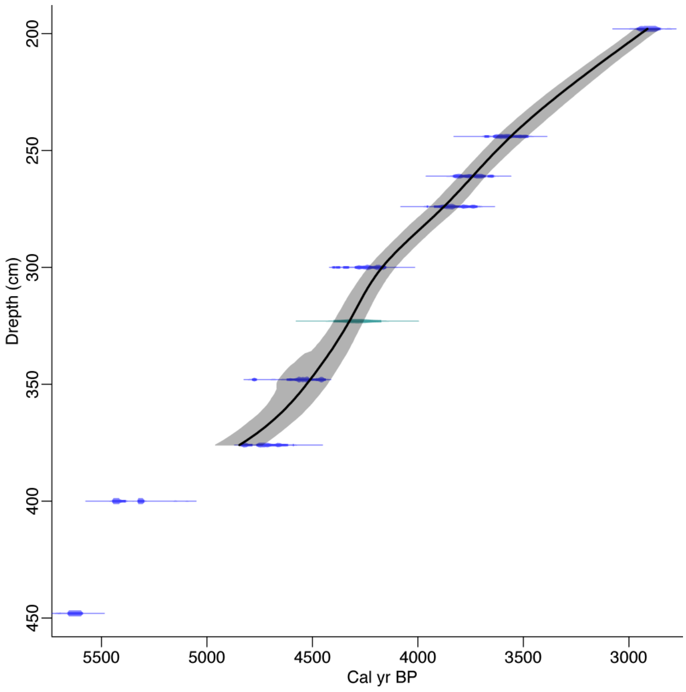


Table 6.13 Details of truncated smooth spline age-depth model developed in *Clam* using radiocarbon ages (FAL376 and above) and Hekla 4 tephra horizon from Fallahogy Bog.

Age-depth model	**Smooth spline** [type=4] with spar = 0.3 [smooth]
Confidence range min.	109
Confidence range max.	241
Confidence range average	148
Goodness-of-fit (-log)	5.97

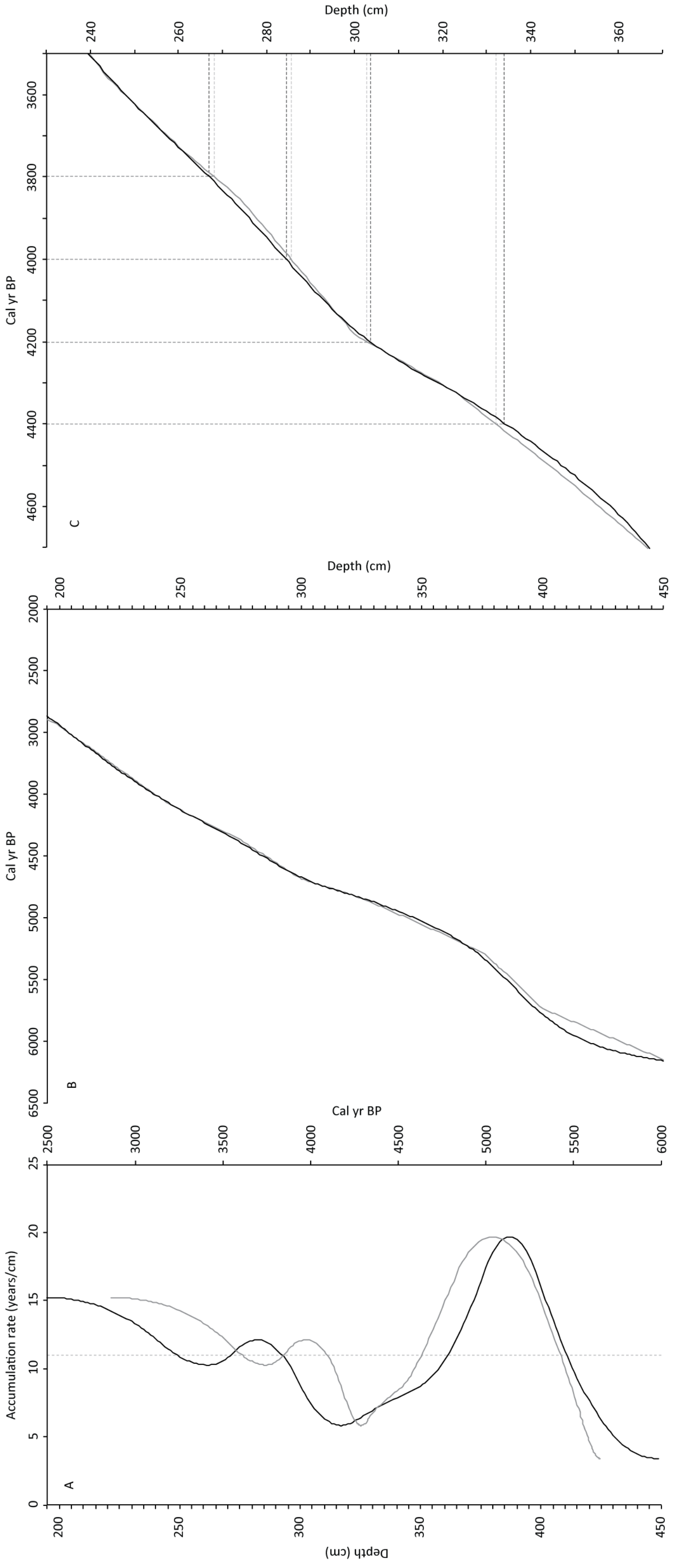
Table 6.14 Comparison of estimated depths for ages associated with the '4.2 kyr event' from the *OxCal P_Sequence* and *Clam* smooth spline models developed for Fallahogy Bog.

Age (cal yr BP)	***Clam* smooth spline estimate (cm)**	***OxCal P_Sequence* estimate (cm)**	**Difference (cm)**
3800	267	268.5	1.5
4000	284.5	285.5	1
4200	303.5	302.5	1
4400	334.5	332.5	2
		Average	1.375

6.5.3 Summary

Ultimately, the optimal age-depth model for a given sequence varies greatly according to its stratigraphy and the number of radiocarbon dates available. After careful consideration, the smooth spline models from both Sluggan Moss and Fallahogy Bog were deemed most appropriate. This uniformity between the two sites' age models also affords greater consistency to any correlations made between the palaeoecological and palaeoclimatological data from these two sites (see Section 2.8). For the purposes of the production, and subsequent discussion, of plots of these data against age, a single 'best' age is often required (Blaauw, 2010). The problems associated with the designation of point estimates to calibrated radiocarbon age ranges have already been discussed at great length (see Section 6.3.4). However, *Clam* provides a more robust alternative by approximating this 'best' point estimate by the weighted mean of all age-model iterations at each depth, through the length of the model. Extrapolation from the model is to be avoided where possible, and has only been conducted over very short lengths at each site (see Sections 6.5.1.3 and 6.5.2.3). Palaeoecological and palaeoclimatological data can now be plotted against these 'best' point estimates, where appropriate, in Chapters 7, 8 and 9.

Figure 6.24 a) Accumulation rates at Fallahogy Bog based on the *Clam* smooth spline model, plotted against depth (black) and age (grey), with average accumulation rate (dashed); b) Comparison of the *OxCal P_Sequence* and *Clam* smooth spline models developed for Fallahogy Bog. Only best point estimates are shown for each model, rather than full age ranges, as it is against these that palaeohydrological data will be plotted (see Section 6.5.3); c) Comparison of the *OxCal P_Sequence* and *Clam* smooth spline models developed for Fallahogy Bog for the period 4700 – 3500 cal yr BP.



Chapter 7 - Results

7.1 Introduction

This chapter presents plant macrofossil, humification, testate amoebae and stable isotopic data from Sluggan Moss and Fallahogy Bog. Results are grouped by site, and then by palaeoecological proxy, with stable isotope data being presented last. Palaeoecological data are described individually, interpreted for changes in BSW/WTD and finally presented in a multi-proxy format. Similarly, isotopic data are described both individually and within the context of the multi-proxy palaeoecological data. Extensive comparison, climatic interpretation and methodological evaluation of these data continues in Chapter 8.

In this chapter, all data are plotted against depth, with the exception of raw plant macrofossil and testate amoebae diagrams, which include a secondary age axis. These axes are developed using 'best' point estimates of smooth spline age-depth models described in Chapter 6 (see Section 6.5.3). In addition, figures presenting proxy-based BSW/WTD reconstructions are formatted in such a way that upward shifts represent changes towards drier and/or warmer climatic conditions, whereas downward shifts represent changes towards wetter and/or cooler conditions

7.2 Sluggan Moss

7.2.1 Plant macrofossils

Plant macrofossil data from Sluggan Moss are presented in Figure 7.1 against depth and with a secondary age axis. Plant macrofossil zones were identified using a combination of the CONISS numerical method within TILIA, and judgement by eye (Gordon and Birks, 1972; Bennett, 1996). Humification and testate amoebae data will also be discussed in relation to these plant macrofossil zones (cf. Hughes *et al.*, 2006). Table 7.1 offers a summarised botanical description of each plant macrofossil zone, together with its approximate depth and age.

Variation in the floral composition at both Sluggan Moss, described here, and Fallahogy Bog (see Section 7.3.1) is typical of raised, ombrotrophic bogs in the northwest Europe (e.g. Barber *et al.*, 1994; Hughes *et al.*, 2000). The sequences are characterised by fluctuations between Sphagna, which frequently dominate, and monocotyledon and ericaceous species.

Of the three *Sphagnum* species present in the Sluggan Moss sequence, *Sphagnum austinii* is the main peat-forming species at this site. This is consistent with other regional peat records in which the oceanic *Sphagnum* species typically colonised bogs in the mid-Holocene during known periods of cooler and/or wetter climatic conditions (Hughes *et al.*, 2007, see Section 8.2.1.1). The total *Sphagnum* component, dominated by *S. austinii*, is periodically interrupted by higher proportions of *Sphagnum* section *Acutifolia*. Two of these interruptions are particularly notable. The first notable peak in *S.* s. *Acutifolia* occurs within the basal zone in this sequence, SM-a. This zone is dominated in large part by monocotyledons, but the subsequent transition to the plant macrofossil zone above does not represent a fen-bog transition, thus suggesting that this event took place prior to 6500 cal yr BP. Species indicative of fen conditions, including *Menyanthes trifoliata*, *Phragmites australis* and *Equisetum* sp. are all absent (Hughes *et al*., 2000). In addition, *Sphagnum palustre*, thought to be an engineer or pioneer species of *Sphagnum* ecosystems, is also absent from the record (Hughes and Barber, 2004). The second notable occurrence of *S.* s. *Acutifolia* occurs in the top zone of the sequence, SM-g. Here *S.* s. *Acutifolia,* which is typically considered a drier species, occurs largely in the absence of ericaceous and monocotyledon species, but also with little or no UOM present, thus suggesting a wetter environment.

In addition to the interchange between *S. austinii* and *S.* s. *Acutifolia*, a notable peak in *Sphagnum* section *Cuspidata* occurs in SM-c, indicating a water table at or around the bog surface, and suggesting the development of a wet microform, most likely a hollow. Overall, variations in the dominance of the *Sphagnum* component in the sequence, as well as the transitions between *Sphagnum* species within total *Sphagnum*, often occur in an abrupt fashion, and suggest that the family, and the species within it, operate via threshold response to variability in the water table. However, the rapid accumulation rates associated with some Sphagna (e.g. *S. austinii*) could account for some of the rapidity seen during these transitions.

The importance of considering the entire species assemblage when inferring climatic variation from a plant macrofossil record is emphasised in Section 3.4.1.1. This is especially true when a sequence is dominated, as it is here, by a hummock-forming species, such as *S. austinii.* Total ericaceous and monocotyledon proportions are largely dominated by root material which is unidentifiable to species level, although small amounts of identifiable *Calluna vulgaris*, *Erica tetralix*, *Eriophorum vaginatum*, *Eriophorum angustifolium* and *Trichophorum cespitosum* were all identified. Generally, ericaceous and monocotyledon remains occur during periods of reduced *Sphagnum* dominance, and independent of one another. Notwithstanding the uppermost zone, SM-a, in which proportions of both are dramatically reduced, ericaceous remains maintain a fluctuating but generally constant presence. Monocotyledon remains, however, experience a more general long-term decline from high initial levels in SM-a to very low levels in SM-h.

The plant macrofossil data from Sluggan Moss were subject to DCA in an effort to produce a quantitative proxy-climate index (cf. Barber *et al*., 1994). At both sites in this study, neither transformation nor down-weighting of rare species was required to achieve satisfactory axis one eigenvalues of close to, or greater than 0.5 for the plant macrofossil data (ter Braak, 1995, see Section 5.4.1.1). Figure 7.2 shows the distribution of species scores against a biplot of DCA axes one and two from Sluggan Moss. Initial inspection of the distribution of species scores along axis one, indicates the presence of a clear water table gradient within the data. The eigenvalue of axis one (0.6145) is also strong, supporting any subsequent interpretations of changing BSW based upon it.

Modern examples of *S. austinii* have been shown to be capable of growing in relatively xeric conditions (Mauquoy and van Geel, 2006). However, the ability of *S. austinii* to grow under wetter bog surface conditions within the fossil records has also been demonstrated (Casparie, 1972; van Geel, 1978; Barber, 1981). Subsequently, the position of *S. austinii* in a relatively central position on the wetness gradient of axis one, albeit towards the ‘drier’ end of the gradient, is unsurprising.

Problems associated with the microscopic identification of certain Sphagna to species level, resulting in the grouping of taxa in sections, can influence the efficacy of palaeoclimatic inferences taken from plant macrofossil data, and are demonstrated well here. Subfossil leaves of members of *S.* s. *Acutifolia* and *S.* s. *Cuspidata* are difficult to identify to species level (Smith, 2004; Wimble, 1986). However, *S.* s. *Sphagnum* (e.g. *S. austinii*) is often identifiable to species level (Mauquoy and Barber, 1999).

The *S.* s. *Cuspidata* species commonly found in the northwestern Europe (e.g. *S. cuspidatum*, *S. recurvum*, *S. balticum*, *S. majus*, *S. pulchrum*; Daniels and Eddy, 1990) all dominate during period of high water tables (Rydin, 1993), and can therefore be simply interpreted as indicative of pool and low lawn ecotopes (Mauquoy and van Geel, 2006). In northwest Europe peatlands, subfossil material, when identifiable, has been assigned as *S. cuspidatum* (e.g. Casparie, 1972; van Geel, 1978; Barber, 1981; Mauquoy, 1997) and it is reasonable to expect that this taxon forms the majority of subfossil *S.* s. *Cuspidata* identified here. *S. cuspidatum* is a pool species, but compact forms have been demonstrated to grow just above the water level (e.g. Ratcliffe and Walker, 1958). As a result, small amounts of *S.* s. *Cuspidata* should be interpreted cautiously, as they may not indicate a wet phase. However, large amounts of subfossil *S.* s. *Cuspidata* would typically indicate the presence of a pool environment. The Section's position at the 'wetter' end of axis one supports the presence of a hydrological gradient in the DCA at Sluggan Moss.

Conversely, *S.* s. *Acutifolia* species commonly found in northwestern Europe have been shown to possess a bimodal distribution, consisting of both hummock (*S. capillifolium* var. *capillifolium*, *S. capillifolium* var. *rubellum*, *S. fuscum*; Daniels and Eddy, 1990*)* and lawn (*S. subnitens*, *S. molle;* Daniels and Eddy, 1990). Common species in the section are difficult to identify beyond section level in a subfossil state (Stoneman, 1993; Mauquoy, 1997). In this DCA, *S.* s. *Acutifolia* is positioned towards the wetter end of the gradient, suggesting the presence of hygrophilous taxa. However, the section is typically considered indicator of drier conditions in peat-based palaeoclimatic studies in the region (e.g. Dupont, 1986; Daley and Barber, 2012), thus highlighting the potential problems associated with its climatic interpretation. In instances where this section dominates, we must consider the coincident taxa, and any other available palaeoecological information, before drawing any climatic inference.

Species associated with dry conditions, such as the Ericaceae family, including *Calluna vulgaris* and *Erica tetralix*, alongside other dry indicators such as *Trichophorum cespitosum* (Mauquoy and van Geel, 2006) are concentrated on axis one with values above 2. In addition, UOM, which is typically associated with ericaceous material and therefore interpreted as an indicator of dry conditions created by high levels of humification, also falls within this group.

Eriophorum vaginatum is a species capable of growing over a range of moisture conditions, hence its position towards the centre of the wetness gradient. Similarly, *Eriophorum angustifolium* is characteristic of drying pools, and its position, further than *E. vaginatum*, towards the 'wetter' end of the gradient, is therefore reasonable (Mauquoy and van Geel, 2006).

Taphonomic issues are not confined to moss remains, however, and concerns have been raised over the overrepresentation of decay-resistant *Eriophorum* roots and *Calluna* rootlets in the macrofossil record (Heal *et al*., 1978). Indeed, non-bryophytic remains in the records from both sites are heavily dominated by undifferentiated ericaceous and monocotyledonous root material and must be interpreted carefully, and in conjunction with data of identified plant remains belonging to the monocotyledon group and Ericaceae family.

Figure 7.3 shows the distribution of sample scores against a biplot of DCA axes one and two from Sluggan Moss (cf. Booth and Jackson, 2003). In general, samples from the same plant macrofossil zone, as calculated using CONISS cluster analysis, group together well, further supporting both the interpretation of the DCA axis one as a valid ecological gradient, but also the application of plant macrofossil zones. CONISS is stratigraphically constrained and therefore some overlap exists between zones. However, the diagram also emphasizes the uniqueness of the species composition experienced in the *S*. s. *Acutifolia* dominated zone, SM-h.

Figure 7.4a displays axis one sample scores against depth for Sluggan Moss, operating as a proxy-climate index and reconstruction of past changes in BSW. By comparing this curve to the plant macrofossil data seen in Figure 7.1, it becomes apparent that the DCA is driven heavily by interchange between species of *Sphagnum*, with noticeable wet periods occurring simultaneously with peaks in *S*. s. *Acutifolia* (316 cm, SM-a) and *S*. s. *Cuspidata* (260 cm, SM-c). In addition, the substantial shift towards *S*. s. *Acutifolia* dominance that occurs in SM-h, above 126 cm is shown by DCA results to be a particularly wet period.

The plant macrofossil data from Sluggan Moss were also transformed using the DHI method (Dupont, 1986) into a quantitative proxy-climate index of past changes in BSW. Figure 7.4b plots the results of DHI analysis, with Figure 7.4c plotting both the DCA and DHI curves against depth to aid comparison. DHI analysis exhibits increased variability within the plant macrofossil data than is shown by DCA. Both curves operate in relative agreement during zones SM-a (344 – 298 cm) and SM-c (274 – 250cm). However, their relationship during SM-b is characterised by poor agreement and potential anti-phase. Zones SM-d to SM–g (250 – 126 cm) are characterised by a moderate agreement but generally greater variability and with occasional poor agreement. This increased variability could be explained by a higher sensitivity of the DHI method to variation in *Sphagnum* species, than vascular species, as the technique was originally intended for use in *Sphagnum* dominated, late-Holocene peats (Dupont, 1986). However, the upper, *S*. s. *Acutifolia* dominated zone, SM-h, is not suggested by DHI analysis to be significantly wetter or drier than the average wetness of the studied sequence. Again, *S*. s. *Acutifolia* is assigned a relatively high Dupont weighting of 6, which is equivalent to *Trichophorum cespitosum*, for example. As DCA shows, in this record *S*. s. *Acutifolia* tends to be associated with wetter bog surface conditions. It could be argued that the DHI is too rigid in the way it incorporates species sections such as *S*. s. *Acutifolia* as some species within the section are more hygrophilous than others and, if identifiable, would be assigned lower Dupont weightings. Whilst DHI is broadly coherent, in comparison with DCA results and the botanical assemblages described in Table 7.1, it was felt that the flexibility and subsequent strong axis one eigenvalue was sufficient to justify using DCA as the primary basis of plant macrofossil

BSW reconstruction at Sluggan Moss. Daley and Barber (2012) argue in favour of wider use of DHI in plant macrofossil-based palaeoclimate studies, as it performs well in instances where DCA results are adversely effected by 'crossed gradients', such as the fen-bog transition. No such crossed-gradients are experienced here, and DCA is more broadly applied in the peat-based palaeoclimatic literature, further justifying its use as a summary tool. The agreement between the DHI and DCA reconstructions is good enough, however, to provide further support for any interpretations of changes in BSW based on the DCA results and species composition outlined in Figure 7.1 and Table 7.1.

7.2.2 Peat humification

Raw percentage light transmission data, obtained from contiguous 1 cm samples through the Sluggan Moss sequence, are presented in Figure 7.5a. Here, a linear regression line has been fitted to demonstrate the long-term down-core trend towards a higher degree of humification, reflecting the continued decomposition of peat matter in the catotelm, and the subsequent increase in the presence of humic acids (Clymo, 1984; Barber and Langdon, 2001). As a result, the humification results were detrended, and normalised to enable comparison with other proxy data derived from this sequence (Blundell and Barber, 2005). These detrended and normalised results are plotted against depth in Figure 7.5b. Three notable shifts towards higher transmission values and lower degrees of humification, indicative of wetter and/or cooler climatic conditions, occur between 274-262 cm, 207-195 cm and 128-114 cm. All three shifts correspond well with the onset of the plant macrofossil zones SM-c (begins 274 cm), SM-e (begins 210 cm) and SM-h (begins 126 cm).

7.2.3 Testate amoebae

Testate amoebae data from Sluggan Moss are presented in Figure 7.6, against depth and age respectively. These data are discussed in relation to the plant macrofossil zones (see Section 7.2.1, cf. Hughes *et al*., 2006). Description of the varying faunal composition, discussed by zone, is presented in Table 7.2.

The studied sequence at Sluggan Moss is overwhelmingly dominated by the interchange between the two dominant species, *Amphitrema flavum* and *Difflugia pulex*. *A. flavum* is generally considered to be a species indicative of wetter bog surface conditions. Conversely, *D. pulex* occupies a relatively dry position in the current pan-European transfer function (Charman *et al*., 2007). Therefore, the interchange between the two species could be interpreted, in large part, as shifts from wetter to drier conditions, and vice versa. However, as discussed in Section 3.4.3.2, there are uncertainties associated with the water table tolerance of *D. pulex*, owing to a lack of modern analogues in surface samples.

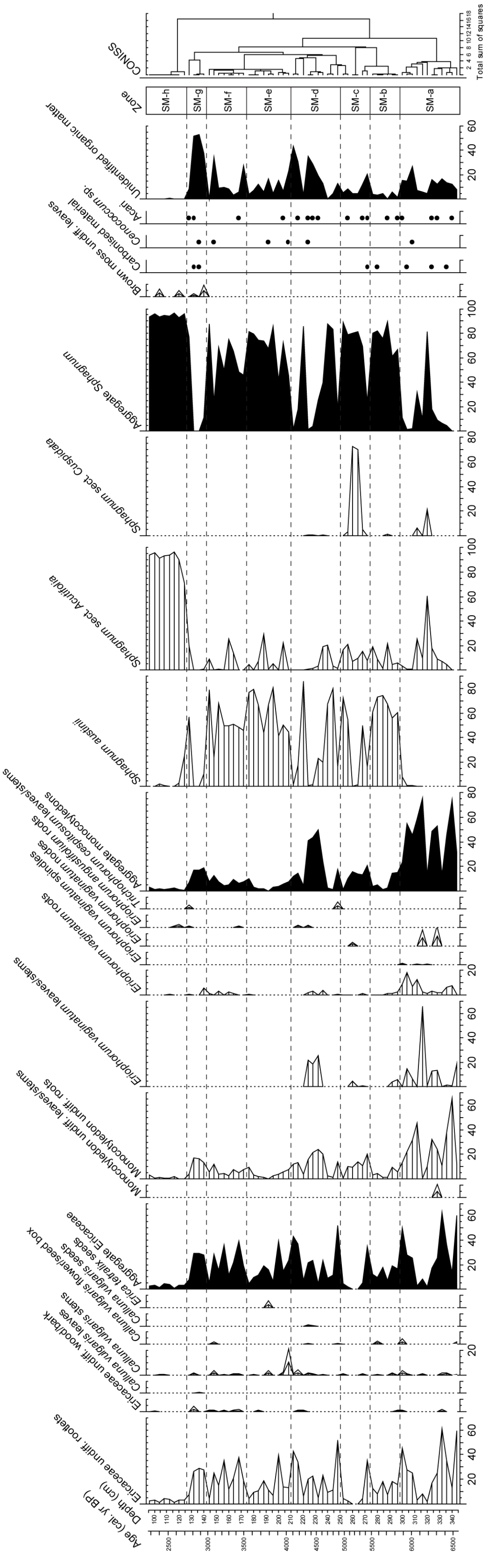


Figure 7.1 Plant macrofossil diagram for Sluggan Moss plotted against depth, with secondary age axis. Peat components displayed with depth bars are derived from averaged quadrat counts (n=15) under low-power magnification (x10). Peat components displayed in solid black denote aggregates of different components. Individual *Sphagnum* leaf counts are also displayed with depth bars, and consist of proportions of aggregate *Sphagnum* based on random selection of 100 leaves identified at high magnification (x400). All data are presented as percentages, with the exception of presence/absence data displayed as black dots. Zones were assigned after CONISS cluster analysis.

Table 7.1 Botanical description of plant macrofossil zones at Sluggan Moss.

Zone	Depth (cm)	Floral assemblage description	Age (cal yr BP)
SM-a	344–298	A variable zone dominated, more than any other, by monocotyledons, with all *Eriophorum vaginatum* components (including leaves/stems, roots/spindles and nodes) present. Ericaceous remains are sporadically dominant, when monocotyledon proportions are reduced, and are dominated by rootlet material. Small amounts of *Calluna vulgaris* are identified. A peak in the presence of *Sphagnum* section *Acutifolia* occurs during this zone, centring around 316 cm, with *Sphagnum* section *Cuspidata* occurring in lesser quantities but a similar pattern. *Sphagnum austinii* is largely absent from this zone until very small amounts appear at 312 cm. UOM is persistent throughout this zone at moderate levels.	6550 – 5850
SM-b	298–274	A zone dominated by *Sphagnum austinii* with consistently high percentages, supplemented with a consistent but lesser presence of *Sphagnum* section *Acutifolia*. Minimal amounts of *Sphagnum* section *Cuspidata* are also present. Monocotyledon remains are dramatically reduced from the previous zone, with small amounts identifiable as *Eriophorum vaginatum*. Ericaceous material persists, dominated by ericaceous rootlets, although *Calluna vulgaris* remains are identifiable. UOM is present in this zone but at low levels.	5850 – 5500
SM-c	274–250	Commencing with a brief and moderate peak in ericaceous material and UOM, this zone is otherwise dominated by *Sphagnum*. Ericaceous remains, occasionally identified as *Calluna vulgaris*, are reduced to their lowest proportion throughout the studied sequence. The zone begins and ends with *Sphagnum austinii* dominant, however, a clear and discrete peak in *Sphagnum* section *Cuspidata* dominates this zone, centring on 260 – 264 cm, and is unique within the studied sequence. Monocotyledons remain present, with *Eriophorum vaginatum* identified in small amounts, but overall proportions are relatively low.	5500 – 4900
SM-d	250–210	A variable zone with dominance shifting between alternating peaks of *Sphagnum*, ericaceous and monocotyledon remains. Two moderate peaks of ericaceous remains, mainly rootlets with some *Calluna vulgaris* identifiable, occur at the beginning and end of the zone. A sizeable peak of monocotyledon remains, largely *Eriophorum vaginatum* leaves and stems, then occurs, bracketed between two notable peaks of *Sphagnum austinii*. Small amounts of *Eriophorum angustifolium* were also identified in the latter half of the zone. A small peak in *Sphagnum* section *Acutifolia* is broadly coherent with the first *Sphagnum austinii* peak, before the former species disappears from the record in the latter half of the zone. Extremely small amounts of *Sphagnum* section *Cuspidata* are present. Small amounts of *Trichophorum cespitosum* leaves and stems are also present at the beginning of this zone. In the latter half of the zone, two major peaks of UOM are evident, occurring in alternation with peaks in *Sphagnum* species.	4900 – 4100
SM-e	210–174	A zone dominated by *Sphagnum austinii* with an underlying presence of *Sphagnum* section *Acutifolia*, culminating in two small peaks. Monocotyledon remains are reduced greatly from the previous zone, with no material identifiable to species level. Ericaceous material generally declines towards the end of the zone from an initial peak, including moderate amounts of *Calluna vulgaris* stems. This zone also possesses small amounts of *Erica tetralix*. UOM is present at steady, low levels throughout the zone.	4100 – 3550
SM-f	174–142	A zone generally dominated by *Sphagnum austinii* with an underlying presence of ericaceous material, marginally increased from the previous zone. A moderate spike of *Sphagnum* section *Acutifolia* occurs centrally in the zone. Small amounts of this ericaceous material are identified as Calluna vulgaris, with the majority being rootlet remains. Monocotyledon material maintains a consistent, but low presence, with a small proportion of root material being identified as *Eriophorum vaginatum*. Very small amounts of *Eriophorum angustifolium* were also identified at the beginning of the zone. The presence of UOM is characterised by two moderate peaks at the beginning and end of the zone.	3550 – 3000
SM-g	142–126	A zone dominated by UOM and total *Sphagnum* disappearing from the record almost entirely, with only small amounts of *Sphagnum austinii* and *Sphagnum* section *Acutifolia* present at the beginning and end of the zone. Moderate amounts of ericaceous material are present, dominated by rootlets with some *Calluna vulgaris* identified. Moderate amounts of monocotyledons are also present, with remains of *Eriophorum vaginatum*, *Eriophorum angustifolium* and *Trichophorum cespitosum* all present within the zone. Brown mosses emerge in small quantities for the first time in the sequence.	3000 – 2700
SM-h	126–96	A zone dominated by *Sphagnum* section *Acutifolia*, with UOM reduced to zero. Proportions of *Sphagnum austinii* are reduced to zero from an initial peak separating this zone from the previous. Monocotyledons are consistently present, but reduced to their lowest proportions in the studied sequence, extremely small quantities of both *Eriophorum vaginatum* and *Eriophorum angustifolium* roots identified. Ericaceous remains also persist, but also at very low proportions, representing their lowest occurrence in the studied sequence. Brown mosses are also sporadically found in small quantities.	2700 – 2300

Figure 7.2 DCA scatter plot for plant macrofossil data at Sluggan Moss, showing species axes one and two scores.

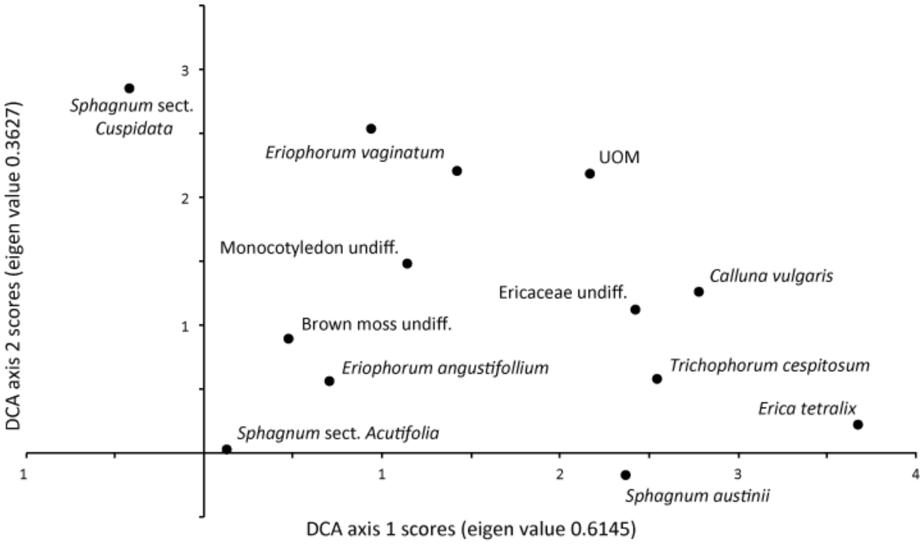


Figure 7.3 DCA scatter plot for plant macrofossil data at Sluggan Moss, showing samples axes one and two scores. Samples are also colour-coded according to CONISS plant macrofossil zone (see Fig. 7.1).

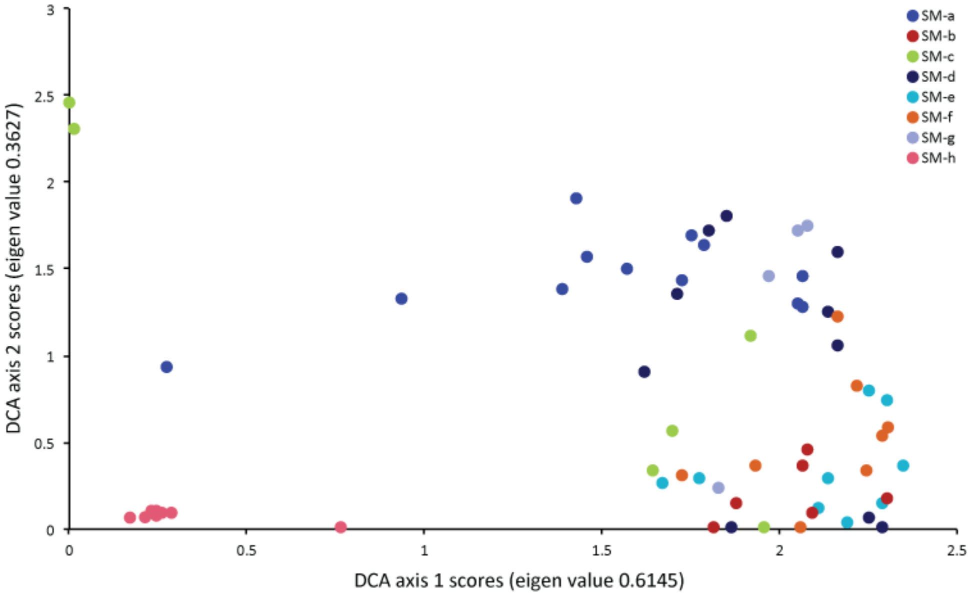


Figure 7.4 Plant macrofossil data from Sluggan Moss displayed as: a) DCA axis one sample scores; b) DHI sample cores; c) comparison of DCA (solid) and DHI (dashed) scores after data normalisation. All curves are plotted against depth.

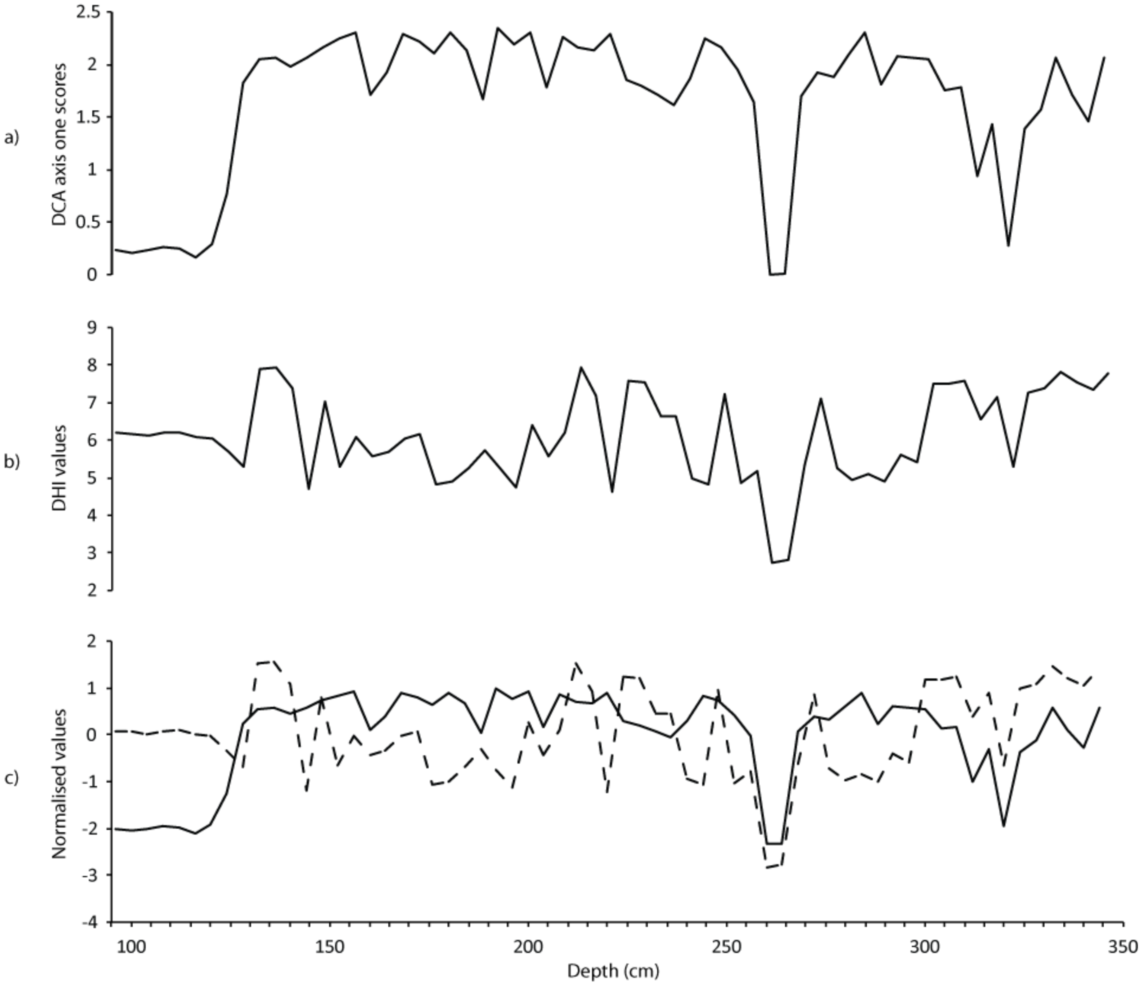
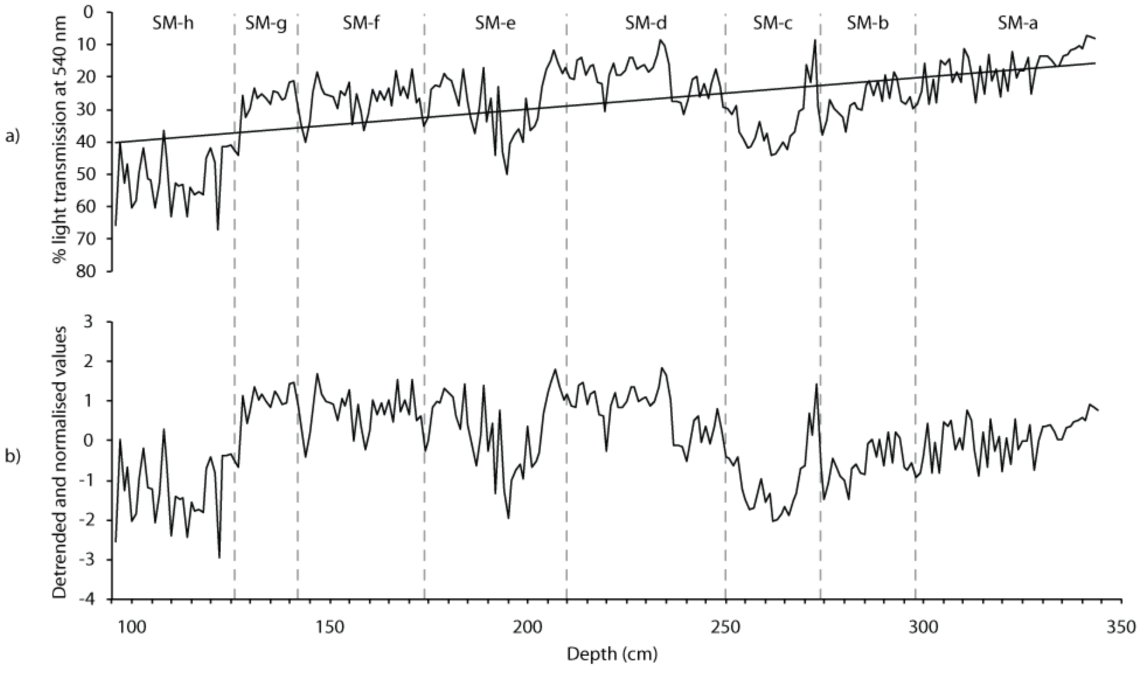


Figure 7.5 Humification data for Sluggan Moss, displayed as: a) 'Raw' values for 540 nm light transmission through humic acid extraction in solution; b) Detrended and normalised values. Both curves are plotted against depth.



Fossil records often demonstrate a dominance of the species downcore (e.g. Gearey and Caseldine, 2006), and it has been suggested anecdotally that the species is heavily associated with *Sphagnum austinii*, with the broad-leaved moss acting as a 'host' for *D. pulex*. Bobrov *et al*. (1999) speculated on the possibility of a link between host plant architecture and testate amoebae, but the idea remains untested. Subsequently, it is possible that the well-documented decline of *S. austinii* in the late-Holocene (Hughes *et al*., 2007) could explain the lack of *D. pulex* in modern samples. These issues are discussed in further detail in Section 8.2.3 and 9.6.1.3.

Alongside *A. flavum* and *D. pulex* there are a number of other species which feature notably in this sequence. *Assulina muscorum*, *Assulina seminulum*, *Cyclopyxis arcelloides* and *Difflugia pristis* are all species associated with wide hydrological tolerances. *Hyalosphenia subflava* and *Trigonopyxis arcula*, however, are generally associated with lower water tables and drier bog surface conditions. On initial inspection variations in the faunal composition of the testate amoebae data at Sluggan Moss are in good agreement with the plant macrofossil zones, initially suggesting that the testate amoebae and bog plants are responding to a common driving mechanism. The raw testate amoebae assemblage data, however, do not demonstrate complete coherence with the plant macrofossil record here. The proxies are in good agreement in some zones, for example, in SM-c a dramatic peak in wet indicator *A. flavum* occurs coincidently with a near disappearance of dry indicator, *D. pulex*. This is consistent with a peak in *S.* s. *Cuspidata* in the plant macrofossil record. However, incoherence is also evident between the two proxies. The first example takes place in SM-a, a zone that could be interpreted as forming within a relatively dry period, as the plant macrofossil record is dominated by monocotyledon and ericaceous species, including identifiable remains of dry indicator *C. vulgaris*, as opposed to species of *Sphagnum*. In the testate amoebae record, however, *A. flavum* is abundant, with *D. pulex* present in relatively low concentrations, when compared to the rest of the studied sequence. Conversely, the uppermost zone, SM-h is shown by plant macrofossil DCA to be a period of distinct wetness, with an almost complete absence of UOM, and ericaceous and monocotyledon species. The testate amoebae record for this zone, clearly demonstrates a period of drier conditions, with a dominance of *D. pulex* being supported by the first significant emergence of dry indicator *Hyalosphenia subflava*. The issues associated with difficulties identifying *S.* s. *Acutifolia*, which dominates this upper zone, to species level have been discussed. It could be suggested that without identifying which *S.* s. *Acutifolia* species is present, the hydrological nature of this zone may remain ambiguous, due to conflicting palaeoecological evidence.

The testate amoebae data from Sluggan Moss were subject to DCA to produce a quantitative proxy-climate index (cf. Barber *et al*., 1994), enabling comparison with other quantitative palaeoecological reconstructions produced by humification and plant macrofossil analyses. At both sites in this study, no transformation was performed, but down-weighting of rare species was required to achieve the highest possible axis one eigenvalues for the testate amoebae data (ter Braak, 1995). The high biodiversity of this record, when compared to the plant macrofossil record from the same core, meant that rare species down-weighting was beneficial as it removed the detrimental effects of rare species, which may have artificially skewed the results.

Figure 7.7 shows the distribution of species scores against a biplot of DCA axes one and two from Sluggan Moss. Initial inspection of the distribution of species scores along axis one, indicates that although the eigenvalue is of only moderate strength (0.2819) there is potential evidence for the presence of a water table gradient within the data. In general, the species dominant in this studied sequence occupy coherent positions, relative to one another along axis one. For example, hygrophilous taxa, *A. flavum* and *A. wrightianum*, are both above 1 on the gradient, intermediate species *A. muscorum*, *A. seminulum* and *D. pristis* fall between 0 and 1, with drier indicators *H. subflava*, *D. pulex* and *T. arcula* all below 0. There are, however, a number of inconsistencies in species placement along the gradient. *Bullinularia indica* and *Arcella catinus*, two species typically associated with very dry conditions are both assigned values above 1 in this DCA. Similarly, *Arcella discoides*, typically interpreted as hygrophilous taxa, falls below 0 on the first axis.

Figure 7.9a displays axis one sample scores against depth. Despite the inconsistencies discussed above, the curve suggests that it is operating as a proxy-climate index and reconstruction of past changes in BSW. By comparing this curve to the testate amoebae data seen in Figure 7.6, it becomes apparent that the DCA is heavily influenced by the interchange between A. *flavum* and *D. pulex*, and generally, the DCA curve is in agreement with interpretations based on the raw data. In addition, there is strong agreement between the DCA axis one scores and the WTD reconstruction, using the ACCROTELM pan-European transfer function (Fig. 7.9b; Charman *et al*., 2007), after data normalisation, further supporting the interpretation of the axis as a hydrological gradient (Figure 7.9d).

Figure 7.8 shows the distribution of sample scores against a biplot of DCA axes one and two from Sluggan Moss. Whilst there is significant overlap between samples from different plant macrofossil zones, there does appear to be some degree of structure, to some extent supporting the earlier assertion that testate amoebae and plant macrofossils are responding to common driving mechanisms.

These data from Sluggan Moss were also transformed using a version of the DHI method (Dupont, 1986) modified here for use with testate amoebae (see Section 5.4.3.3), into a quantitative proxy-climate index of past changes in BSW. Figure 7.9c displays the results of DHI analysis with species scores plotted against depth. By comparing this curve to the raw testate amoebae assemblage data seen in Figure 7.6, the application of a modified DHI analysis can initially be considered a success, as the DHI curve is consistent with interpretations based on the raw data. As with the plant macrofossil analyses from this site, DHI results exhibit marginally more variability than DCA results after normalisation (Figure 7.9d), although the two are in generally good agreement.

Furthermore, Figure 7.9d plots the normalised results of DHI and DCA analysis alongside the transfer function water table reconstruction and against depth to aid comparison. This demonstrates a good level of agreement between all three statistical techniques. However, DCA is not as broadly applied to testate amoebae, when compared to plant macrofossils, in peat-based palaeoclimatic studies and this, together with a mediocre axis one eigenvalue of 0.2819, suggests that the reconstruction should not be used as the primary summary for the testate amoebae data.

Furthermore, the use of the DHI method for testate amoebae is also not widely adopted, and no known example of its use can be found in the current literature. Its inclusion in this study was intended as an exercise to ascertain the potential for the development of the method for future studies, as the high-levels of biodiversity often witnessed in testate amoebae records, and their associated ecological complexity, can preclude the production of a robust axis one from DCA. Instead, the reconstruction of past changes in WTD from testate amoebae using transfer functions is widely accepted as the optimal method, with the ACCROTELM transfer function (Charman *et* al., 2007) employed here frequently preferred in the literature (see Section 3.4.3.1). Subsequently, the transfer function reconstruction is adopted as the primary quantitative summary tool for the testate amoebae data from Sluggan Moss as it is both robust and strengthens any inter-site comparisons with existing testate amoebae records. The transfer function WTD reconstruction exhibits a number of potential wet shifts commencing at 344, 276, 222, 192, 148 and 132 cm.

Figure 7.10 presents a comparison of the pan-European ACCROTELM (Charman *et al*., 2006) and regionally specific North of Ireland (Swindles *et al*., 2009) transfer function water table reconstructions (see Section 5.4.3.2). The two reconstructions demonstrate a good degree of agreement, with the key shifts observed above present in both curves. The North of Ireland model does, however, appear to predict generally drier conditions throughout, with this being emphasised during certain periods, particularly those associated with species such as *H. subflava* (e.g. SM-h, 126 cm and above), and display generally a higher degree of variability than the ACCROTELM model.

7.2.4 Multi-proxy comparison of palaeoecological records

Figure 7.11 presents a summary of the proxy reconstructions from plant macrofossil, humification and testate amoebae analyses at Sluggan Moss, which have been normalised to aid comparison (cf. Blundell and Barber, 2005; Hughes *et al*., 2006). The plant macrofossil assemblage at Sluggan Moss is typical of an oceanic raised bog in the region (Hughes and Barber, 2004) and the presence of a *S.* s. *Cuspidata* suggests the bog is climatically sensitive (Barber *et al*., 1994), and there are a number of occasions during which the proxy summary curves (i.e. testate amoebae transfer function, plant macrofossil DCA and 5-pt moving average humification) are in broad agreement, displaying shifts towards wetter conditions, albeit with variation in the magnitude of the changes (Fig. 7.11). Of particular palaeoclimatic note is a multi-centennial wet period, during the period c. 275 – 245 cm, which is recorded in all three proxies. Similarly, there appears to be a significant wet shift c. 130 cm. However, it is worth nothing that there is also significant proxy disagreement within the studied sequence (e.g. c. 215 – 170 cm, 120 – 95 cm).

By discussing each proxy individually, and then comparing the three, a number of issues are raised with regard to the development of palaeoclimatic inference from these records. These include problems associated with: being unable to identify *Sphagnum* section *Acutifolia* to species level; the incorporation of species with bi-modal hydrological preferences within, and also of grouping species for the purposes of, DCA; the auto-correlative tendencies of the existing techniques for determining the degree of peat humification. Annotations included in Figure 7.11 attempt to outline the likely causes of these discrepancies, and these issues will discussed at greater length in Chapter 8.

Table 7.2 Faunal description of testate amoebae record at Sluggan Moss, by plant macrofossil zone.

Zone	Depth (cm)	Faunal assemblage description	Age (cal yr BP)
SM-a	344–298	This basal zone is characterised by a dominance of *Amphitrema flavum*. High levels of *Assulina muscorum* and *Assulina seminulum* are seen at the base of the zone, but reduce quickly to a constant but low presence. *Difflugia pristis* are present in moderate quantities with a noticeable peak occurring at 324 cm. *Difflugia pulex* increases through the zone in a staggered fashion with an initial peak at 328 cm, followed by a sharp decline, before resuming an upward trend to dominate the composition by the close of the zone.	6550 – 5850
SM-b	298–274	A zone dominated by *Difflugia pulex* with a reduction in the presence of *Amphitrema flavum*, troughing at 286 cm. Towards the top of the zone, *Amphitrema flavum* increases to dominate, with a sharp decline in *Difflugia pulex* to almost zero marking the end of the zone. *Assulina muscorum* persists in low but consistent proportions. *Difflugia pristis* maintain a low but persistent presence.	5850 – 5500
SM-c	274–250	A zone characterised by a dominance of *Amphitrema flavum*, with *Difflugia pulex* experiencing proportions close to zero for the majority of the zone. This zone represents *Amphitrema flavum*'s most notable and dominant peak in the studied sequence. *Assulina muscorum* is reduced to minimal levels, as is *Difflugia pristis*. Notably, *Amphitrema wrightianum* and *Arcella artocrea* are also present in this zone, albeit in small quantities.	5500 – 4900
SM-d	250–210	A zone of transition, which begins with a peak and associated trough in *Difflugia pulex* and *Amphitrema flavum*, respectively. This pattern is subsequently reversed before a longer-term decline in *Amphitrema flavum*. This decline operates in phase with a staggered increase in *Difflugia pulex*. *Assulina muscorum* is also present in notable proportions compared to previous zones. *Difflugia pristis* and *Assulina seminulum* persist at low levels. This zone also marks a prolonged period of *Cyclopyxis arcelloides* presence, which is unmatched anywhere else in the studied sequence.	4900 – 4100
SM-e	210–174	The boundary between this zone and the previous zone is characterised by a peak in *Amphitrema flavum* and a trough in *Difflugia pulex*. However, this event represents merely a diversion from this zone's overall trend, continued from the previous zone, which sees *Amphitrema flavum* decline and *Difflugia pulex* increase in proportion. *Difflugia pristis* persists at slightly higher levels than the previous zone. *Assulina muscorum* is present throughout, with a notable peak at 192 cm. *Amphitrema wrightianum* is also present in this zone, albeit in relatively low concentrations.	4100 – 3550
SM-f	174–142	This zone sees the end of the trend, which has persisted through the previous two zones, with *Amphitrema flavum* and *Difflugia pulex* troughing and peaking, respectively, at 168 cm. A sharp decline then occurs in *Difflugia pulex* back to moderate levels, which then persist until the end of the zone. A notable peak in *Amphitrema flavum* occurs in this zone peaking at 158 cm. This presence then falls again towards the end of the zone, with the taxa replaced by rising quantities of *Assulina muscorum*, *Assulina seminulum* and *Difflugia pristis*. Notably, *Trigonopyxis arcula* also persists at its highest levels for the studied sequence.	3550 – 3000
SM-g	142–126	This zone is characterised by a significant peak in *Difflugia pristis* at the expense of the previously dominant species *Amphitrema flavum* and *Difflugia pulex.* The presence of *Trigonopyxis arcula* from the previous zone continues, but with small peaks alternating with similar peaks in *Amphitrema wrightianum*. *Assulina muscorum* continues a presence at moderate levels, with *Assulina seminulum* persistent in low proportions.	3000 – 2700
SM-h	126–96	The uppermost zone is dominated by *Difflugia pulex*, with low levels of *Amphitrema flavum* relative to the rest of the studied sequence. *Assulina muscorum* and *Assulina seminulum* continue as they did in the previous zone, and *Difflugia pristis* is reduced to levels approaching zero. Notably, *Hyalosphenia subflava* emerges to its highest persistent level in the sequence.	2700 – 2300

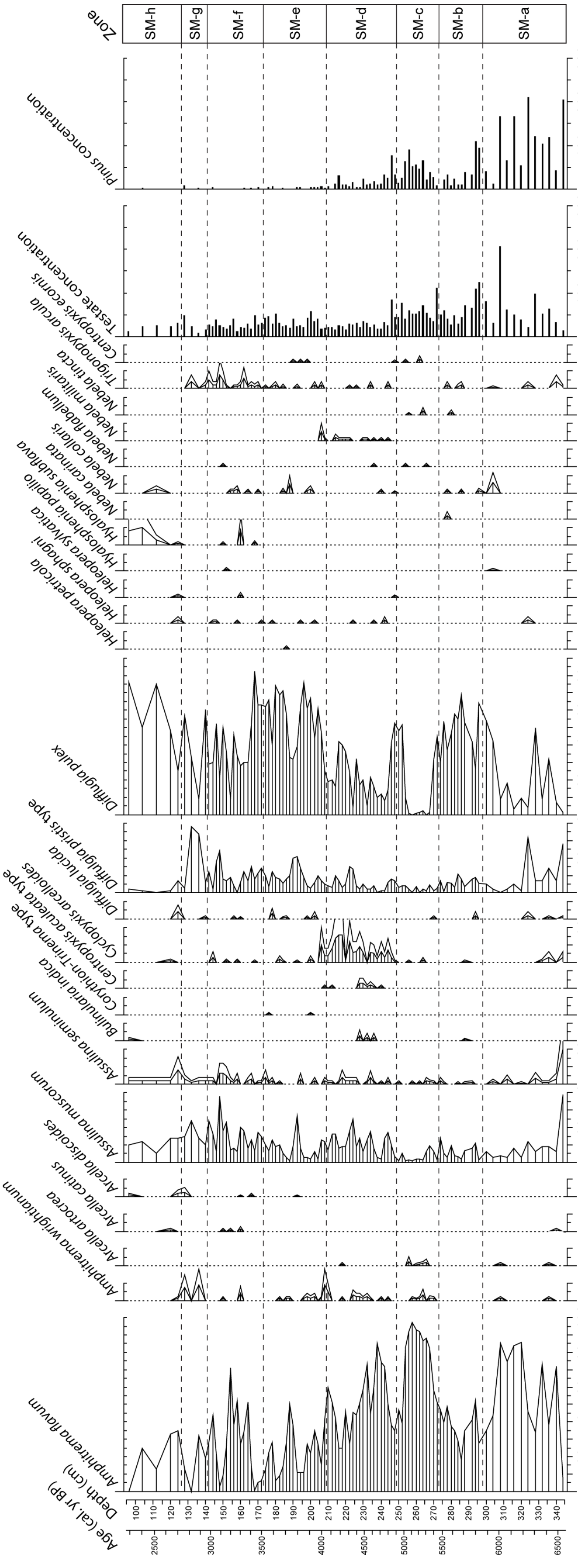


Figure 7.6 Testate amoebae diagram for Sluggan Moss plotted against depth. All data are percentages of the total number counted per level. Concentration data are displayed in bar form. Zones follow CONISS cluster analysis of plant macrofossil data (see Fig. 7.1).

Figure 7.7 DCA scatter plot for testate amoebae data at Sluggan Moss, showing species against axes one and two.

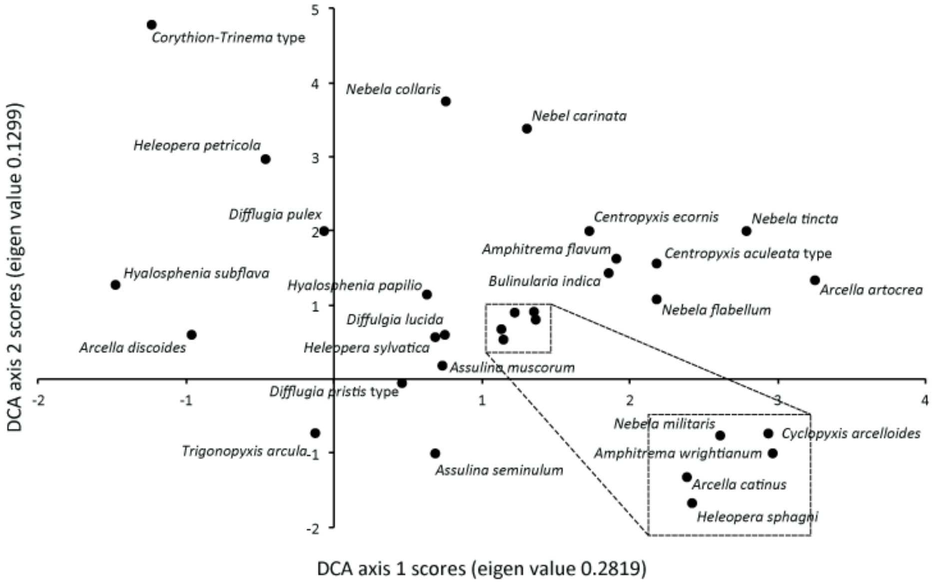
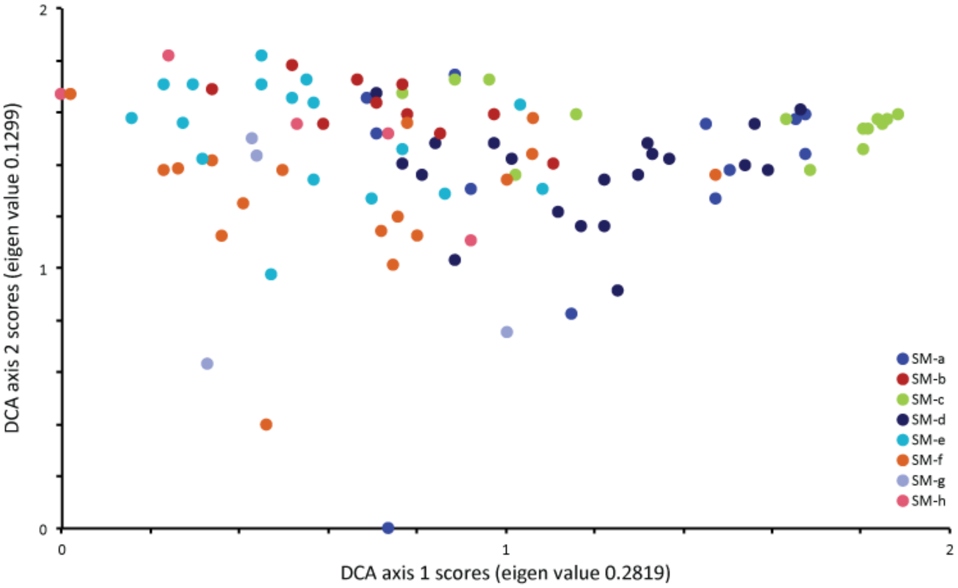


Figure 7.8 DCA scatter plot for testate amoebae data at Sluggan Moss, showing samples against axes one and two. Samples are also colour-coded according to plant macrofossil CONISS zones.



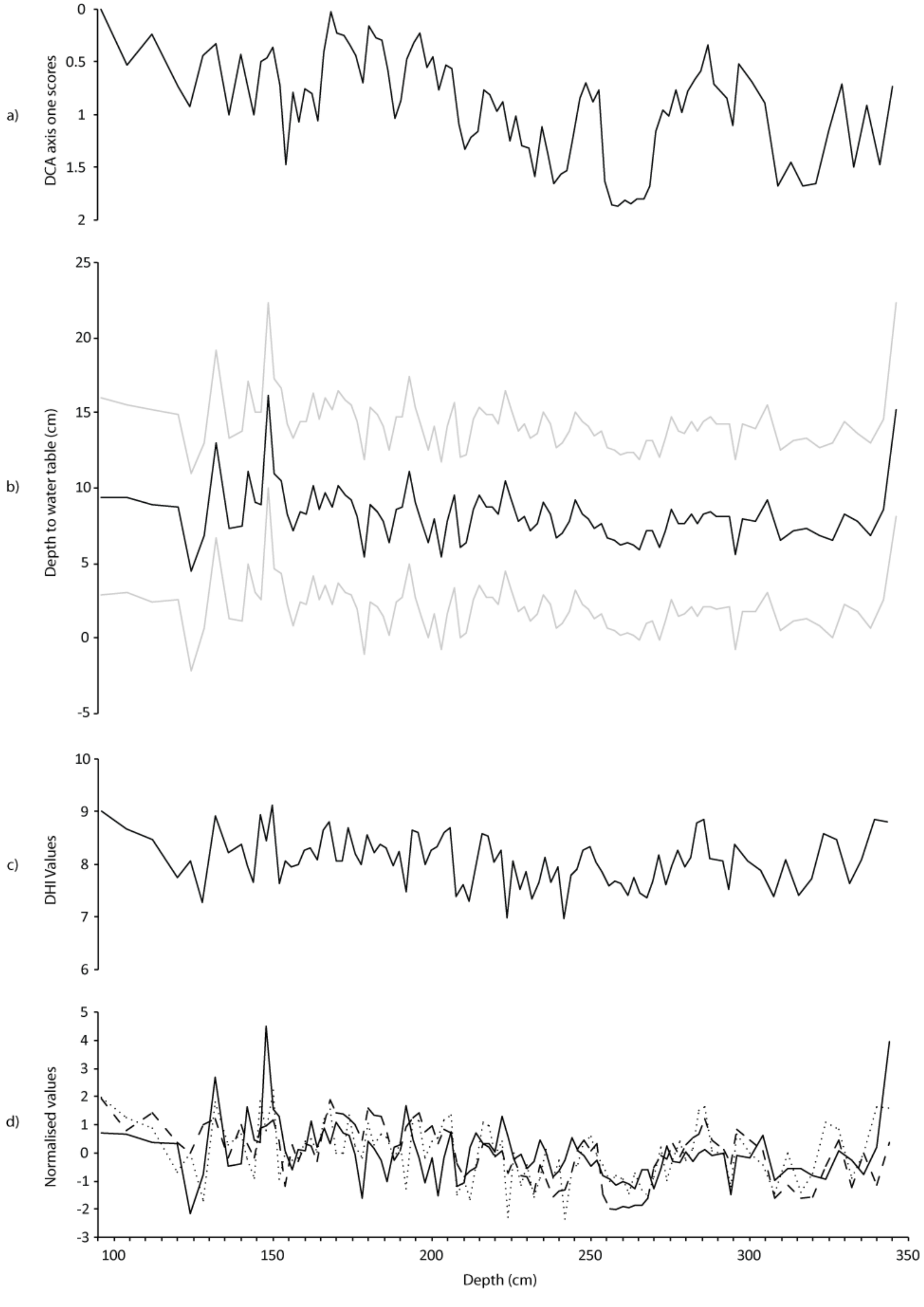


Figure 7.9 Testate amoebae data from Sluggan Moss displayed as: a) inferred WTD reconstructions based on the ACCROTELM pan-European transfer function (Charman *et al*., 2007) with calculated error displayed in light grey; b) DCA axis one sample scores; c) testate amoebae-adapted DHI sample cores; d) comparison of transfer function reconstruction (solid), DCA (dashed) and DHI (dotted) scores after data normalisation. All curves are plotted against depth.

Figure 7.10 Comparison of transfer function water table reconstructions using the a) ACCROTELM model (Charman *et al.*, 2007; solid), and b) North of Ireland model (Swindles *et al.*, 2009; dashed) for testate amoebae data at Sluggan Moss. Errors are not shown for ease of reference. ACCROTELM reconstruction average error = 6.146 cm; North of Ireland reconstruction average error = 5.639.

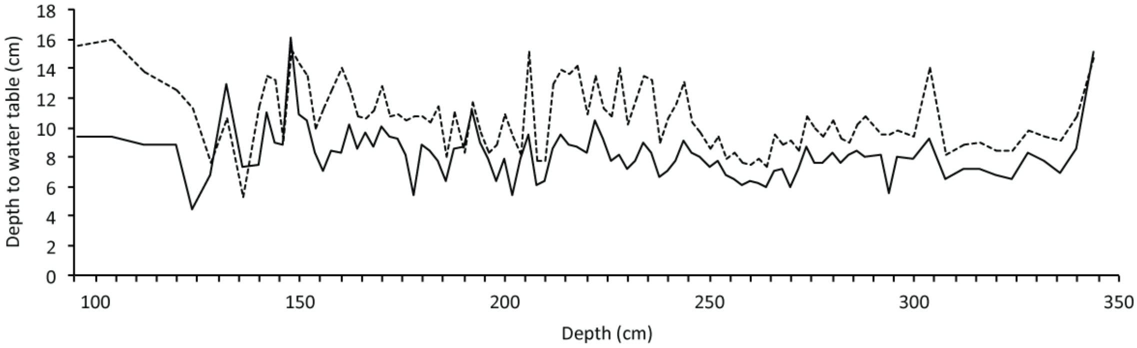
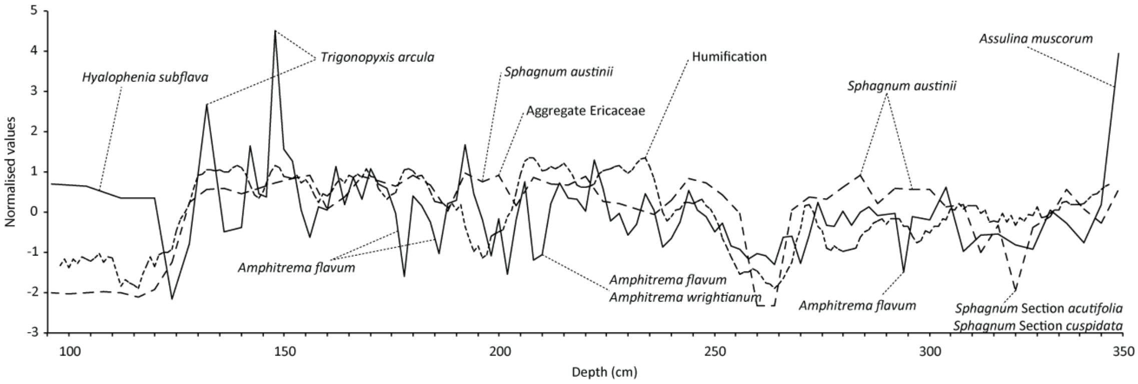


Figure 7.11 Comparison of normalised data for testate amoebae-based reconstructed water table, humification and plant macrofossil DCA normalised palaeoecological data at Sluggan Moss, plotted against depth. Annotations are provided to explain disparity between the proxies.



7.2.5 Stable isotopes

Figure 7.12 presents the results of stable isotopic analysis of *Sphagnum* α-cellulose samples from Sluggan Moss. Error bars present 2 standard error ranges based on repeated measurements of the same sample. Triplicate measurements were taken for the majority of samples but, as detailed in Table 7.3, there were a small number of instances where only duplicate or single measurements were possible, as a result of small sample size and/or analytical error. Stratigraphical resolution of the analyses reflects: i) the distribution of suitable *Sphagnum* material in the plant macrofossil record; and ii) an effort to focus analyses around the marker horizon of the Hekla 4 tephra.

Figure 7.12a presents $\delta^{18}O$ values analysed from *Sphagnum* α-cellulose samples at this site. Analyses within the studied sequence exhibit a slight upward trend in $\delta^{18}O$ values (i.e. becoming less depleted) with decreasing age. Values ranged from the highest excursion (25.18‰) at 180 cm (c. 3600 cal yr BP), to the lowest (20.9‰) at 232 cm (c. 4550 cal yr BP), producing a total range of isotopic variation in the record of 4.28‰. Apart from four larger (i.e. >1‰) excursions towards isotopically depleted values, $\delta^{18}O$ values generally vary within ± 1‰ of 24‰. These larger excursions are centred on 312, 260, 232, 188 cm, and date to approximately 6050, 5150, 4500 and 3750 cal yr BP, respectively, rounded to the nearest 50 years.

Table 7.3 Sample isotopic measurements per sample depth

Sample depth (cm)	Sample measurements	
	$\delta^{18}O$	$\delta^{13}C$
144 cm	3	2
148 cm	1	1
152 cm	2	2
156 cm	2	2
164 cm	1	1
196 cm	2	2
244 cm	2	2
All other samples (*n* = 40)	3	3

Figure 7.12b presents $\delta^{13}C$ values analysed from *Sphagnum* α-cellulose samples. Note that the y-axis has been inverted for ease of comparison in Figure 7.12c. Analyses within the studied sequence exhibit an upward trend in $\delta^{13}C$ values (i.e. becoming more depleted with decreasing age. Values ranged from the highest (-21.46‰) at 260 cm (c. 5150 cal yr BP), to the lowest (-27.25‰) at 220 cm (c. 4300 cal yr BP), producing a total range of isotopic variation in the record of 5.79‰. Apart from three larger (i.e. >1‰) excursions towards isotopically enriched values, $\delta^{13}C$ values generally vary within ± 1‰ of -26‰. These larger excursions are centred on 312, 260 and 232 cm, and date to approximately 6050, 5150, 4550 cal yr BP, respectively, rounded to the nearest 50 years.

When $\delta^{18}O$ and $\delta^{13}C$ values are plotted together against depth (Fig. 7.12c) some agreement appears to exist between the two curves. $\delta^{13}C$ values exhibit marginally more isotopic variation (5.79‰) than $\delta^{18}O$ (4.28‰). Of the three isotopic excursions seen in both records (i.e. 312, 260 and 232 cm), the magnitude of change is almost identical in the lower two. The excursion at 232 cm, whilst a distinct excursion in its own right (i.e. characterised by excursion, followed recovery to previous levels) exists in the $\delta^{18}O$ record, in the $\delta^{13}C$ record appears the excursion appears to be consumed within a broader enrichment trend, following the previous large excursion at 260 cm. However, this relationship appears to vary either side of 180 cm. Statistical analysis, by calculating Pearson product moment correlation coefficient (PPMCC), reveals a moderate negative correlation between all $\delta^{18}O$ and $\delta^{13}C$ data at Sluggan Moss, which was statistically significant (PPMCC = -.576, $p < .05$; Fig. 7.13a, Tab. 7.4). This negative correlation is strengthened in the data between 312 and 180 cm depth (PPMCC = -.680, $p < .05$; Fig. 7.13b, Tab. 7.13), suggesting that as $\delta^{18}O$ becomes enriched, $\delta^{13}C$ becomes depleted, and vice versa. However, it is worth noting that this relationship is likely to be driven largely by the group of 6 samples (>-24 ‰ $\delta^{13}C$, <23 ‰ $\delta^{18}O$) which occur at 232, 254 – 264 (4 samples), and 312 cm and therefore coincide with the three isotopic excursions described earlier (i.e. >1‰), suggestive of a common driver during big excursions. The data 180 – 96 cm, whilst exhibiting a positive correlation, and visually suggesting coincident enrichment and depletion between isotopes, this relationship is not statistically significant (PPMCC = .320, $p < .05$; Fig. 7.13c, Tab. 7.13).

Figure 7.12 Stable isotope data from Sluggan Moss, as derived from *Sphagnum* α-cellulose, including: a) $\delta^{18}O$ and b) $\delta^{13}C$, where y-axis error bars denote 2σ ranges on replicate analyses from a single level where possible; c) comparison of $\delta^{18}O$ (solid) and $\delta^{13}C$ (dashed) data. All data are plotted against depth. $\delta^{13}C$ values are inverted for reference.

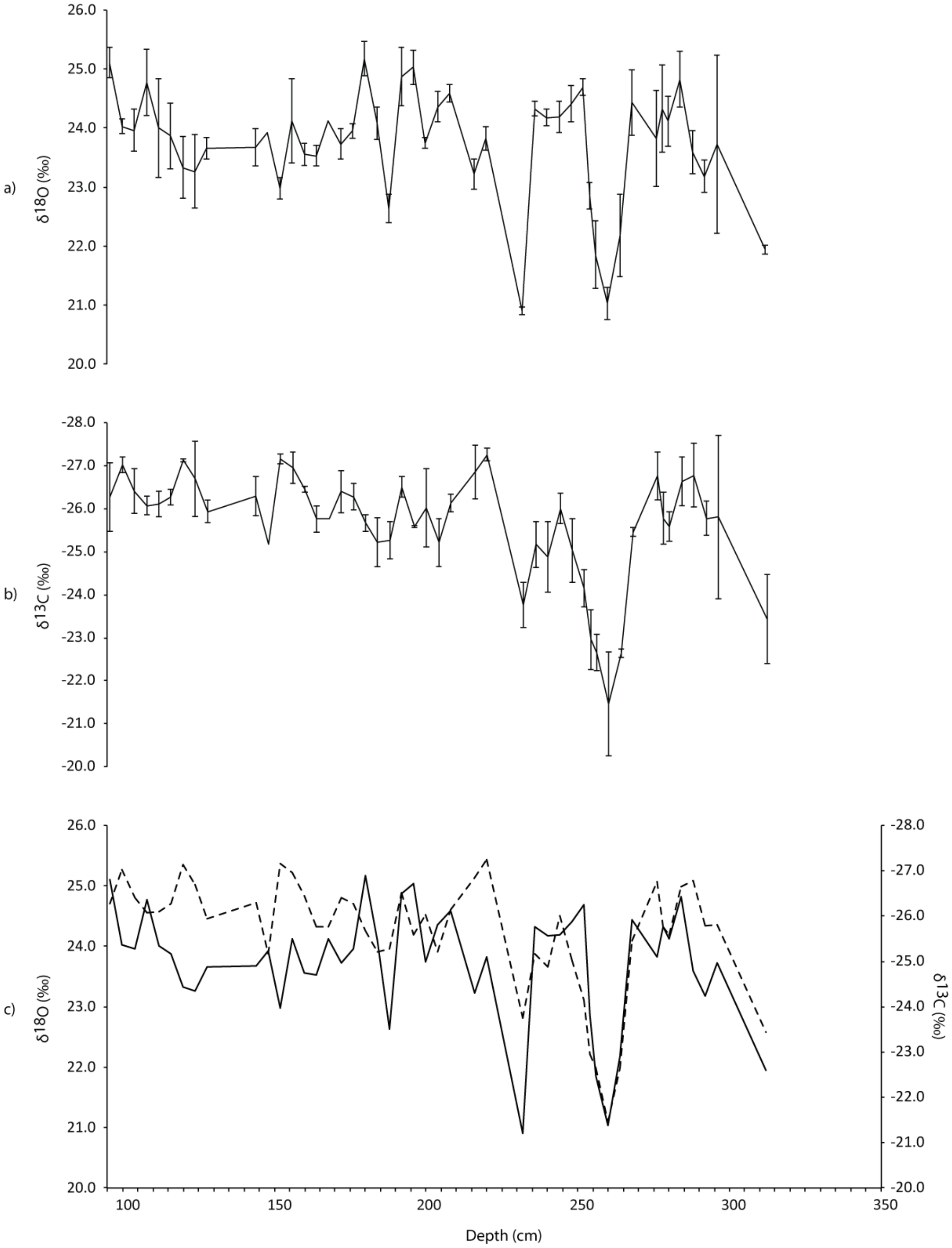


Figure 7.13 Biplot for Sluggan Moss showing δ^{18}O and δ^{13}C data; a) all data; b) 312 – 180 cm; c) 180 – 96 cm. Linear regression lines have been fitted for reference.

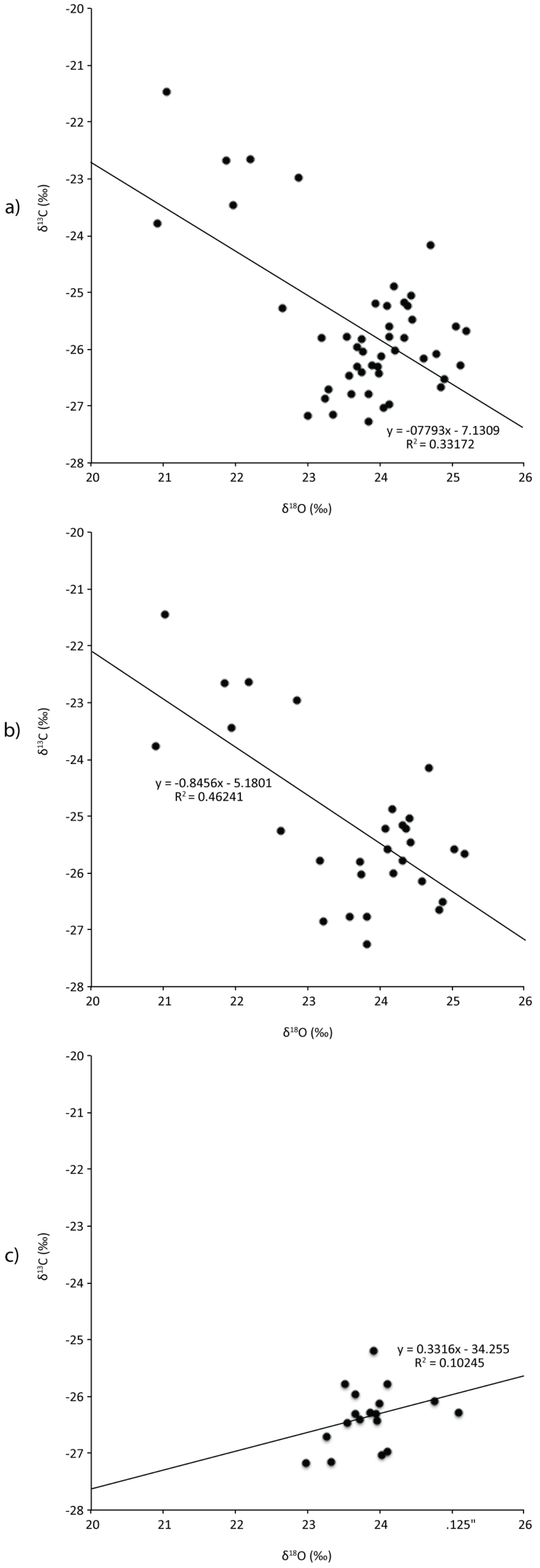


Table 7.4 Pearson product moment correlation coefficient of $\delta^{18}O$ and $\delta^{13}C$ from Sluggan Moss. Asterisks denote statistical significance.

	PPMCC	**Sig. (2-tailed, p<0.05)**	**n**	**R^2**
a) All data	-.576	.000*	47	0.33172
b) 312 – 180 cm	-.680	.000*	29	0.46241
c) 180 – 96 cm	.320	.195	18	0.10245

7.3 Fallahogy Bog

7.3.1 Plant macrofossils

Plant macrofossil data from Fallahogy Bog are presented in Figure 7.14, against depth and age respectively. Again, plant macrofossil zones were identified, using a combination of the CONISS numerical method within TILIA, and judgement by eye (see Section 7.2.1). In addition, Table 7.5 offers a summarised botanical description of each plant macrofossil zone, together with its approximate depth and age.

As at Sluggan Moss, the floral composition, and the variation within it, at Fallahogy Bog is typical of raised, ombrotrophic bogs in the region (e.g. Barber *et al.*, 1994; Hughes *et al.*, 2000), and subsequently characterised by fluctuations between *Sphagnum*, monocotyledon and ericaceous species. Two *Sphagnum* species are present in this section of the sequence, with the dominant *S. austinii* functioning as the main peat-forming species. *S. austinii* is present from the base of the studied sequence, FAL-a, and therefore a precise date of its emergence as a dominant species cannot be ascertained. However, a minimum of age of c. 5650 cal yr BP places the emergence of this taxon significantly earlier than at sites in Great Britain (see Section 8.2.1.1). Within this basal zone, species indicative of fen conditions, or associated with the fen-bog transition (Hughes *et al.*, 2000; Hughes and Barber, 2004) are absent. Instead, this early dominance of *S. austinii* indicates that the fen-bog transition occurred prior to c. 5650 cal yr BP.

The lower portions of the studied sequences zones FAL-a to FAL–d are characterised by alternate dominance of *Sphagnum*, monocotyledon and ericaceous species. Generally, ericaceous and monocotyledon remains occur during periods of reduced *Sphagnum* dominance and, on two distinct occasions (FAL-b and FAL-d) peaks in ericaceous remains are followed by monocotyledon peaks potentially suggestive of autogenic ecological succession. In this lower section, total ericaceous and monocotyledon proportions are largely dominated by unidentified root material. However, small amounts of *Calluna vulgaris*, *Erica tetralix*, *Vaccinium oxycoccus*, *Eriophorum vaginatum* and *Eriophorum angustifolium* were identified. Generally, concentration of UOM also varies coincidently with ericaceous remains although there appears to be no long-term trend in the concentrations of ericaceous material through the sequence. Monocotyledon remains, however, exhibit a broad, long-term increase towards the top of the studied sequence.

The total *Sphagnum* component is overwhelmingly dominated by *S. austinii*, with *S.* s. *Acutifolia* achieving concentrations over 20% only three times within the studied sequence. The plant macrofossil zone FAL-e, represents the period of most prolonged *S. austinii* dominance which is coupled with prolonged sequence minima for UOM, monocotyledon and ericaceous species, therefore implying a shift towards longer term wetter and/or cooler conditions. There is, however, one notable peak of dry indicator *T. cespitosum* within this zone, and small amounts of *C. vulgaris*, *E. tetralix*, *V. oxycoccus*, *E. vaginatum* and *E. angustifolium* were also identified. Again, variations in the dominance of the total *Sphagnum* component in the sequence often occur in an abrupt fashion, suggesting that the genus operates via threshold response to variability in the water table. Although, as previously mentioned, the rapid accumulation rates associated with this species could account for some of the rapidity seen during these transitions.

Finally, the upper portion of the studied sequence (FAL-f and FAL-g) is characterised by a shift towards an increased dominance of monocotyledon and ericaceous species with significantly less *Sphagnum* present. Monocotyledon species identified include *T. cespitosum*, *E. vaginatum* and *E. angustifolium*, alongside *C. vulgaris*. In addition, a substantial upward trend in UOM concentration begins, and continues to the top of the upper zone. This trend, combined with associated floral composition, implies a shift toward higher levels of humification and drier bog surface conditions.

The plant macrofossil data from Fallahogy Bog were subject to DCA to produce a quantitative proxy-climate index, enabling comparison with other quantitative palaeoecological reconstructions. Procedure matched that at Sluggan Moss, with no rare species down-weighting or transformation (ter Braak, 1995). Figure 7.15 shows the distribution of plant species scores against a biplot of DCA axes one and two from Fallahogy Moss. Initial inspection of the distribution of species scores along axis one, suggests the presence of a water table gradient within the data. The eigenvalue of axis one (0.4738) is also moderately strong, supporting any subsequent interpretations of changing BSW based upon it.

Both *Sphagnum* species are located on the right hand side of the axes, with drier species to the left, at varying degrees. Dry indicators such as *T. cespitosum* and *E. tetralix*, as well as unidentified ericaceous and organic material were all located at 1.5 on axis one. Surprisingly, the ericaceous *V. oxycoccus*, usually indicative of dry conditions, was located on the far right of the first axis, above both *Sphagnum* species. This can be explained by the fact that it occurs just once, and in very small quantities, in the studied sequence and subsequently its position on axis one is relatively unreliable but influential. However, *V. oxycoccus* is often characteristic of *Sphagnum* dominated areas of the bog surface as it is unable to compete in areas of denser ericaceous or monocotyledon growth, which, when combined with its sporadic appearance, may go some way to explaining its position on the DCA axis.

Figure 7.16 shows the distribution of sample scores against a bi-plot of DCA axes one and two from Fallahogy Bog. Samples from the same plant macrofossil zones group together well, with FAL-a, FAL-e and FAL–g most tightly constrained. Samples from FAL-b, FAL-c and FGAL-d are less tightly constrained but maintain a degree of coherent structure in their distribution.

Figure 7.17a displays axis one sample scores against depth for Fallahogy Bog. By comparing this curve to the plant macrofossil data seen in Figure 7.14, it becomes apparent that the DCA is heavily influenced by interchange between *Sphagnum* and non-*Sphagnum* species, with drier conditions being expressed during periods of lower *Sphagnum* concentrations. A notable dry peak occurs 304 cm and is clearly driven by a corresponding peak in the concentration of dry indicator, *T. cespitosum*.

The plant macrofossil data from Fallahogy Bog were also transformed using the DHI method (Dupont, 1986) into a quantitative proxy-climate index of past changes in BSW. Figure 7.17b plots the results of DHI analysis, with Figure 7.17c plotting both the DCA and DHI curves against depth and normalised to aid comparison. Here DHI and DCA exhibit strong coherence, with extremely similar variability both in terms of timing and magnitude. The only difference of note is the magnitude of the dry shift at 304 cm, which, as previously discussed was caused by a peak in *T. cespitosum*. The good level of agreement between the DHI and DCA reconstructions provides further support for any interpretations of changes in BSW based on the DCA results and species composition outlined in Figure 7.14 and Table 7.5.

7.3.2 Peat humification

Raw percentage light transmission data, obtained from contiguous 1cm samples through the Fallahogy sequence, are presented in Figure 7.18a. A linear regression line, fitted to the raw data, demonstrates that no long-term downcore trend exists in this instance. As a result, the humification results required no detrending, but were normalised to enable comparison with other proxy data from this sequence (Blundell and Barber, 2005) and plotted against depth in Figure 7.18b. The data exhibit a high degree of variability with a number of shifts characterised by prolonged (>10cm) trends towards higher transmission values and lower degrees of humification, indicative of wetter and/or cooler climatic conditions, that begin at c. 450, 387, 355, 329, 305, 245 cm. Two of these shifts, at 355 and 245 cm correspond broadly with the onset of plant macrofossil zones FAL-e (begins 350 cm) and FAL-f (begins 258 cm), which represent a substantial shift in botanical composition to, and from, a dominance of *S. austinii*, respectively.

7.3.3 Testate amoebae

Testate amoebae data from Fallahogy Bog are presented in Figure 7.19, against depth and age respectively. Again, these data are discussed in relation to the plant macrofossil zones (see Section 7.3.1, cf. Hughes *et al.*, 2006). Description of the varying faunal composition, discussed by zone, is presented in Table 7.6.

The studied sequence at Fallahogy Bog is generally dominated by four species, *Amphitrema flavum*, *Assulina muscorum*, *Difflugia pristis* type and *Difflugia pulex*, with two other species, *Cyclopyxis arcelloides* and *Trigonopyxis arcula*, also present in significant quantities at varying points in the sequence. The highest peaks in recorded taxa presence occur in *A. flavum* and *D. pulex*, with the latter being the most dominant of all taxa present.

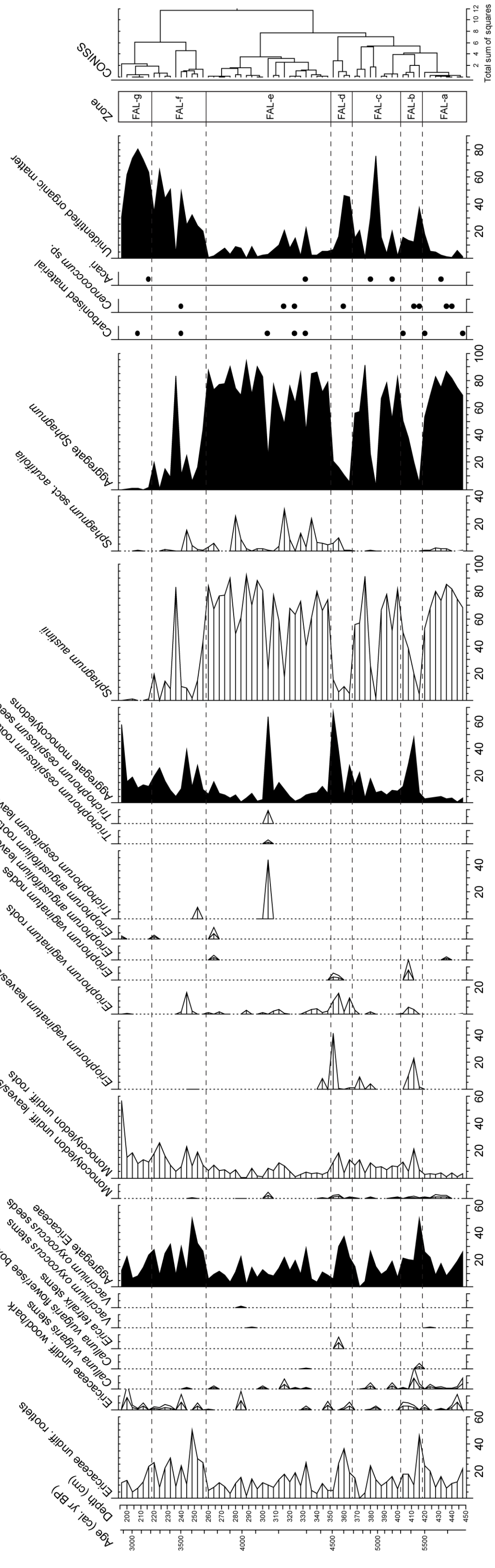


Figure 7.14 Plant macrofossil diagram for Fallahogy Bog plotted against depth. Peat components displayed with depth bars are derived from averaged quadrat counts (n=15) under low-power magnification (x10). Peat components displayed in solid black denote aggregates of different components. Individual *Sphagnum* leaf counts are also displayed with depth bars, and consist of proportions of aggregate *Sphagnum* based on random selection of 100 leaves identified at high magnification (x400). All data are presented as percentages, with the exception of presence/absence data displayed as black dots. Zones were assigned after CONISS cluster analysis.

Table 7.5 Botanical description of plant macrofossil zones at Fallahogy Bog.

Zone	Depth (cm)	Floral assemblage description	Age (cal yr BP)
FAL-a	448–418	A zone dominated by *Sphagnum austinii* with only very small amounts of *S.* s. *Acutifolia* present. There is a low, but persistent, presence of monocotyledon material. This material consists mainly of unidentified roots, but *Eriophorum angustifolium* is identified in small quantities. Ericaceous material is present in moderate quantities, consisting mostly of rootlets, but also some material identifiable as *Calluna vulgaris*. Very small amounts of *Vaccinium oxycoccus* were also found. UOM is present but only in small quantities.	5650 – 5500
FAL-b	418–402	A zone characterised by a significant reduction in *Sphagnum austinii*, coincident with a moderate peak of ericaceous material, of which some was identified as *Calluna vulgaris*. Immediately after this peak, a peak of monocotyledon remains occurred, with significant presence of all *Eriophorum vaginatum* components, including leaves, stems, roots and nodes. UOM peaks and falls coherently with the ericaceous remains.	5500 – 5300
FAL-c	402–366	A variable zone characterised, in large part, by alternating peaks of *Sphagnum austinii* and UOM. Generally, *Sphagnum austinii* proportions remain high, with a significant peak of UOM occurring at 384cm. Monocotyledon remains are relatively stable and persistent, but low, with small amounts of *Eriophorum vaginatum* identified. Ericaceous material, of which small amounts can be identified as *Calluna vulgaris*, is present throughout the zone, but drops to zero at 372cm	5300 – 4700
FAL-d	366–350	A zone characterised by coincident, significant peaks in both ericaceous material and UOM, centring at 360cm, with remains of *Erica tetralix* present. *Sphagnum* is dramatically reduced, with *Sphagnum austinii* present at significantly lower levels than previous zones. *S.* s. *Acutifolia*, however, emerges in noticeable quantities for the first time in the sequence, half way through this zone. Towards the end of this zone monocotyledon remains peak at their highest in the entire sequence. Significant amounts of all *Eriophorum vaginatum* components (inc. leaves, stems, roots and nodes) were identified.	4700 – 4500
FAL-e	350–258	A zone characterised by the prolonged domination of *Sphagnum austinii*. This is punctuated by a series of smaller peaks in *S.* s. *Acutifolia* at 336, 316 and 280 cm. Ericaceous material maintains a constant but relatively low presence, with *Calluna vulgaris* and small amounts of *Vaccinium oxycoccus* remains identified. Similarly, monocotyledon remains are consistently low in proportion with some *Eriophorum vaginatum* roots identified. A significant peak in *Trichophorum cespitosum* is witnessed at 304 cm, with leaf, stem root and seed components identified. *Eriophorum angustifolium* was also identified in small quantities at the end of the zone. UOM maintains a low but constant presence throughout the zone.	4500 – 3700
FAL-f	258–218	A transitional zone characterised by increases in ericaceous material and UOM, coincident with a significant reduction in the presence of *Sphagnum*. The zone begins with a peak in ericaceous material, where small amounts of *Calluna vulgaris* have been identified. Monocotyledon remains increase slightly on the previous zone, with *Eriophorum vaginatum, Trichophorum cespitosum, Eriophorum angustifolium* all present at varying points. Although *Sphagnum* is vastly reduced on the previous zone, a significant peak of *Sphagnum austinii* takes place at 236 cm, with relatively small amounts of *S.* s. A*cutifolia* present until the middle of the zone. UOM increases steadily from zero, until it dominates the composition, with the only significant reduction occurring coincidently with the *Sphagnum austinii* peak at 236 cm.	3700 – 3200
FAL-g	218–196	A zone characterised by dominance of the composition by UOM, predominantly at the expense of the *Sphagnum* species, with *Sphagnum austinii* and *S.* s. *Acutifolia* both present only in very small quantities. Ericaceous material is reduced compared to the previous zone and was not identified to species level. Monocotyledon material maintains a steady presence throughout the zone, peaking significantly at the end of the zone. *Eriophorum vaginatum* and *Eriophorum angustifolium* were both identified in low quantities.	3200 – 2850

Figure 7.15 DCA scatter plot for plant macrofossil data at Fallahogy Bog, showing species against axes one and two.

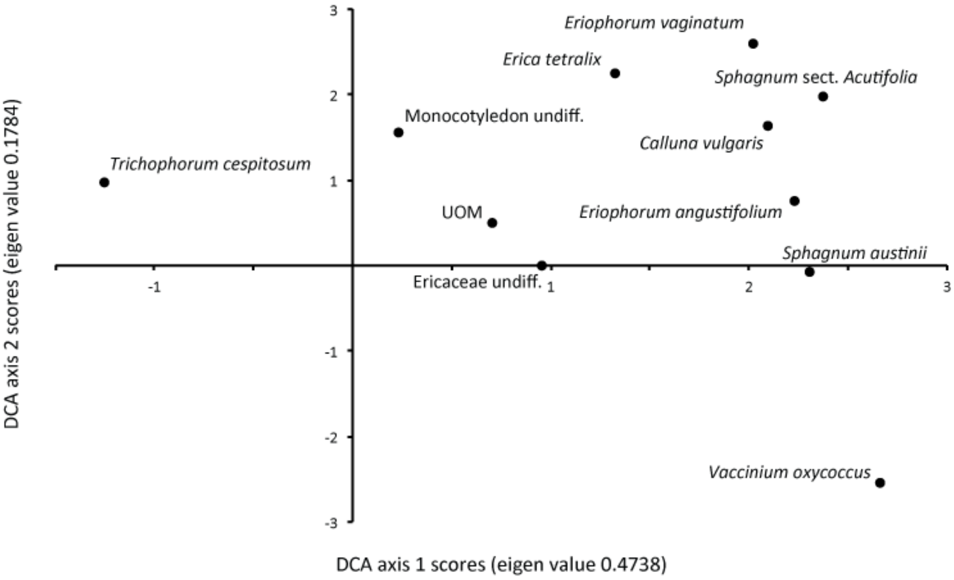


Figure 7.16 DCA scatter plot for plant macrofossil data at Fallahogy Bog, showing samples against axes one and two. Samples are also colour-coded according to CONISS zone (see Fig. 7.14).

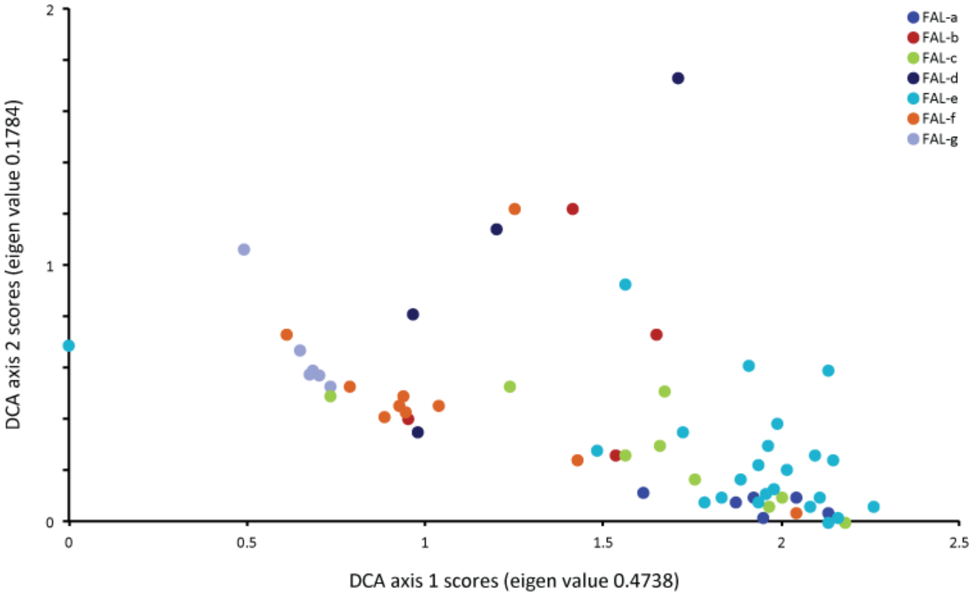


Figure 7.17 Plant macrofossil data from Fallahogy Bog displayed as: a) DCA axis one sample scores; b) DHI sample cores; c) comparison of DCA (solid) and DHI (dashed) scores after data normalisation. All curves are plotted against depth.

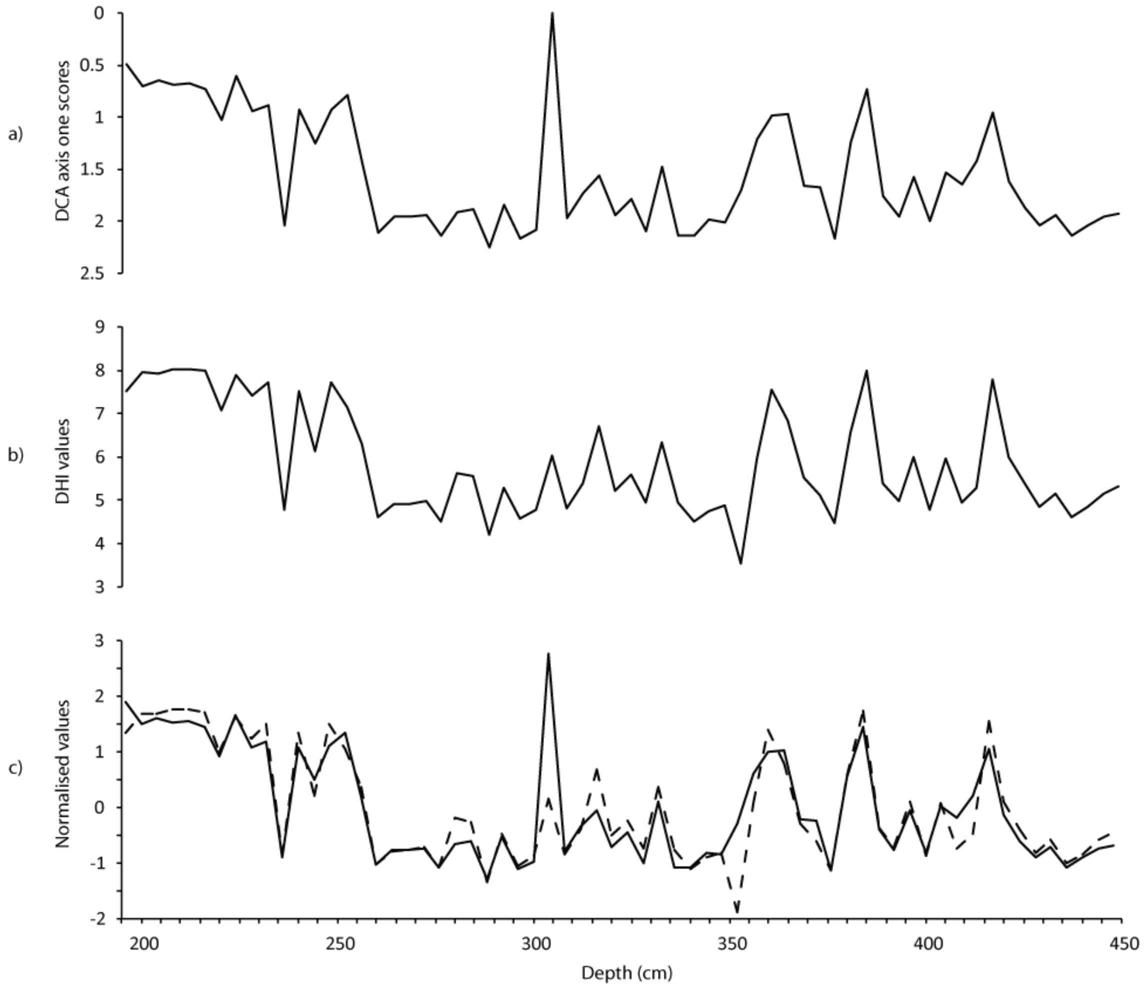
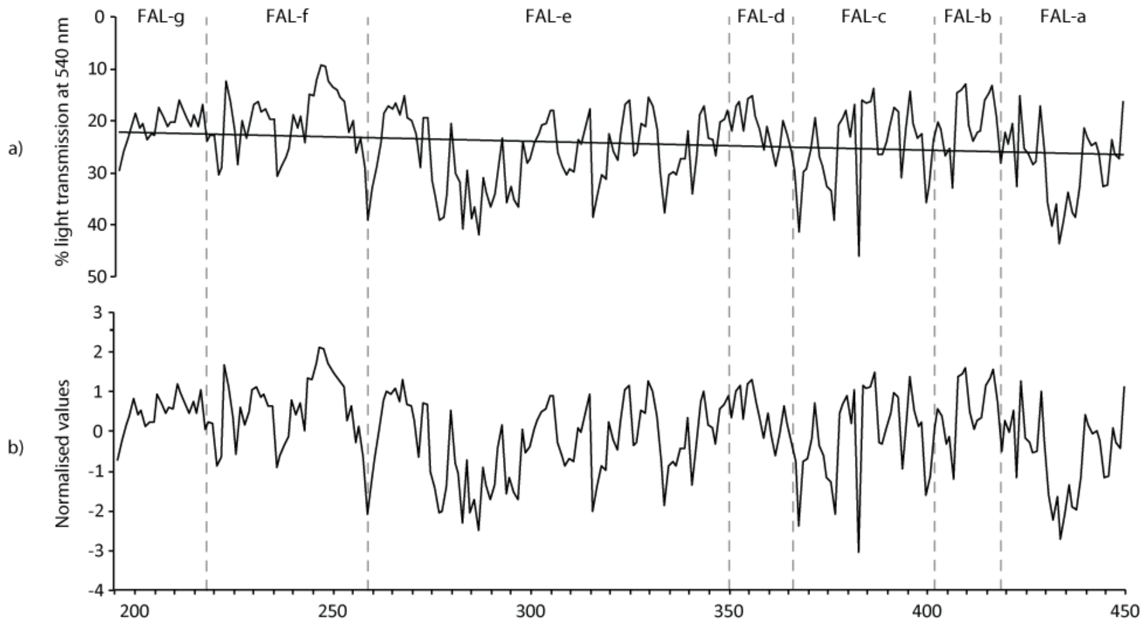


Figure 7.18 Humification data for Fallahogy Bog, displayed as: a) 'Raw' values for 540 nm light transmission through humic acid extraction in solution; b) Normalised values. Both curves are plotted against depth.



Typically, *A. flavum* is generally associated with wetter conditions, and *D. pulex* with relatively dry conditions. *A. muscorum, C. arcelloides* and *D. pristis* type all possess wide hydrological tolerances. *T. arcula* is, however, associated with lower water tables and drier bog surfaces (Charman *et al.*, 2000, 2007). Down-core variation in the faunal composition of the testate amoebae record is broadly coherent with these interpretations. For example, peaks in dry indicator *T. arcula* correspond well with points in the record where wet indicator *A. flavum* is largely absent; *D. pulex* and *A. flavum* operate in broad anti-phase; intermediate species, such as *A. muscorum*, *C. arcelloides* and *D. pristis* type, whilst maintaining a relatively low, but continuous presence, tend to peak during periods of low *A. flavum* and *D. pulex*.

On initial inspection variations in the faunal composition of the testate amoebae data at Fallahogy Bog demonstrate relatively little agreement with the plant macrofossil zones, suggesting that the testate amoebae and bog plants are not responding to a common driving mechanism. There is however a general drying trend towards the top of the sequence characterised by the continued dominance of *D. pulex*, together with a notable increase in *T. arcula* and decline in *A. flavum*, from 292 cm to the top of the sequence. A drying trend is also seen in the plant macrofossil record, but begins slightly later, from 258 cm upwards, in zones FAL-f and FAL-g.

The raw testate amoebae assemblage data, however, do not provide evidence of a potential host-commensal symobiont relationship between *S. austinii* and *A. flavum*, which has been suggested (see Section 7.2.3; Section 8.2.3.2), with the latter persisting strongly throughout periods of reduced S. austinii dominance, such as FAL-b, FAL-d, FAL-f and FAL-g. The testate amoebae data from Fallahogy Bog were subject to DCA to produce a quantitative proxy-climate index, enabling comparison with other quantitative palaeoecological reconstructions. Procedure matched that at Sluggan Moss, as detailed in Section 7.2.3, with rare species down-weighting and no transformation (ter Braak, 1995).

Figure 7.20 shows the distribution of species scores against a biplot of DCA axes one and two from Fallahogy Bog. Again, despite a poor axis one eigenvalue (0.2369), a case could be argued for the presence of a water table gradient. *T. arcula* is located on the left-hand side of the axis, with a score below -1. *Arcella catinus*, also a prominent dry indicator, is also located towards the left of the axis. Conversely, *A. flavum* is located on the far right of the axis, with a value approaching 2.5. *Nebela carinata*, another wet indicator, is also located above 2 on axis one. Intermediate, broad tolerance species previously discussed, such as *D. pristis*, *A. muscorum* and *A. seminulum*, occupy fairly central positions on the axis. There are, however, some inconsistencies with intermediate species *C. arcelloides*, located towards the far-left of the axis, with a score of c. 0.5 and *D. pulex*, considered in some studies a drier indicator, occurring centrally with an axis one score of approximately 1, for example.

Figure 7.21 shows the distribution of sample scores against a biplot of DCA axes one and two from Sluggan Moss. In general, samples falling within the same plant macrofossil zone are poorly constrained, in terms of their axis one and two distributions, supporting the earlier assertion that variation in the faunal composition of the testate amoebae record does not conform well to these subdivisions.

Figure 7.22a displays axis one sample scores against depth for Fallahogy Bog testate amoebae data. Despite the inconsistencies discussed above, the broad coherence between the DCA axis one score curve and the WTD reconstruction using the ACCROTELM pan-European transfer function (Fig. 7.22b; Charman *et al.*, 2007) suggests that the former is operating as an indicator of past changes in BSW, supporting the interpretation of the axis as a hydrological gradient. This is best observed when both indexes are normalised and presented together in Figure 7.22d.

As at Sluggan Moss, raw testate amoebae data were also transformed using a version of the DHI method (Dupont, 1986; see Section 7.2.3) modified here for use with testate amoebae into a quantitative proxy-climate index of past changes in BSW. Figure 7.22c displays the results of DHI analysis with species scores plotted against depth. Again, comparison of this curve with DCA and transfer function reconstructions, after normalisation, in Figure 7.22d demonstrates moderate coherence and subsequent success of the modified DHI technique. Again, the DHI technique exhibits marginally more variability than the curve produced by DCA.

Figure 7.22d plots the normalised results of DHI and DCA analysis alongside the normalised transfer function water table reconstruction and against depth to aid comparison, and demonstrates a good level of agreement between all three statistical techniques. For the majority of the studied sequence the transfer function-based WTD reconstruction exhibits a degree of variability centered around a WTD of c. 10 cm, but from which no discernible wet or dry shifts can be identified (Figure 7.22b), although it could be argued that there is a slight, longer-term drying trend. Towards the top of the sequence a distinct shift to drier conditions can be seen, commencing at 220 cm and peaking at 208 cm, before returning to a WTD that is comparable to the earlier section of the sequence. When the data are normalised and presented alongside DHI and DCA curves, other potential wet shifts emerge, where all three techniques are in good agreement. Whilst many of these shifts are minor and short-lived, one appears to be a more prolonged shift towards wetter conditions and commences at 280 cm, with a downward trend to wetter conditions persisting until 244 cm.

Figure 7.23 presents a comparison of the pan-European ACCROTELM (Charman *et al.*, 2006) and regionally specific North of Ireland (Swindles *et al.*, 2009) transfer function water table reconstructions (cf. Langdon *et al.*, 2012; see Section 5.4.3.2). Again, the two reconstructions demonstrate a good degree of agreement, with the key shifts observed above present in both curves. As at Sluggan Moss, the North of Ireland model does, however, appear to suggest a higher degree of variability than the ACCROTELM model, with lower average WTD throughout. When comparing each transfer function's reconstruction of an Irish dataset, Langdon *et al.* (2012) observed greater agreement at the wetter, rather than the drier, end of the water table gradient. It would be difficult to judge whether the same is true here, however.

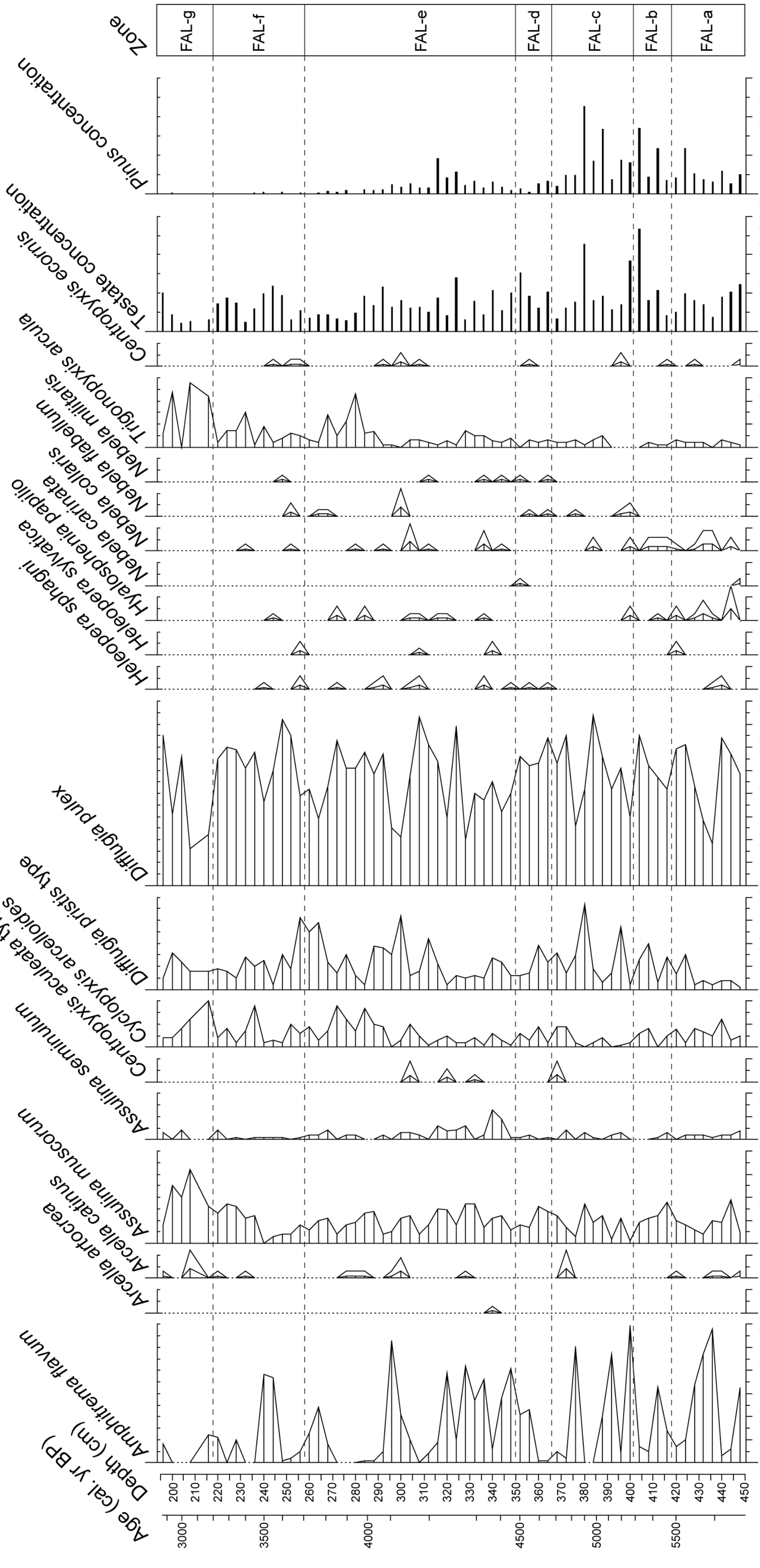


Figure 7.19 Testate amoebae diagram for Fallahogy Bog plotted against depth. All data are percentages of the total number counted per level. Concentration data are displayed in bar form. Zones follow CONISS cluster analysis of plant macrofossil data (see Fig. 7.14).

Table 7.6 Faunal description of testate amoebae record at Fallahogy Bog, by plant macrofossil zone.

Zone	Depth (cm)	Faunal assemblage description	Age (cal yr BP)
FAL-a	448–418	This basal zone, like much of this sequence, is dominated by *Difflugia pulex*. However, a notable peak in *Amphitrema flavum* centring at 436 cm at the expense of *Difflugia pulex*. *Assulina muscorum*, *Assulina seminulum*, *Cyclopyxis arcelloides* and *Difflugia pristis* all persist at levels which are characteristic of their presence in this sequence, with no notable peaks or troughs. *Trigonopyxis arcula* is also present in low levels.	5650 – 5500
FAL-b	418–402	A zone dominated by *Difflugia pulex* with *Amphitrema flavum* peaking only moderately at 312 cm. *Assulina muscorum* and *Difflugia pristis* continue at moderate levels. *Assulina seminulum* and *Cyclopyxis arcelloides* are also both present, but drop in proportion to zero at points within this zone. *Trigonopyxis arcula* is also present in low levels.	5500 – 5300
FAL-c	402–366	A zone dominated by *Difflugia pulex* with *Amphitrema flavum* peaking notably three times at 400, 392 and 376 cm. Each peak is characterised by a rapid rise and fall in *Amphitrema flavum* concentrations, together with an associated trough in *Difflugia pulex*. There are also two peaks in *Difflugia pristis*, which operate in anti-phase with those of *Amphitrema flavum*. *Assulina muscorum* persists at its moderate level, and *Assulina seminulum* and *Cyclopyxis arcelloides* are both present at reduced levels. *Trigonopyxis arcula*, after absence at the beginning of the zone, begins a low, but stable presence in the sequence, which persist for a number of zones.	5300 – 4700
FAL-d	366–350	A zone dominated by *Difflugia pulex* but with *Amphitrema flavum* rising from low levels towards the end of the zone. *Assulina muscorum*, *Cyclopyxis arcelloides*, *Difflugia pristis* all persist at moderate levels. *Assulina seminulum* is reduced to almost zero, and *Trigonopyxis arcula* continues to be present at low, but stable levels.	4700 – 4500
FAL-e	350–258	A zone characterised by an overall dominance of *Difflugia pulex*. However, the first half of the zone sees an alternation in dominance between *Difflugia pulex* and *Amphitrema flavum*. Above 320 cm, *Amphitrema flavum* is reduced dramatically, at times to zero, save for a notable peak at 296 cm. This upper section also witnesses an increase in *Difflugia pristis* and *Trigonopyxis arcula*. Here, the latter taxon experiences the first of two periods of moderately high concentration in this sequence, which coincides with the reduction of *Amphitrema flavum* to near zero. *Assulina muscorum* and *Cyclopyxis arcelloides* are present in notably constant, moderate levels. Generally, levels of *Assulina seminulum* exhibit a declining trend towards the end of this zone.	4500 – 3700
FAL-f	258–218	A zone heavily dominated by *Difflugia pulex* with only one notable peak in *Amphitrema flavum* centring around 240 cm. *Assulina muscorum* and *Cyclopyxis arcelloides* are, again, relatively constant at moderate levels. *Difflugia pristis* declines in this zone, from a relatively high level to resume a constant, but moderate presence. *Trigonopyxis arcula* also maintains a constant, moderate presence. *Assulina seminulum* experiences its lowest persistent values in the studied sequence.	3700 – 3200
FAL-g	218–196	This upper zone is dominated by *Difflugia pulex* with *Amphitrema flavum* present in only very low levels, relative to the rest of the studied sequence. *Assulina muscorum* and *Trigonopyxis arcula* are both present in notable concentrations, representing their highest, persistent presence anywhere in the studied sequence. *Cyclopyxis arcelloides* falls in concentration from the beginning of the zone and *Difflugia pristis* maintains a moderate but constant presence. *Assulina seminulum* is present only in small amounts. This zone is also notable for its lack of species diversity. Unlike the other zones in this sequence, which feature a large number of species at low concentration levels, this zone features only the species mentioned above. Of note is also a level, 212 cm, in which testate amoebae were in such a poor state of preservation that it had to be omitted from the analyses.	3200 – 2850

Figure 7.20 DCA scatter plot for testate amoebae data at Fallahogy Bog, showing species against axes one and two.

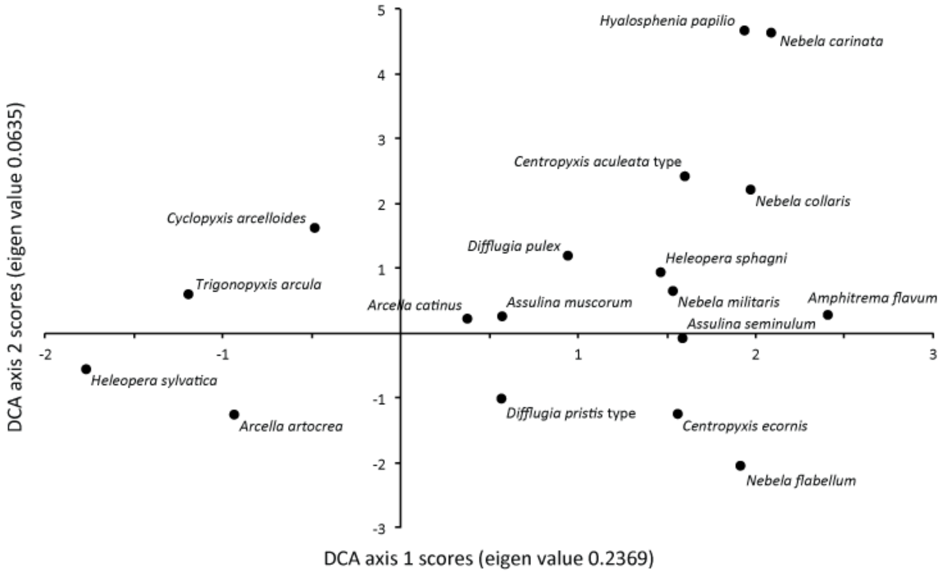


Figure 7.21 DCA scatter plot for testate amoebae data at Fallahogy Bog, showing samples against axes one and two. Samples are also colour-coded according to plant macrofossil CONISS zones.

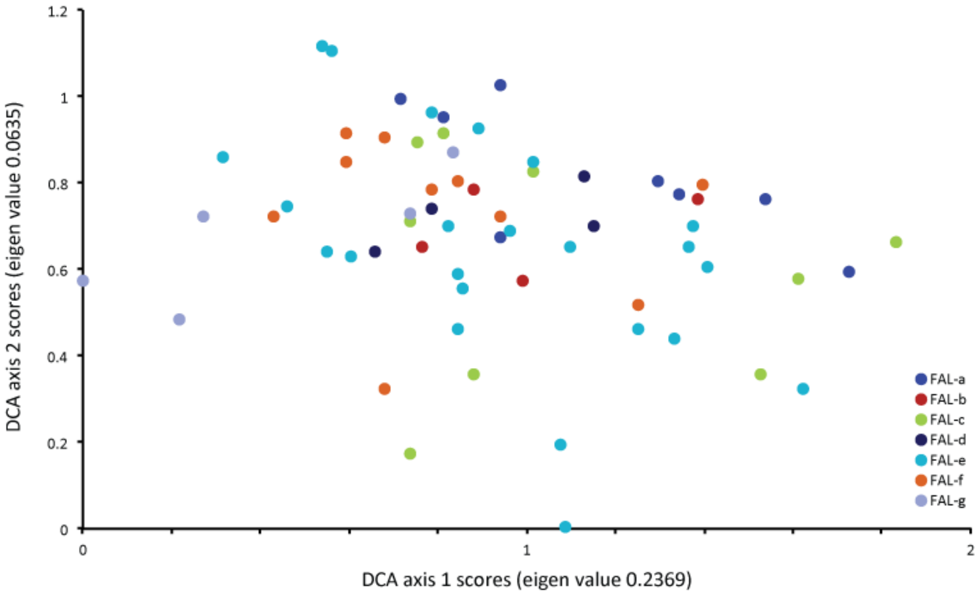


Figure 7.22 Testate amoebae data from Fallahogy Bog displayed as: a) inferred WTD reconstructions based on the ACCROTELM pan-European transfer function (Charman *et al*., 2007) with calculated error displayed in light grey; b) DCA axis one sample scores; c) testate amoebae-adapted DHI sample cores; d) comparison of transfer function reconstruction (solid), DCA (dashed) and DHI (dotted) scores after data normalisation. All curves are plotted against depth.

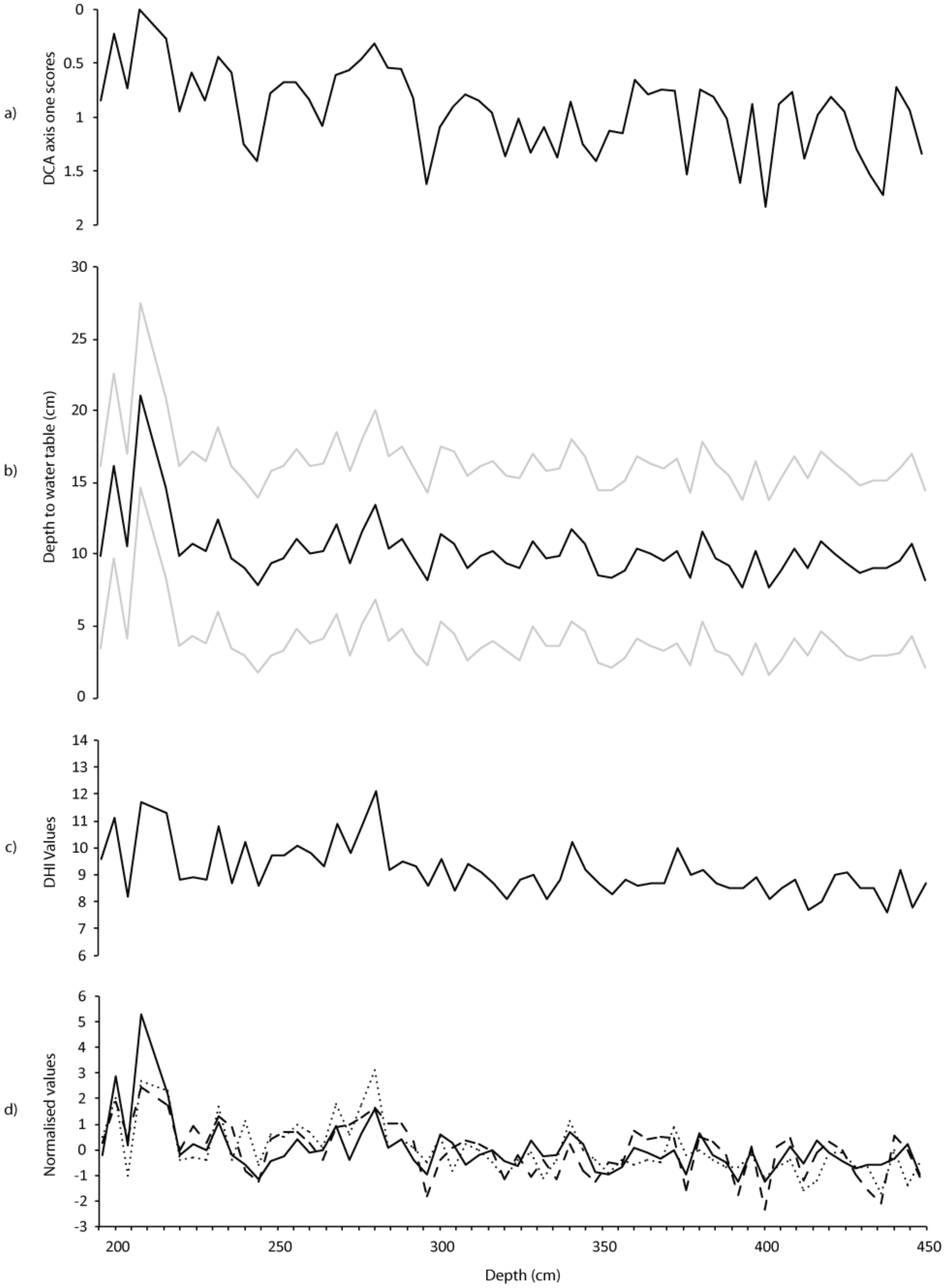
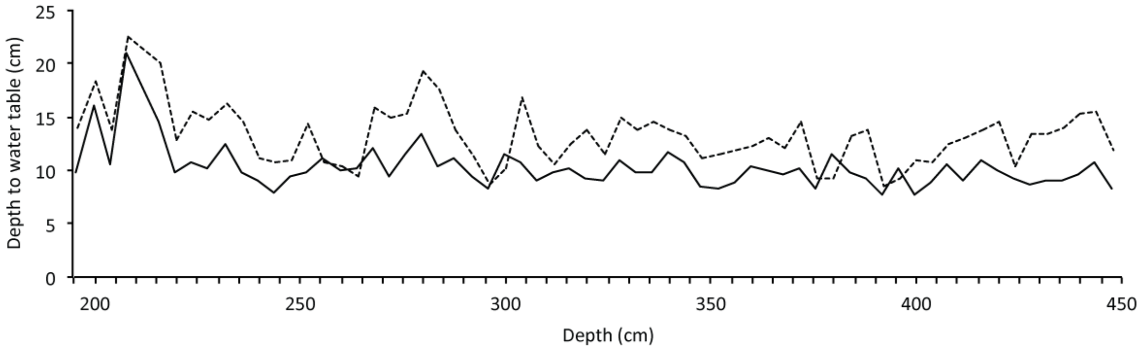


Figure 7.23 Comparison of transfer function water table reconstructions using the ACCROTELM (Charman *et al.*, 2007; solid) and North of Ireland (Swindles *et al.*, 2009; dashed) models for testate amoebae data at Fallahogy Bog. Errors are not shown for ease of reference. ACCROTELM reconstruction average error = 6.251 cm; North of Ireland reconstruction average error = 5.907cm.

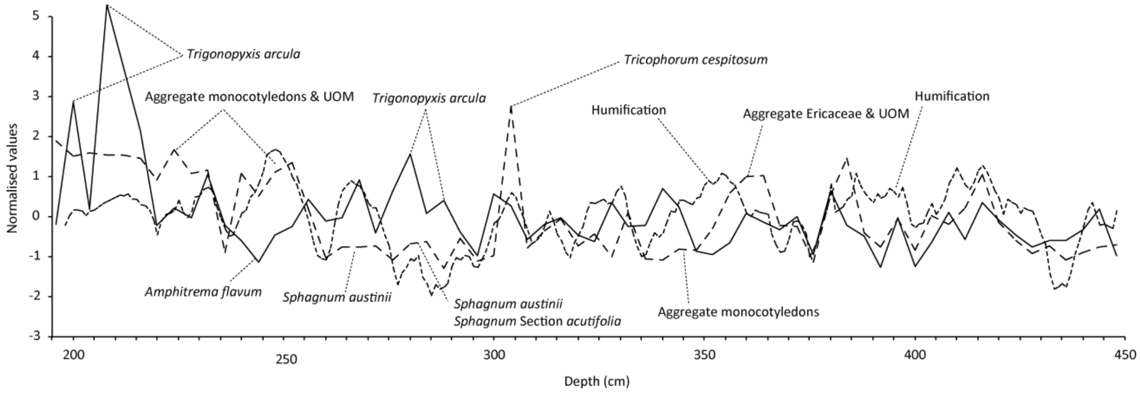


7.3.4 Multi-proxy comparison of palaeoecological records

Figure 7.24 presents a summary of the proxy reconstructions from plant macrofossil, humification and testate amoebae analyses at Fallahogy Bog, which have been normalised to aid comparison (cf. Blundell and Barber, 2005; Hughes *et al.*, 2006). The three palaeoecological proxies from Fallahogy Bog show only sporadic agreement, mostly limited to dry phases in the record. Records often appear to be uncorrelated, and occasionally in active disagreement. Notable periods of clear incoherence occur between 365-335 cm, 290-275 cm, and 255-240 cm.

Whilst the plant macrofossil record presents assemblages that are fairly typical of the region's oceanic raised bogs (Hughes and Barber, 2004), the periodically high amounts of UOM and a lack of *S.* s. *Cuspidata* would suggest that this was a relatively dry site during the period encompassed by this study. Both the other proxies support this hypothesis. Light transmission values are relatively low, indicating the presence of humified peats, and by inference, dry conditions. *D. pulex*, and indicator of drier conditions and/or fluctuating water tables overwhelmingly dominate the testate amoebae record at Fallahogy Bog, with very wet indicators, such as *A. wrightianum* failing to appear. In addition, *A. flavum*, a wetter indicator and one that frequently dominates testate amoebae records from the region, fails to establish any real form of dominance here. There are, however, a number of occasions during which the proxy summary curves could be said to be in broad agreement, displaying shifts towards wetter conditions (Fig. 7.24), of which the most marked occurs 416cm and persists, with the proxies not recovering to their previous levels until c. 380 cm, potentially extending the duration of this event, depending on interpretation. As already mentioned, a significant dry shift towards the end of the record is seen in all proxies. In the plant macrofossil record this shift begins 258 cm with a significant rise in UOM and decline in *Sphagnum*, a pattern supported by the humification record. In the testate amoebae record a broad trend to drier conditions appears to begin slightly earlier, at 296 cm, indicated by increasing *T. arcula* and further declining *A. flavum*. Again, discussing and comparing these proxy records, both individually and together, raises a number of issues with regard to the development of palaeoclimatic inference from these records, which will be discussed in Chapter 8.

Figure 7.24 Comparison of normalised data for testate amoebae-based reconstructed water table (solid), humification (finely dashed) and plant macrofossil (dashed) DCA normalised palaeoecological data at Fallahogy Bog, plotted against depth. Annotations are provided to explain disparity between the proxies.



7.3.5 Stable isotopes

Figure 7.26 presents the results of stable isotopic analysis of *Sphagnum* α-cellulose samples at Fallahogy Bog. Errors bars present 2 standard error ranges based on repeated measurements of the same sample. As at Sluggan Moss, triplicate measurements were taken for the majority of samples but, as detailed in Table 7.7, there were a small number of instances where only duplicate or single measurements were possible, due to small sample size and/or analytical error. Stratigraphical resolution of the analyses reflects: i) the distribution of suitable *Sphagnum* material in the plant macrofossil record; and ii) an effort to focus analyses around the marker horizon of the Hekla 4 tephra.

Table 7.7 Sample isotopic measurements per sample depth.

Sample depth (cm)	***Sample measurements***	
	$\delta^{18}O$	***$\delta^{13}C$***
220 cm	1	1
228 – 306 cm (*n* = 19)	3	1
256 cm	2	1
336 cm	2	2
344 cm	2	3
380 cm	2	2
All other samples (*n* = 19)	3	3

Figure 7.26a presents δ^{18}O values analysed from *Sphagnum* α-cellulose samples at this site. Analyses within the studied sequence exhibit no overall upward or downward trend δ^{18}O values. Values ranged from the highest excursion (26.02‰) at 306 cm (c. 4200 cal yr BP), to the lowest (21.77‰) at 396 cm (c. 5200 cal yr BP), producing a total range of isotopic variation in the record of 4.25‰. Apart from two larger (i.e. >1‰) excursions towards isotopically depleted values, and two towards enriched values, δ^{18}O values generally vary within ± 1‰ of 23.5‰. The two depleted excursions centred on 264 and 396 cm, and date to approximately 3800 and 5200 cal yr BP, respectively, rounded to the nearest 50 years. There are also two significant shifts towards isotopic depletion, which occur within normal variability, but are over 1‰ in magnitude. These centre on 332 and 224 cm, and date to c. 4400 and 3300 cal yr BP, respectively, rounded to the nearest 50

years. The two larger excursions towards enriched δ^{18}O values occur at 306 and 324 cm, and date to 4200 and 4350 cal yr BP, respectively. However, these excursions are associated with particularly high error bars, potentially indicating significant variation in intra-sample δ^{18}O measurement. Both samples were subject to measurements in triplicate and, upon closer inspection of these measurements, appear to have one outlying measurement. Such intra-sample variation is most likely caused by poor homogenisation of the α-cellulose fraction during laboratory preparation. Table 7.8 highlights these potential outliers and demonstrates that upon their exclusion, the average δ^{18}O value for each sample falls within the range of normal variability within the core (i.e. <1‰ of 23.5‰) for the sample at 324 cm, but not at 306 cm. This suggests that the isotopic excursion at 306 cm (c. 4200 cal yr BP) represents a genuine enrichment event, whereas the excursion at 324 cm (c. 4350 cal yr BP) is the result of analytical error.

Table 7.8 Problematic isotopic measurements per sample depth. Potential 'outliers' are italicised.

Sample depth (cm)	δ^{18}O measurements (‰)			Average (‰ ± 1σ)	
	1	*2*	*3*	*With 'outlier'*	*Without 'outlier'*
306 cm	27.165	27.362	*23.540*	26.02 ± 2.15	27.26 ± 0.14
324 cm	23.627	23.803	*26.749*	24.73 ± 1.75	23.72 ± 0.12

Figure 7.26b presents δ^{13}C values analysed from *Sphagnum* α-cellulose samples at this site. Note that the y-axis has been inverted for ease of comparison in Figure 7.25c. Analyses within the studied sequence exhibit a slight downward trend in δ^{13}C values (i.e. becoming more enriched). Values ranged from the highest (-26.25‰) at 344 cm (c. 4350 cal yr BP), to the lowest (-28.89‰) at 360 cm (c. 4600 cal yr BP), producing a total range of isotopic variation in the record of 2.64‰. Apart from three larger (i.e. >1‰) excursions towards isotopically enriched or depleted values, δ^{13}C values generally vary within ± 1‰ of -27.5‰. The two enriched excursions centred on 302 and 344 cm, and date to approximately 4200 and 4500 cal yr BP, respectively. However, there are also three significant shifts towards isotopic enrichment, which occur within normal variability, but are over 1‰ in magnitude. These centre on 280, 312 and 336 cm, and date to approximately 3950, 4250 and 4400 cal yr BP.

The enriched excursion at 344 cm possesses large error bars potentially indicating significant variation in intra-sample δ^{13}C measurements, probably caused by poor sample homogenisation. Table 7.9 demonstrates that no one measurement can be identified as a likely outlier and, therefore, the average δ^{13}C measurement (± 1σ error) must be accepted. As a result, caution must be adopted when interpreting and discussing this excursion.

Table 7.9 Problematic isotopic measurements per sample depth.

Sample depth (cm)	δ^{18}O measurements (‰)			Average (‰ ± 1σ)
	1	2	3	
344 cm	-25.017	-26.368	-27.35	-26.25 ± 1.17

The depleted excursion centres on 360 cm, and dates to approximately 4500 cal yr BP. In addition, there are also four significant shifts towards isotopic depletion, which occur within normal variability, but are over 1‰ in magnitude. These centre on 292, 328, 340 and 420 cm and date to 4100, 4350, 4450 and 5500 cal yr BP, respectively, rounded to the nearest 50 years. As a result, the section of the sequence 344 – 288 cm (c. 4450 – 4050 cal yr BP) can be classified as a period of generally higher isotopic variability.

When $\delta^{18}O$ and $\delta^{13}C$ values are plotted together against depth (Fig. 7.26c), little relationship appears to exist between the two curves. There are sporadic suggestions of coherence between the proxies. For example, between 344 and 300 cm the curves appear to operate within a lagged regime, with $\delta^{18}O$ lagging behind changes in $\delta^{13}C$ by approximately 30 years. However, as both measurements were taken from the same sample, the issue cannot be chronological in nature. Statistical analysis, by calculating Pearson product moment correlation coefficient (PPMCC), reveals only a very weak positive relationship, which is not statistically significant (PPMCC = -.156, $p < .05$; Fig. 7.25, Tab. 7.10), supporting the above suggestion that no coherent relationship exists between the respective depletion and enrichment of $\delta^{18}O$ and $\delta^{13}C$ data at Fallahogy Bog.

Figure 7.25 Biplot for Fallahogy Bog showing $\delta^{18}O$ and $\delta^{13}C$ data. A linear regression lines have been fitted for reference.

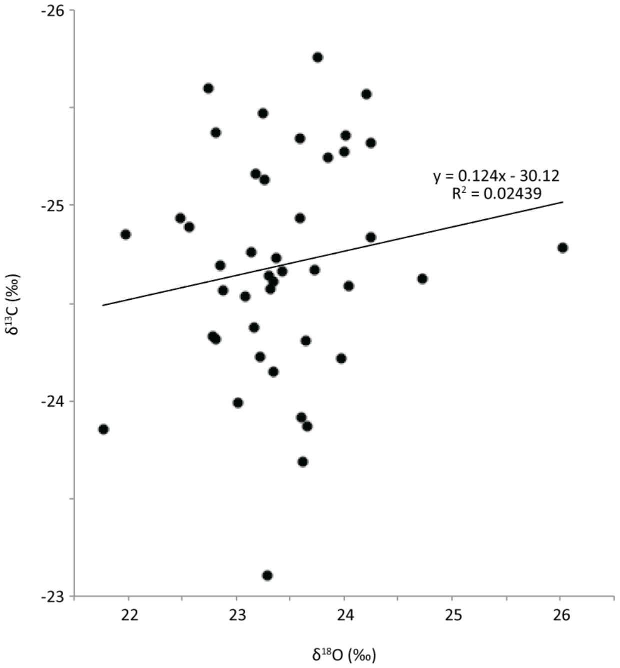
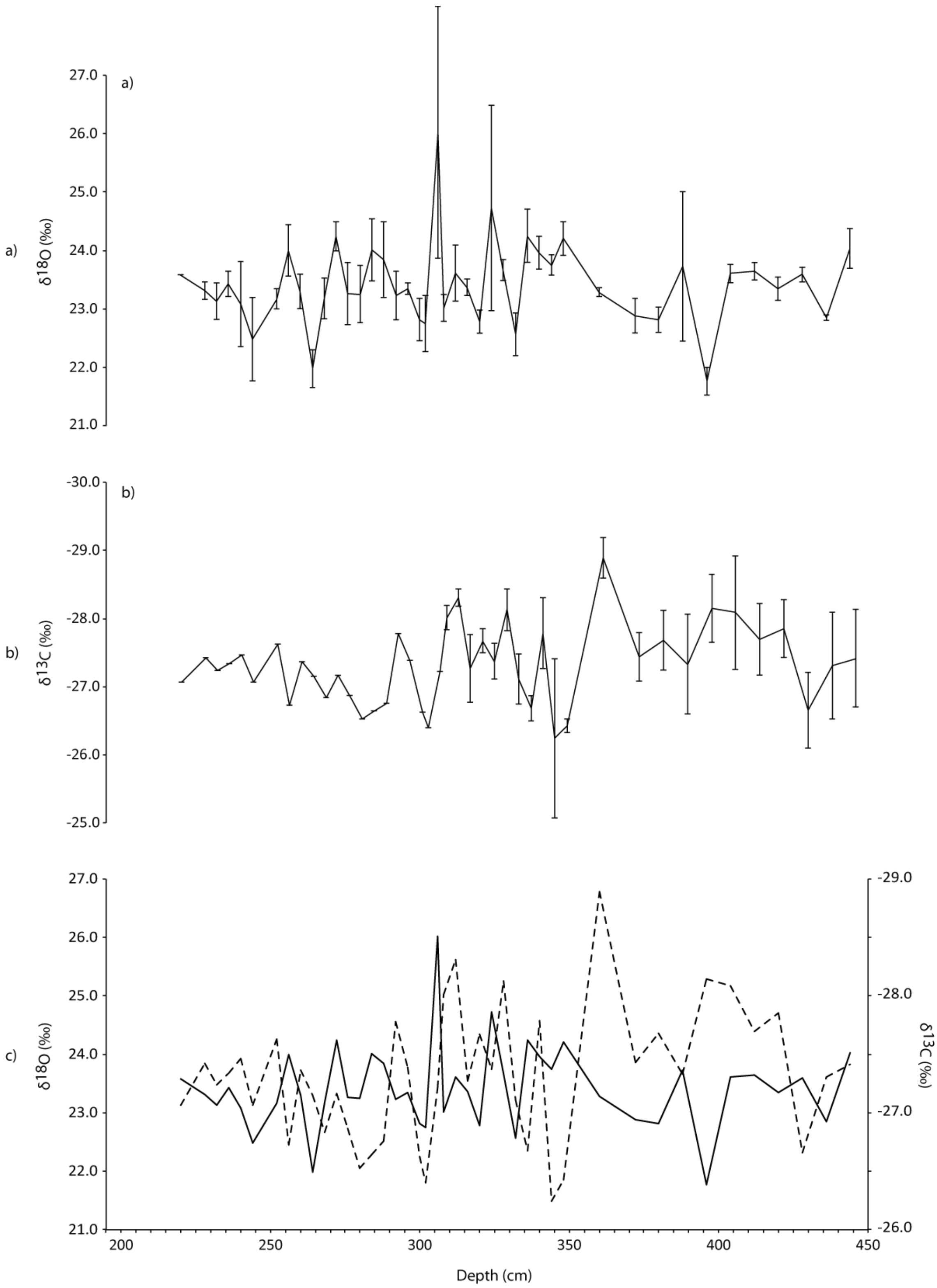


Table 7.10 Pearson product moment correlation coefficient of $\delta^{18}O$ and $\delta^{13}C$ isotope data from Fallahogy Bog.

	PPMCC	**Sig. (2-tailed, p<0.05)**	**n**	**R²**
All data	.156	.317	43	0.02439

Figure 7.26 Stable isotope data from Fallahogy Bog, as derived from *Sphagnum* α-cellulose, including: a) $\delta^{18}O$ and b) $\delta^{13}C$, where y-axis error bars denote 2σ ranges on replicate analyses from a single level where possible; c) comparison of $\delta^{18}O$ (solid) and $\delta^{13}C$ (dashed) data. All data are plotted against depth.



Chapter 8 – Discussion I

8.1 Introduction

In Chapter 1, a series of research questions are proposed:

1. Based on existing evidence, is it possible to characterise the timing and nature of a global '4.2 kyr event'?
2. Can a coherent '4.2 kyr event' signal be characterised in Great Britain and Ireland?
3. Can peat-based stable isotopic analysis contribute to our understanding of the nature and/or cause of the '4.2 kyr event'?

Chapter 2 sought to address the first research question by examining critically a large number of palaeoclimatic and palaeoenvironmental records for evidence of the event. It was concluded here that a global '4.2 kyr event' could be characterised, both climatically and temporally, but only in restricted regions of the globe. In other regions, and often at higher latitudes, a defined 'event' signal is less apparent. In northwestern Europe, and particularly in Great Britain and Ireland, a number of studies have found evidence for a period of significant climatic change between c. 4400 and 3500 cal yr BP (e.g. Anderson *et al.* 1998; Hughes *et al.*, 2000; Barber *et al.*, 2003). The precise nature and timing of these changes, however, is ambiguous and their relationship to a global 4.2 kyr event remains uncertain.

Consequently, the remainder of this thesis addresses the remaining two research questions. A novel methodology, presented in Chapter 5, aimed to combine a 'traditional' multi-proxy approach, incorporating plant macrofossil, humification and testate amoebae analyses, with simultaneous measurement of the stable isotopes of oxygen and carbon in *Sphagnum* α-cellulose (cf. Daley *et al.*, 2010; Woodley *et al.*, 2011). If interpretations of the $\delta^{18}O$ and $\delta^{13}C$ records derived from ombrotrophic, *Sphagnum*-rich bogs are correct (cf. Daley *et al.*, 2010; Loisel *et al.*, 2010), a record of past changes in atmospheric circulation and/or BSW should be preserved in the isotope records. The regional focus of this project, in northern Ireland, was chosen to: i) maximise the potential of an atmospheric circulation record (i.e. by proximity to the North Atlantic); and, ii) provide a key temporal marker (i.e. Hekla 4 tephra horizon) around which to focus analyses and construct a high-quality ^{14}C chronology (see Section 4.2.2).

This chapter will begin with a discussion of the palaeoecological data developed during the course of this study, building on the basic interpretations offered in Chapter 7. The isotopic data will then be discussed both in terms of its potential as a climatic record, and within the broader context of multi-proxy peat-based palaeoclimate studies. Palaeoecological and isotopic records from both sites will then be contextualised within both their regional and temporal settings, of northwestern Europe and the mid-Holocene, respectively, in Chapter 9. Where figures aim to compare multi-proxy data from the same site for methodological purposes, the x-axis is given as depth. Where data are presented on an inter-site basis, it is done so against a common age x-axis. As in Chapter 7, figures are formatted in such a way that upward shifts represent changes towards drier and/or warmer climatic conditions, whereas downward shifts represent changes towards wetter and/or cooler conditions, unless otherwise stated.

8.2 Palaeoecological evidence

Previously, composite BSW curves have been produced in an effort to summarise multi-proxy palaeoecological data (cf. Charman *et al*., 1999; Blundell and Barber, 2005; Langdon *et al*., 2005; Hughes *et al*., 2006; Daley and Barber, 2012). However, the equally weighted averaging of normalised plant macrofossil, humification and testate amoebae BSW/WTD reconstructions has the potential to reduce, or even remove, any climate signal present in the data, as proxy records, with potentially differing relationships with climatic drivers, are amalgamated. Similarly, the attribution of wet shifts, as defined by agreement between at least two of the three palaeoecological proxies (cf. Hughes *et al*., 2006), possesses significant potential to mask climatic signals (see Section 9.6.3).

Discussion of the palaeoecological data will subsequently be organised primarily by proxy, rather than by site, in order to: i) ascertain whether each proxy is presenting a regionally coherent BSW and hence climatic signal; and, ii) provide a basis for a methodological appraisal of each proxy technique. From this, it will be possible to assess: i) whether a 4.2 kyr event can be characterised in the region; and, ii) how peat records can be employed best to examine mid- to late-Holocene climatic events. Detailed palaeoecological and palaeoclimatic interpretations of each proxy record are presented in Chapter 7.

8.2.1 Plant macrofossils

8.2.1.1 Establishment of *Sphagnum austinii*-dominated assemblages

The plant macrofossil records from both sites in this study are heavily dominated by peat-forming species, *S. austinii*. This is consistent with other regional peat records, in which the oceanic species typically colonised northwest European bogs in the mid-Holocene during periods of cooler and/or wetter climatic conditions (Hughes *et al*., 2007), before experiencing a pan-European range contraction in the late-Holocene (van Geel and Middeldorp, 1988; Stoneman *et al*., 1993; Mauquoy and Barber, 1999; Langdon and Barber, 2005; Hughes *et al*., 2007). The establishment of a *S. austinii*-dominated assemblage has been described as indicating the establishment of a 'true' oceanic raised-bog community, often after a prolonged period of relatively dry *Eriophorum*/*Calluna* domination (Hughes and Barber, 2004), and is subsequently of great palaeoecological interest.

This pathway to ombrotrophy is clearly demonstrated at Sluggan Moss, where above ground *Eriophorum vaginatum* and *Calluna vulgaris*, accompanied by *Sphagnum* section *Acutifolia*, typify plant macrofossil zone SM-a (see Section 7.2.1). Such a composition is typical of pioneer assemblages in raised bog development (Hughes *et al*., 2007), which frequently provide the foundation necessary for maintaining a stable water mound on which *Sphagnum*-rich raised bogs can develop (Hughes and Barber, 2004). *S. austinii* subsequently became dominant c. 5850 cal yr.

BP at Sluggan Moss. Similarly, the studied sequence at Fallahogy Bog commences c. 5650 cal yr BP, with *S. austinii* clearly dominating in the first plant macrofossil zone, FAL-a.

This is notably earlier than the first instances of *S. austinii* dominance in Great Britain, including sites in Wales (e.g. Cors Canon; 4870 cal yr BP, Hughes and Schulz, 2001), the Angle-Scottish border region (e.g. Walton Moss; 4410-3990 cal yr BP, Hughes *et al.*, 2000; c. 4700 cal yr BP, Daley and Barber, 2012; Bolton Fell Moss; 4860 cal yr BP, Mauquoy and Barber, 1999; c. 5255 cal yr BP, Hughes and Barber, 2004; Raeburn Flow; 4790 cal yr BP, Mauquoy and Barber, 1999) and central Scotland (e.g. Temple Hill Moss; 4690 cal yr BP, Langdon and Barber, 2005). Extensive raised bog systems in Ireland are typically, although not exclusively, concentrated in two regions: the Bann and Main valley complex in the northeast, and the central plain and Shannon basin. Upland and lowland blanket bogs dominate other areas, particularly toward the west. Sites from the central concentration of raised bogs experience a range of dates for the first dominance of *S. austinii* (e.g. Mongan Bog, c. 2700 cal yr BP; Abbeyknockmoy Bog, c. 4830 cal yr BP; Hughes and Barber, 2004; Derryville Bog, c. 2350 cal yr BP; Ardkill Moss, min. c. 2150 cal yr BP; Blundell *et al.*, 2008), with some sites never witnessing significant dominance (e.g. Raheenmore; Daley, 2007).

The plant macrofossil records from this study are, to the author's knowledge, the first high-resolution, mid-Holocene records from the northeastern region. Other sites from the region have produced only low-resolution records dating from the late-Holocene. At two such sites, Glen West and Slieveanorra, *S. austinii* failed to establish dominance until the late Holocene, if at all (Swindles *et al.*, 2007b). At a third site, Dead Island Bog, *S. austinii* is present in varying quantities since at least c. 4500 cal yr BP (Swindles, unpub.). This early establishment of ombrotrophic raised bog assemblages in the region subsequently reinforces earlier assertions that palaeoecological and palaeoclimatic proxy data derived for the sites in this study provide an excellent opportunity to characterise any existent 4.2 kyr event signal.

8.2.1.2 Subsequent dominance of *Sphagnum austinii*

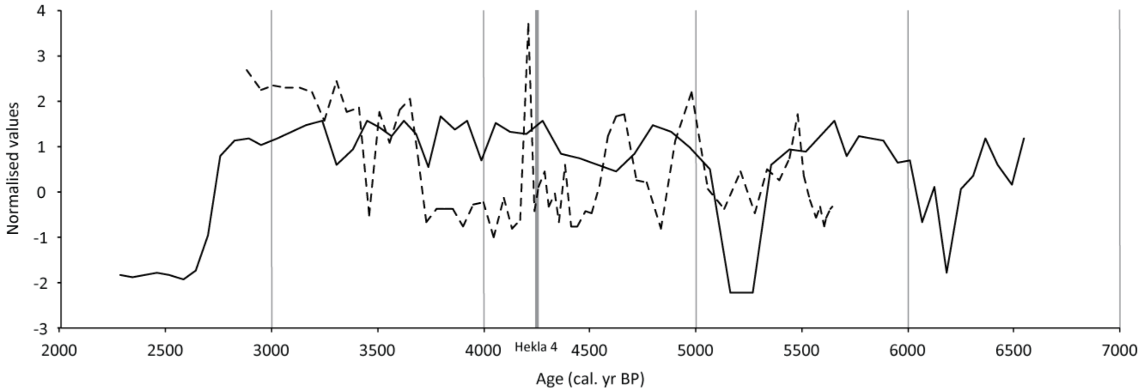
Whilst the early establishment of a *Sphagnum*-rich botanical assemblage is beneficial with regard to the formation of an ombrotrophic raised bog, capable of recording a climatic signal via a range of proxies, the dominance of single species, in this instance *S. austinii*, can be problematic. Owing to the relatively high water-table tolerance of *S. austinii*, records such as these are potentially climatically insensitive, demonstrating complacency towards small-scale changes of either short duration or low magnitude (e.g. Blundell and Barber, 2005; Hughes *et al.*, 2006). Barber *et al.* (1994) found plant macrofossil records to be at their most climatically sensitive when populated by a range of *Sphagnum* species, rather than a single eurytopic taxa.

Examination of the DCA axis one sample scores for Sluggan Moss (Fig. 7.7) in particular, supports an assertion of climatic complacency. Scores here closely mirror the proportion of *S. austinii* in the plant macrofossil record and, subsequently, significant climatic variation is only implied when other *Sphagnum* species proliferate. For the period of overlap (i.e. the duration of studied sequence at Fallahogy Bog), the largely complacent record presented here is, however, punctuated by a significant wet phase in zone SM-c (c. 5500 – 4900 cal yr BP), characterised by a near disappearance of ericaceous material and a substantial peak in *Sphagnum* section *Cuspidata*.

8.2.1.3 A regional climate signal?

As outlined in Chapter 4, the two sites are c. 30 km from one another and possess very similar topographical and environmental contexts. Therefore, for a climatic signal to be identified in the proxy records, there should be significant inter-site replication. Figure 8.1 plots plant macrofossil-based DCA BSW reconstructions from both sites against a common chronology. This demonstrates a lack of regional coherence between the records, implying that climatic signal within the records is weak. Consequently, a coherent regional palaeoclimatic reconstruction cannot be derived from the plant macrofossil data. Whilst no *Sphagnum* section *Cuspidata* was identified in the plant macrofossil record from Fallahogy Bog, the DCA reconstruction implies a slightly wetter phase, concurrent with the major shift towards wetter conditions described at Sluggan Moss c. 5500 – 5000 cal yr BP. It could therefore be argued that this phase represents the only period during which the regional plant macrofossil record could confidently be said to be responding to climate.

Figure 8.1 Plant macrofossil normalised DCA axis one scores from Sluggan Moss (solid) and Fallahogy Bog (dashed).



8.2.1.4 Quantitative reconstructions: DCA vs. DHI

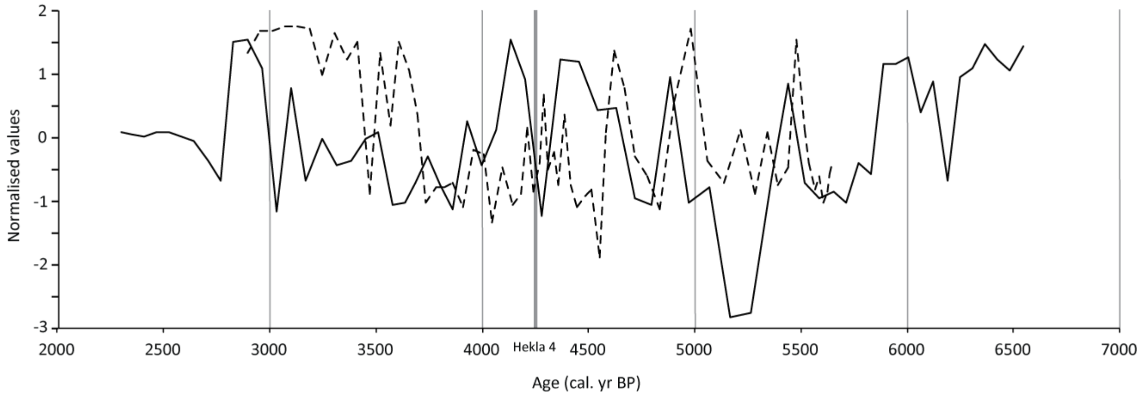
The benefits, limitations and rationale for producing a quantitative proxy-climate index using DCA from plant macrofossil data have been discussed (see Section 5.4.1) and many of these limitations are demonstrated here. Where a record is dominated by a single eurytopic species, other species can be treated as 'outliers' regardless of their own hydrological tolerance. If Sluggan Moss is taken as an example, the dominance of *S. austinii* is clearly artificially influential, as implied climatic variation only occurs when its dominance is challenged by other species. The technique also assumes a unimodal distribution, meaning that it can often fail to accurately represent species with a broad, and occasionally bi-modal, hydrological tolerances, such as *S. austinii* (Flatberg, 1986; Hill, 1988; Daniels and Eddy, 1990).

In addition, difficulties associated with identifying *Sphagnum* section *Acutifolia* to species level from sub-fossil material means that the section, which contains species of varying hydrological tolerances (Daniels and Eddy, 1990) must be assigned a single unimodal distribution within these analyses. At Sluggan Moss, there are two occasions during which members of the section proliferate.

These two instances occur within contrasting botanical contexts, indicative of drier (SM-a) and wetter (SM-h) conditions. In addition, these macrofossil zones are separated by c. 3500 cal yrs BP, presenting the possibility that material in these two zones represents two different species, each with their own ecological preferences, potentially negating the accuracy of the BSW reconstruction. In addition, the botanical composition of SM-h is so radically different (>95% *Sphagnum* section *Acutifolia*, near 0% UOM, ericaceous and monocotyledonous material) that, whilst not a crossed gradient (see Section 5.4.1.1), such significant compositional change is likely to have skewed the analyses. Issues associated with the DCA reconstructions also exist at Fallahogy Bog, where the susceptibility of the analysis to transient species domination is demonstrated by a sharp, ecologically unrealistic dry peak, occurring as the result of a short-lived dominance of *Trichophorum cespitosum* (c. 4150 cal yr BP).

The DHI index, an alternative method for developing quantitative BSW reconstructions from plant macrofossil data, is presented in Section 5.4.1.2. This method has been successfully employed in other studies, and a case has recently been made to suggest that DHI was capable of providing a statistically better reconstruction than DCA, when compared with other multi-proxy data (Daley and Barber, 2012). Figure 8.2 plots plant macrofossil-based DHI BSW reconstructions from both sites against a common chronology, demonstrating very little regional agreement between the records, further reinforcing the lack of a regional climatic signal in the plant macrofossil data. These curves, particularly at Sluggan Moss, demonstrate more variability than their DCA equivalents, and may subsequently, in the first instance, be interpreted as being more representative of BSW change. However, as outlined earlier, the principles of DCA are founded on a number of key assumptions, which do not apply in all situations. The broad hydrological tolerance of *S. austinii* and potentially bi-modal nature of the *Sphagnum* section *Acutifolia* grouping, found in these records, cannot be adequately addressed using the single weighting assigned to species in DHI. In addition, the 1-8 indicator value scale employed in DHI does not adequately reflect the relative distribution of species along the hydrological gradient in raised bog ecosystems. As a result of these fundamental discrepancies, no recommendation can be made that DHI should be accepted as a reliable climatic reconstruction ahead of DCA, and therefore the conclusion stands that, outside the period c. 5500 – 5000 cal yr BP, the plant macrofossil data from these sites do not present a regional climatic signal and should not be used to examine the timing and nature of the 4.2 kyr event.

Figure 8.2 Plant macrofossil normalised DHI scores from Sluggan Moss (solid) and Fallahogy Bog (dashed).



8.2.2 Peat humification

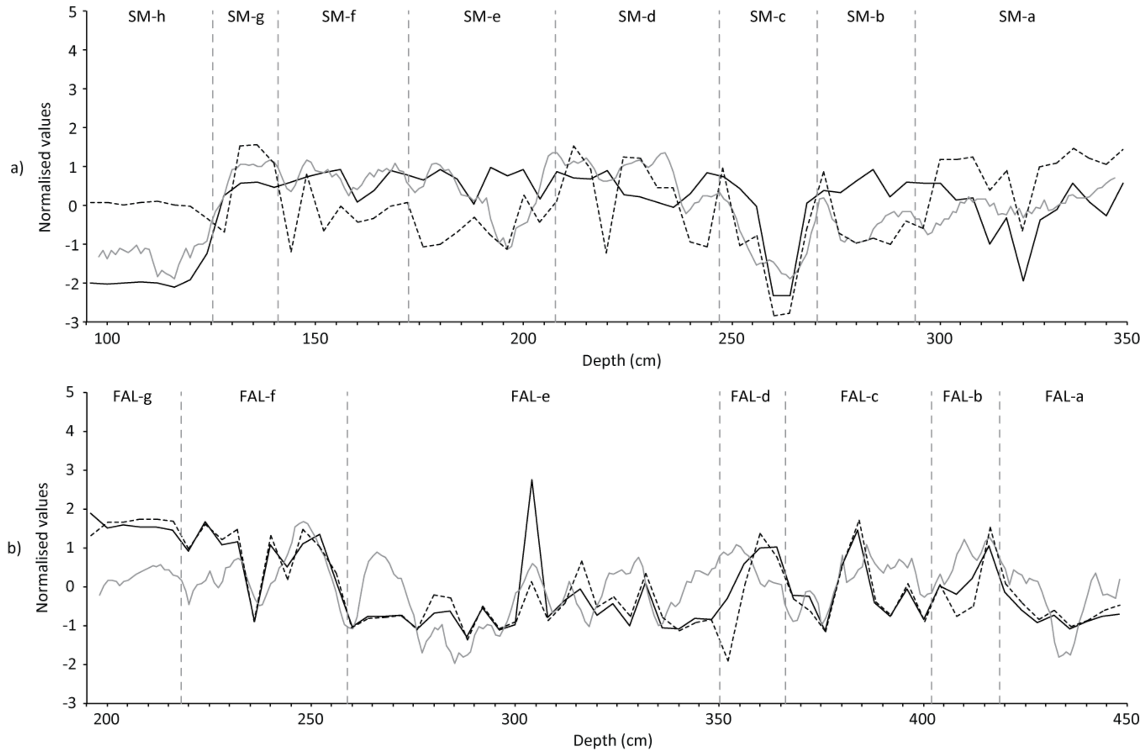
8.2.2.1 A viable palaeoclimatic tool?

As already demonstrated, the humification record from Sluggan Moss is characterised by moderate variability through the studied sequence (Fig. 7.5). There are, however, patterns in the data that correspond closely with the plant macrofossil based BSW reconstructions (Fig. 8.3a), particularly with the DCA axis one scores. The record of humification begins by indicating relatively dry conditions, consistent with the *Eriophorum*/*Calluna* dominated peat of SM-a. There is a slight wetting trend through zones SM-a and SM–b, again consistent with the emergence of *Sphagnum austinii* and the onset of the typical oceanic raised bog assemblage. This wetting trend continues into a notable wet period, coincident with the wettest plant macrofossil zone, SM-c. It is worth noting that this represents the single period of most agreement between normalised DHI, DCA and humification curves, centred around 5200 cal yr BP, providing further evidence of a major wet phase at this time.

Transmission values then indicate a return to drier levels as *S. austinii* regains dominance. The lowest (i.e. driest) transmission values occur coincidently with the sizeable monocotyledon and UOM peaks identified in SM-d, potentially highlighting a dry phase not recorded in DCA. A shift towards wetter conditions occurs coincidently with the reassertion of *S. austinii* dominance in SM-e, before continuing in a stable fashion, reflecting the continued *S. austinii*-dominated assemblages. Transmission levels then take a notable shift towards wetter values, coincident with the onset of the *Sphagnum* section *Acutifolia*-rich plant macrofossil zone, SM-h. This supports the suggestion that this zone represents a wet period and that, consequently, the *Sphagnum* taxa present are likely to be a more hygrophilous member of the section.

As at Sluggan Moss, the humification record from Fallahogy Bog is characterised by a high degree of variability through the studied sequence (Fig. 7.18), although there are patterns, which correspond well to the plant macrofossil data (Fig. 8.3b). In this instance, humification values are relatively and consistently low, supporting earlier assertions that this site was relatively dry. Figure 8.3b demonstrates that, at this site, humification appears to be responding more notably to dry shifts, rather than wet. The 5 point moving average humification curve shows good agreement with dry phases centring c. 416, 384, 360 and 304 cm, or c. 5500, 5000, 4600 and 4200 cal yr BP, respectively, characterised in the plant macrofossil record by reduction in *Sphagnum* and increases in aggregate Ericaceae, monocotyledons and UOM. The most notable dry shift in the humification record, as in the plant macrofossils, occurs at the beginning of FAL-f (c. 3700 cal yr BP). As at Sluggan Moss, the humification and plant macrofossil records are often in broad agreement, but never fully correspond, further emphasising the importance of considering raw plant macrofossil data when constructing palaeoclimatic inferences.

Figure 8.3 Normalised humification values (solid grey, 5-point moving average) plotted against depth with DHI (dashed) and DCA axis one scores (solid black) calculated from plant macrofossil data from a) Sluggan Moss, and b) Fallahogy Bog.



In recent years, the efficacy of the humification technique has been questioned, with suggestions that differential decay rates between species result the reflection of botanical changes, alongside or in place of any direct climatic signal (Yeloff and Mauquoy, 2006). Although there have been a number of studies in which variation in botanical composition does not appear to have had significant effects on the humification record (e.g. Chambers *et al.*, 1997; Blundell, 2002), transmission values at Sluggan Moss and Fallahogy Bog are shown to correspond strongly with the plant macrofossil record (Figs. 8.3), potentially indicating the presence of a 'species signal'. Figure 8.4 presents humification curves from Sluggan Moss (Fig. 8.4a) and Fallahogy Bog (Fig. 8.4b) against graphs depicting the changing composition of the plant macrofossil assemblage, between aggregate *Sphagnum*, monocotyledons, Ericaceae and UOM in an effort to examine for evidence of a species signal in the humification data. At both sites, the humification record responds strongly to botanical changes suggesting the presence of a 'species signal'.

Furthermore, statistical analysis, by calculating Pearson product moment correlation coefficient (PPMCC), reveals a moderately strong correlations between all humification values and plant macrofossil DCA axis one scores, which are statistically significant, at both Sluggan Moss (PMCC = -.602, $p < .05$; Fig. 8.5a, Tab. 8.1a) and Fallahogy Bog (PPMCC = .441, $p < .05$; Fig. 8.5b, Tab. 8.1b).

Figure 8.4 Normalised humification curves (N.B. Inverted axes, negative = less humified; positive = more humified) from a) Sluggan Moss and b) Fallahogy Bog, plotted against aggregate *Sphagnum* (dark grey), monocotyledon (mid-grey), ericaceous (light-grey) remains and UOM (white), on a common age axis.

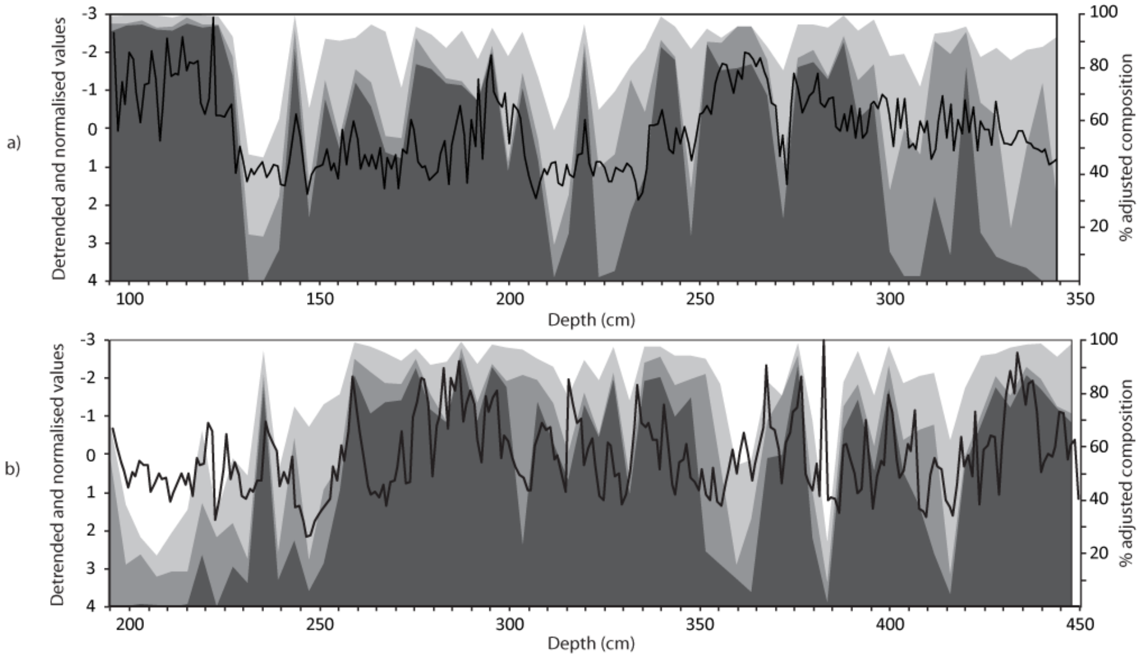


Figure 8.5 Biplot of humification values against plant macrofossil DCA axis one scores for a) Sluggan Moss and b) Fallahogy Bog. Linear regression lines have been fitted for reference.

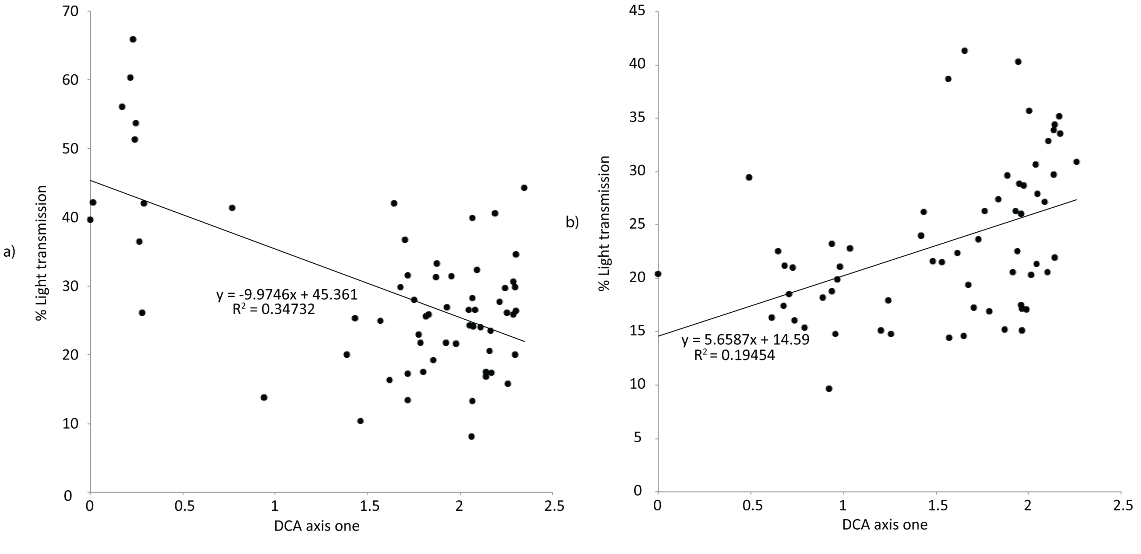
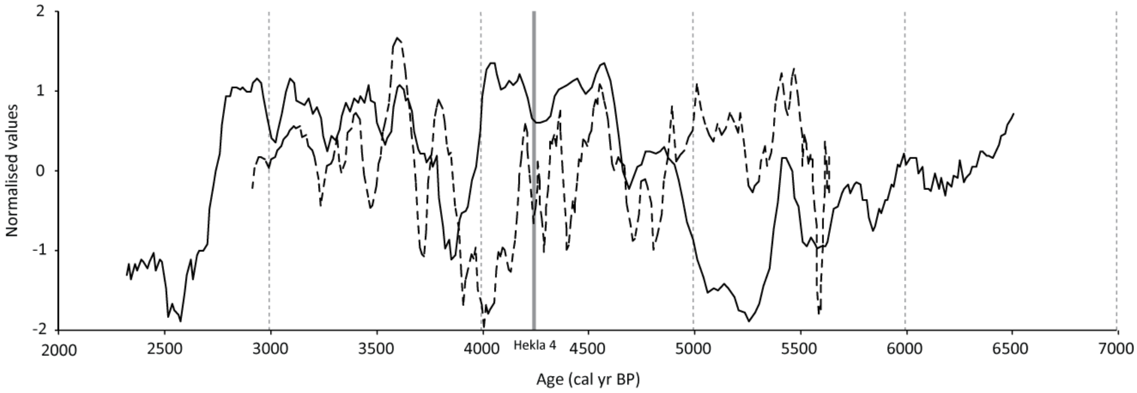


Table 8.1 Pearson product moment correlation coefficient of $\delta^{18}O$ and $\delta^{13}C$ from Sluggan Moss. Asterisks denote statistical significance.

	PPMCC	**Sig. (2-tailed, p<0.05)**	**n**	**R²**
a) Sluggan Moss	-0.602	.000*	63	0.34732
b) Fallahogy Bog	0.441	.000*	64	0.19454

Figure 8.6 Normalised humification curves (5-point moving average) from Sluggan Moss (solid) and Fallahogy Bog (dashed) on a common age axis.



Section 8.2.1.3 has already demonstrated that the plant macrofossils from these sites present only a weak climate signal. The corresponding humification records, strongly affected by botanical composition, are therefore unlikely to present a coherent regional climatic signal. By plotting both records on a common age axis, it becomes clear that only limited regional coherence exists (Fig. 8.6).

A simultaneous shift towards wetter conditions is implied c. 5400 cal yr BP, coincident with the major wet phase implied in the plant macrofossil record. Two other phases of broad agreement occur, with shifts towards drier conditions occurring at c. 4650 cal yr BP at both sites, and 4000 and 3800 cal yr BP at Fallahogy Bog and Sluggan Moss, respectively. However, the overall lack of coherence between these two records implies that the humification analyses employed here has not yielded a reliable climatic signal, and so should not be used to examine the timing and nature of the 4.2 kyr event.

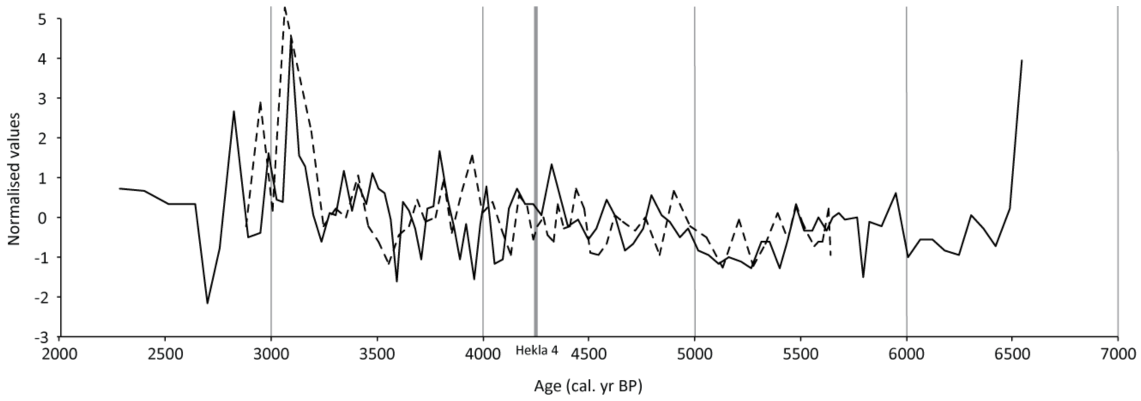
8.2.3 Testate amoebae

8.2.3.1 A regional signal?

Figure 8.7 demonstrates that of all the palaeoecological proxy data developed within this study, the testate amoebae records possess the most regional coherence, thus suggesting the presence of a climatic signal.

In both records, shifts toward a prolonged wetter phase, coincident with changes seen in the plant macrofossil and humification records, are observed during the period 5500 – 5000 cal yr BP, further supporting the attribution of this period to a significant shift in the prevailing climate to wetter and/or cooler conditions. In addition, broadly coincident shifts towards wetter conditions occur c. 4900/4800 cal yr BP, c. 4100 cal yr BP, c. 4000/3900 cal yr BP, c. 3100 cal yr BP and c. 3000/2900 cal yr BP. In addition, a coincident major dry shift is seen in both records c. 3200 cal yr BP.

Figure 8.7 Testate amoebae inferred WTD reconstructions from Sluggan Moss (solid) and Fallahogy Bog (dashed) on a common age axis.



Whilst agreement between the two sites is generally good, there also appears to be a degree of lead/lag between the records. However, these chronological differences fall well within the centennial-scale error margins associated with the age-depth models (see Chapter 6). As a result, it is likely that these are merely chronological discrepancies, rather than representations of truly time-transgressive climatic events. As a result, the testate amoebae records from these sites can be accepted as records of palaeoclimatic change (see Section 8.2.3.3).

8.2.3.2 Taxon dominance

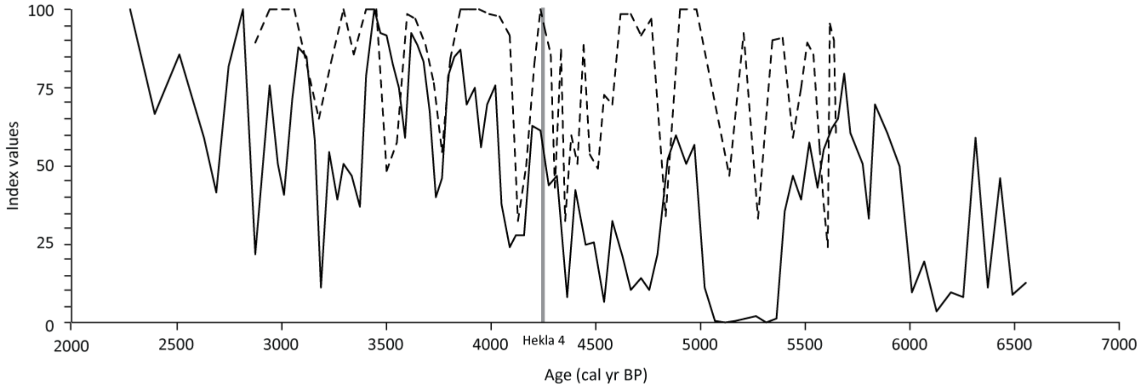
A common theme, at both Sluggan Moss and Fallahogy Bog, is the co-dominance of *Amphitrema flavum* and *Difflugia pulex*. Both taxa have been shown to possess relatively broad hydrological tolerances, albeit with the latter generally considered an indicator of slightly drier conditions (Charman *et al*., 2000, 2007), and both are often ubiquitous in bog records from the region (e.g. Swindles *et al*, 2010a). The problems associated with the domination of a small number of eurytopic taxa has been discussed previously, with reference to plant macrofossil records (e.g. Barber *et al*., 1994; Blundell and Barber, 2005; Hughes *et al*., 2006), and these issues can be extended to apply to testate amoebae records. As outlined in Section 8.2.1.2, such assemblages can often exhibit a degree of climatic complacency. However, as Section 8.2.3.1 has shown, the testate amoebae records from Sluggan Moss and Fallahogy Bog exhibit a degree of regional coherence, indicating the presence of a regional climatic signal. It appears that in spite of the dominance of *A. flavum* and *D. pulex*, the testate amoebae record from the two sites is not climatically complacent. However, it is possible that the relatively small magnitudes of the changes recorded are the result of a signal muted by the dominance of these two species.

The transfer function approach used for testate amoebae records is based on observed ecological data, notably depth to water table preferences, for each taxon (e.g. Charman *et al*., 2000, 2007), and is arguably a more robust reconstruction of palaeohydrological change than other methods, such as the DCA or DHI approaches more commonly used for plant macrofossil data. However, as discussed in Section 3.4.3.2, *D. pulex* is severely underrepresented in modern surface samples, leading to a degree of uncertainty as to its precise ecological preferences. Although the taxon is generally considered to be relatively eurytopic, with a tendency towards drier conditions, Sullivan and Booth

(2011) have suggested that *D. pulex* can be interpreted as an indicator of fluctuating water table conditions, further complicating the interpretation of records in which the taxon is ubiquitous. Conversely, they also suggest that *A. flavum* should be interpreted as being representative of stable conditions. The alternating dominance of these two taxa in both these records suggests that this interpretation could be accurate. However, no more effective test of this hypothesis can be gleaned from these data due to the lack of an independent record of water table stability. Such uncertainties surrounding the ecology of a species so prevalent could potentially mute any climatic signal present. It is possible that this has occurred here, as evidenced by the relatively small magnitudes of change discussed earlier, but not to an extent that the climatic signal is masked completely.

In light of Sullivan and Booth's (2011) interpretation, Figure 8.8 plots the ratio of *D. pulex* to *A. flavum* as a potential index of water table stability at each site. Whilst no obvious regional consistency exists, which could be expected if this index were valid, it is difficult to quantify to what degree autogenic ecohydrological response (cf. Swindles *et al.*, 2012a) may be influencing any potential water table stability signal. As a result, further investigation into the ability of testate amoebae to reconstruct past changes in water table stability is required.

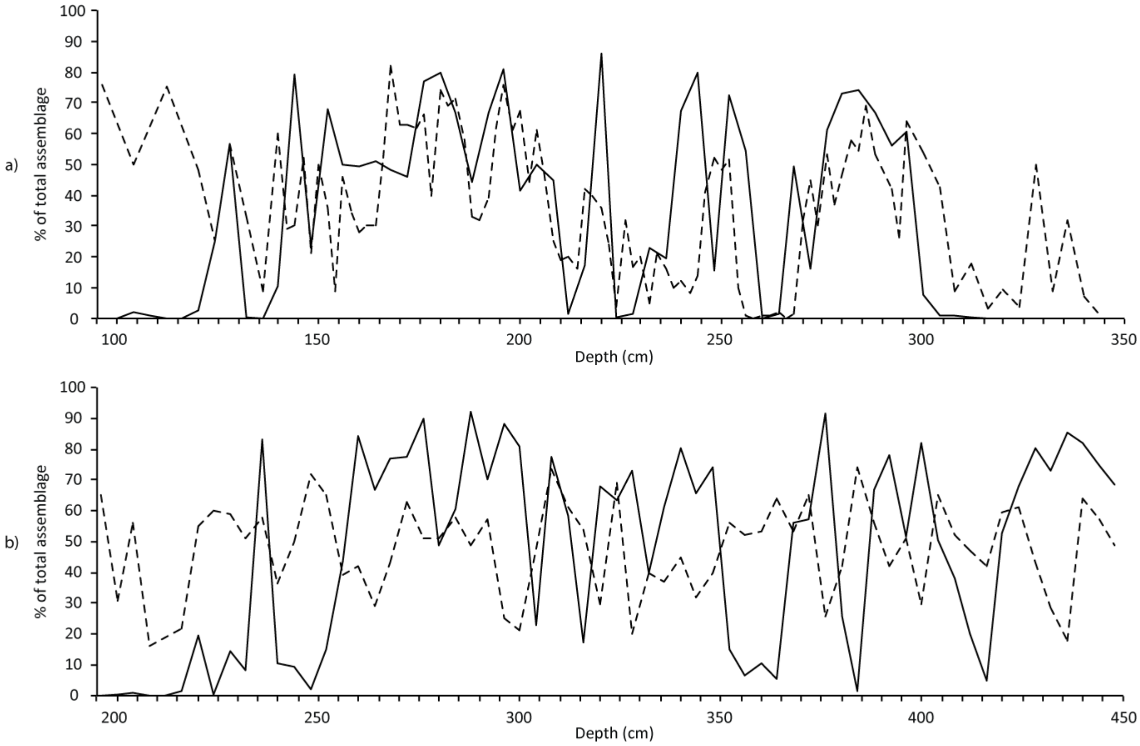
Figure 8.8 Comparison of *Difflugia pulex*:*Amphitrema flavum* ratio as an index of variability from Sluggan Moss (solid) and Fallahogy Bog (dashed). Higher values represent higher proportions of *D. pulex*, indicating increased variability (cf. Sullivan and Booth, 2011).



The broad-ranging hydrological distributions enjoyed by both *D. pulex* and *Sphagnum austinii* (see Section 8.2.1.2) has led to informal suggestions that the testate amoebae enjoys some form of commensalistic relationship with the bryophyte, owing possibly to a favourable micro-ecosystem presented by the plants structure (see Section 7.2.3). Figure 8.9 presents a comparison of the varying proportions of *D. pulex* and *S. austinii* as a percentage of total testate amoebae and plant macrofossil assemblage, respectively. However, the data here demonstrate no significant correlation, suggesting that no relationship exists between the two taxa. Some similarity exists at Sluggan Moss, but as *D. pulex* thrives in periods where *S. austinii* is absent, it is likely that this coherence is merely the result of a similarity in the response of each taxon to changes in BSW/WTD.

Generally, the testate amoebae records from both Sluggan Moss and Fallahogy Bog are considered here to be robust, yielding a coherent regional climatic signal and, as result, can be used to assess for the presence, timing and nature of a significant climatic event c. 4.2 kyr BP.

Figure 8.9 Comparison of *Difflugia pulex* (dashed) and *Sphagnum austinii* (solid) as percentage total assemblage from a) Sluggan Moss and b) Fallahogy Bog.



8.2.3.3 Tuning and stacking: a regional palaeoclimatic curve

The importance of obtaining multiple records from the same region to combat the inherent uncertainties of any single proxy climate record is emphasised by Charman *et al.* (2006). The technique of 'tuning and stacking' is used to address a number of problems typically encountered when interpreting and reconciling multiple records from a region, including chronological differences, locally specific ecosystem noise and analytical error. Palaeoclimatological records are commonly tuned in order to dated records to provide additional chronological control (e.g. Shackleton, 2000; Bond *et al.*, 2001; Hughen *et al.*, 2006), but the technique's application to peat records is more recent (Charman *et al.*, 2006; Blundell *et al.*, 2008; Swindles *et al.*, 2010a). Prior to this, studies have compared the timing of wet and dry conditions between sites to assess for synchronicity of change (e.g. Hughes *et al.*, 2000; Barber and Charman, 2003; Langdon *et al.*, 2003), often with inconclusive results, thus providing a rationale for the development of a more advanced technique of comparison. Previously, tuned and stacked records have used both single proxy records (e.g. testate amoebae, Charman *et al.*, 2006; Swindles *et al.*, 2010a) and inferred BSW curves based on stacked multi-proxy records (e.g. Blundell *et al.*, 2008). In addition, regional, tuned and stacked composite BSW curves have been constructed using various numbers of sites (e.g. 12, Charman *et al.*, 2006; 2, Blundell *et al.*, 2008; Swindles *et al.*, 2010a). Langdon and Barber (2005) also presented a stacked, but untuned, multi-proxy BSW record from 7 sites in Scotland. Rationale for the selection of sites to combine should be based on their distribution, chronological quality, and methodological similarity (Charman *et al.*, 2006).

In line with the research aims, this study attempted to construct a regional climatic reconstruction for the time period encompassing the 4.2 kyr event, using sites possessing greatest potential for climatic sensitivity and robust chronology. As discussed previously, only the testate amoebae data provide a robust climatic signal, demonstrating clear correspondence of the main shifts in surface wetness (Fig. 8.8). There are, however, slight differences in the magnitude and timing of the shifts. The latter is not unexpected given the potential for chronological error in any chronology derived from radiocarbon dating (up to ± 150 cal years for AMS ^{14}C ages). Acknowledging age-depth models as best age estimates allows us to tune chronologies to one another, facilitating the stacking of records to produce a regionally combined palaeoclimatic curve.

This study followed the approach of previous methodologies for the tuning and stacking of climatic records (e.g. Charman *et al*., 2006; Blundell *et al*., 2008; Swindles *et al*., 2010a). After conversion to normalised values, key points of change, based on peaks and troughs, were identified in each record, with an average age calculated for each 'match point' based on the original age-depth models presented in Chapter 6. An adjusted age-depth model was produced based on linear interpolation, in preference to polynomial and regression models (cf. Telford *et al*., 2004), between these match points. The revised chronology also included the Hekla 4 tephra isochron as an independent and precise pinning point (e.g. Langdon and Barber, 2004).

Charman *et al*. (2006) state that whilst this process cannot guarantee that the revised chronology is any nearer to the true age-depth relationship, it is important that the age estimates are consistent with the original age-depth model for any of the incorporated records. Subsequently, the revised chronology must be plotted alongside each original chronology, to ensure that it falls within the initial age estimates (at 2σ error). In addition, it is imperative that the revised model is inspected for unlikely accumulation rate changes (Blaauw, 2012).

In a review of the tuning and stacking process, Swindles *et al*. (2012c) emphasise the importance of presenting alternative tuning scenarios, based on plausible combinations of match points. Two possible scenarios are presented here in Figures 8.10 and 8.12, with match point ages detailed in Tables 8.2 and 8.3. Finally, comparisons of each resultant revised age-depth models with the original age-depth models from both sites (Figs. 8.11 and 8.13).

Figure 8.11 demonstrates that the initial revised age-depth model, calculated from match points between records detailed in Figure 8.10 and Table 8.2 shows some disagreement with the original age-depth models from both sites, and can be rejected. Figure 8.13 demonstrates that the revised age-depth model, calculated from match points between records detailed in Figure 8.12 and Table 8.3, shows good agreement with the original age-depth models from both sites, consistently falling within the latter's estimated error margins. It is therefore suitable for stacking the two records. The composite record was produced by merging data from each record using a 100-year moving average. By combining these records, it is hoped that long-term climate signals are retained, whilst reducing chronological and ecological uncertainties specific to each site's record (Charman *et al*., 2006; Blundell *et al*., 2008). As the studied sequence is temporally shorter at Fallahogy Bog, values from Sluggan Moss were taken as being representative of the region for the periods at either end of the combined record. The result, presented in Figure 8.14, is a regional palaeoclimate curve spanning approximately 4000 years, between 2500 – 6500 cal yr BP.

Figure 8.10 Normalised water table reconstructions from a) Sluggan Moss and b) Fallahogy Bog and tuning match points (A-J) used to tune the chronologies. Hekla 4 tephra is labelled. Negative values indicate wetter conditions.

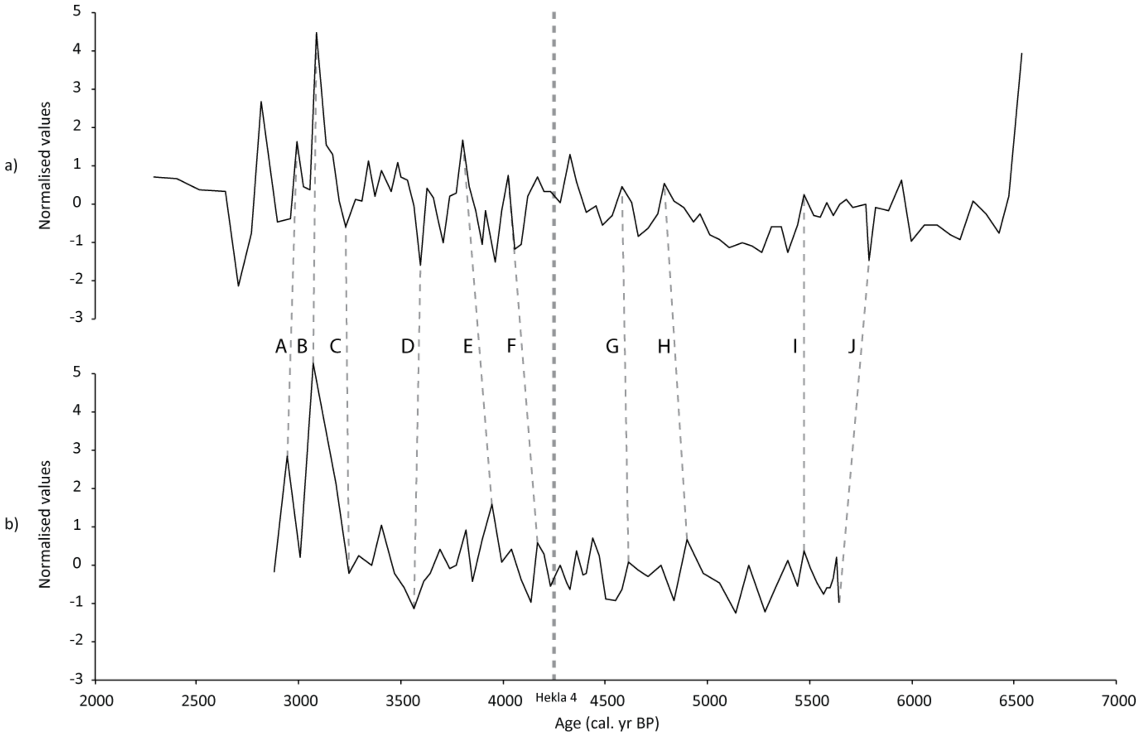


Table 8.2 Details of tuning match points outlined in Figure 8.10. Ages from each site represent best estimates as calculated by *Clam* (Blaauw, 2010); see Chapter 6 for more details.

Match point	Sluggan Moss	Fallahogy Bog	Average	Difference
A	2981	2940	2960.5	41
B	3085	3061	3073	24
C	3230	3238	3234	8
D	3593	3555	3574	38
E	3796	3947	3871.5	151
F	4018	4171	4094.5	153
Hekla 4	4287	4287	4287	0
G	4581	4620	4600.5	39
H	4797	4902	4849.5	105
I	5483	5479	5481	4
J	5803	5649	5726	154

Figure 8.11 Comparison of the revised chronologies developed from match points outlined in Table 8.2 (red), plotted against the original age-depth models (black = best estimate; grey = 2σ error) for a) Sluggan Moss and b) Fallahogy Bog.

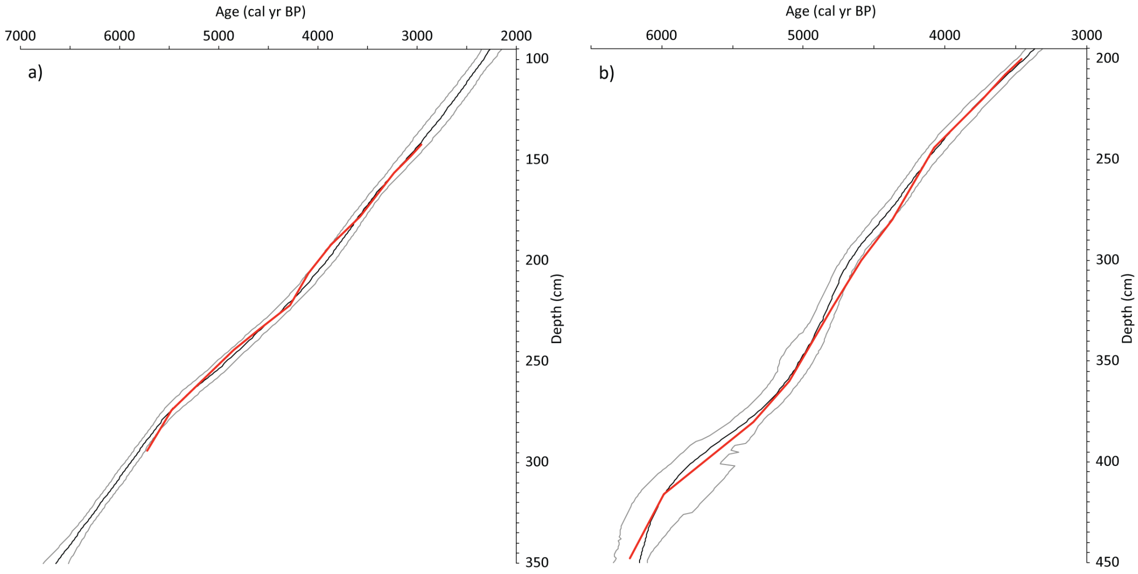


Figure 8.12 Normalised water table reconstructions from a) Sluggan Moss and b) Fallahogy Bog and tuning match points (A-J) used to tune the chronologies. Hekla 4 tephra is labelled. Negative values indicate wetter conditions.

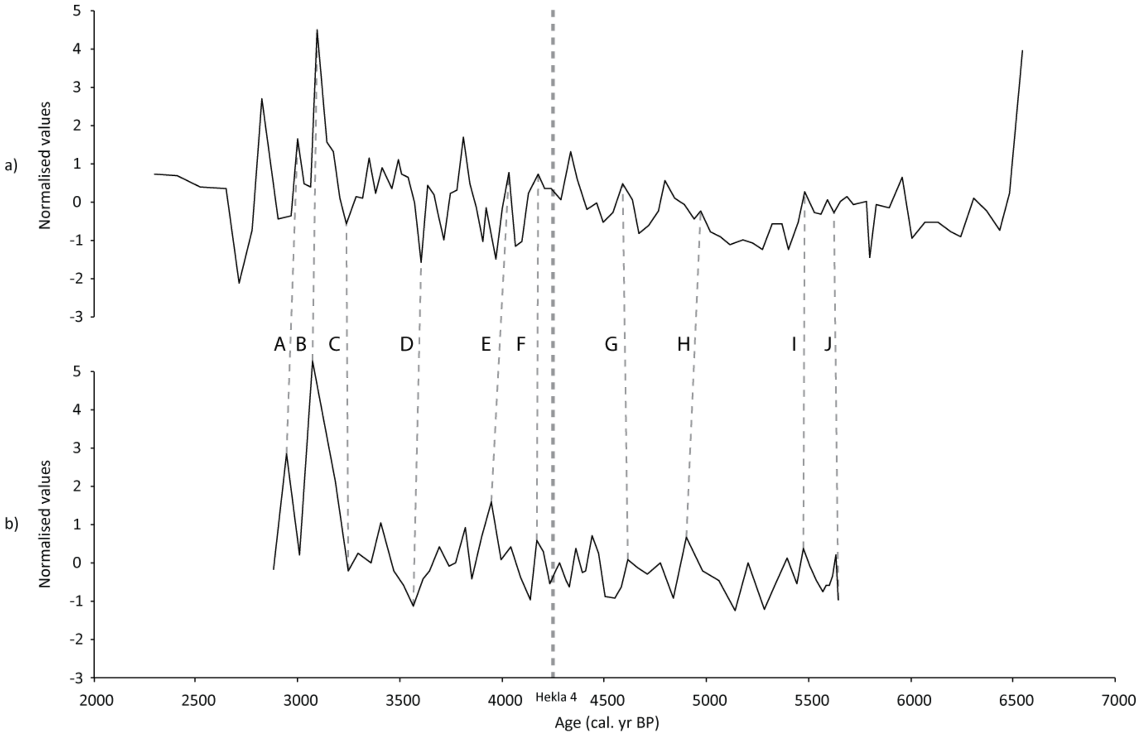


Table 8.3 Details of tuning match points outlined in Figure 8.12. Ages from each site represent best estimates as calculated by *Clam* (Blaauw, 2010); see Chapter 6 for more details.

Match point	Sluggan Moss	Fallahogy Bog	Average	Difference
A	2981	2940	2960.5	41
B	3085	3061	3073	24
C	3230	3238	3234	8
D	3593	3555	3574	38
E	4018	3947	3982.5	71
F	4161	4171	4166	10
Hekla 4	4287	4287	4287	0
G	4581	4620	4600.5	39
H	4973	4902	4937.5	71
I	5483	5479	5481	4
J	5622	5649	5635.5	27

Figure 8.13 Comparison of the revised chronologies developed from match points outlined in Table 8.3 (red), plotted against the original age-depth models (black = best estimate; grey = 2σ error) for a) Sluggan Moss and b) Fallahogy Bog.

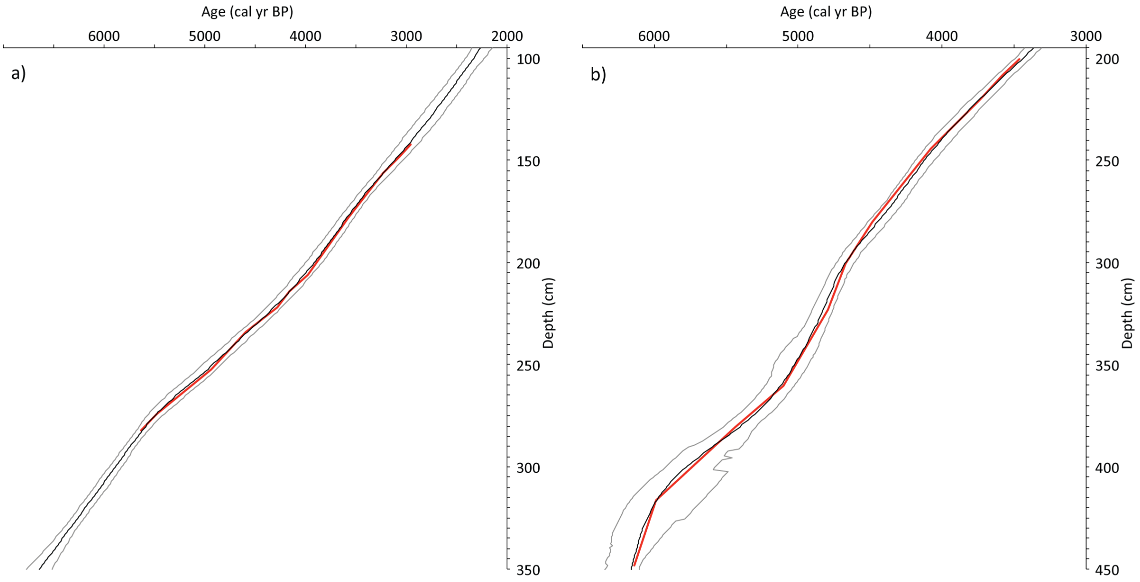
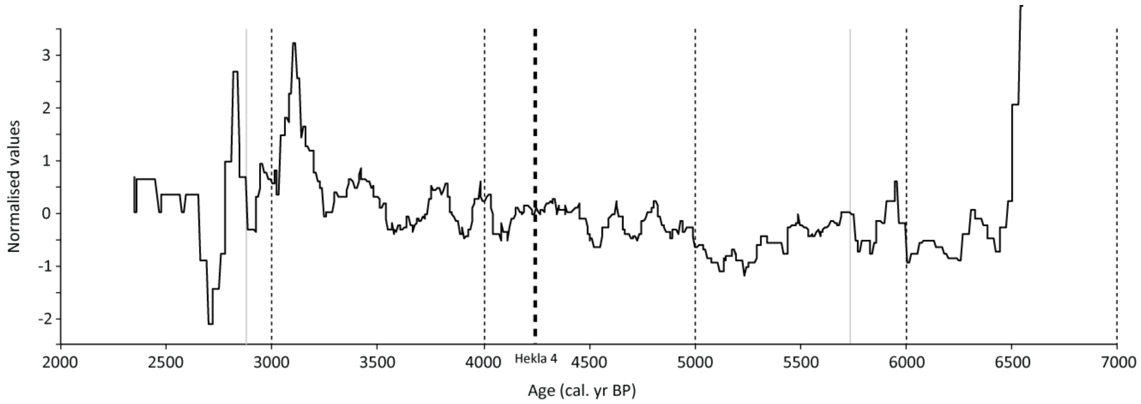


Figure 8.14 A regional composite water table record for northern Ireland, based on testate amoebae data from Sluggan Moss and Fallahogy Bog. Section during which records from both sites overlap lies between grey lines.



The tuning and stacking approach is not without limitations. Blaauw (2012) describes a degree of subjectivity that exists in the assignation of match points, based on major and minor peaks, along with the general shape of the proxy curve. However, attempts were made here to minimise the degree to which the original age-depth models were 'stretched'. Table 8.3 demonstrates that the average difference between the two original dates for each match point was significantly reduced when compared to the first tuning attempt outlined in Table 8.2, thus rendering the revised model a more robust age approximation. In addition, the presence of an objective, precisely dated, geochemically identified tephra layer affords a large degree of chronological integrity to the combined record (Blaauw, 2012).

Blaauw (2012) also warns against the tuning and stacking of records from a wide spatial distribution, suggesting that records should only be combined within regions that have been shown to possess the same climatic regime. As McCarroll (2010) outlines, precipitation patterns can significant vary spatially, especially with complex topographical influences, and as a result stacking of records should only be undertaken over relatively small areas. As discussed in Chapter 4, the geographic context and climatic data of these two sites are very similar and representative of the region. In addition, other studies have tuned and stacked similar peat-based proxy records from other records, comparable spatially and temporally, achieving successful results (e.g. Charman *et al.*, 2006; Blundell *et al.*, 2008; Swindles *et al.*, 2010a). As a result, whilst bearing in mind the potential for chronological and reconstructive error (Swindles *et al.*, 2012c), the tuned and stacked record developed here can be presented as a regional palaeoclimatic curve, overcoming problems that beset the correlation of such records, and providing a more reliable basis for comparisons with other regional records (Charman *et al.*, 2006).

8.3 Discussion of the stable isotopic data

In Section 3.4.4 the proposed mechanisms governing variation of $\delta^{18}O$ and $\delta^{13}C$ in *Sphagnum* α-cellulose are summarised. The isotopic data from Sluggan Moss and Fallahogy Bog will subsequently be discussed in that context.

8.3.1 $\delta^{18}O$ data

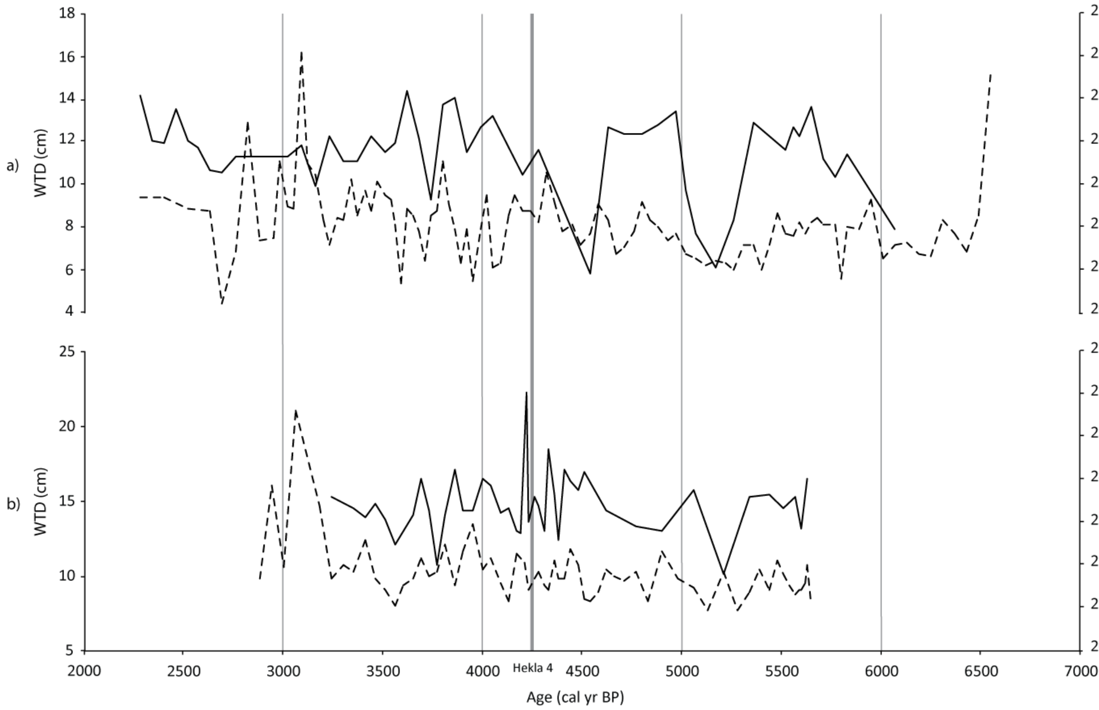
A modern calibration dataset compiled by Daley *et al.* (2010) suggests that water used for *Sphagnum* cellulose synthesis is sourced from precipitation and therefore captures the $\delta^{18}O$ signal of meteoric water. This can be explained by the mechanically simple pathway through which the plant assimilates water. $\delta^{18}O$ of meteoric precipitation is strongly influenced by water vapour source and trajectory history, thus enabling the reconstruction of past atmospheric circulation (Anderson *et al.*, 2002; Vinther *et al.*, 2003; Burnett *et al.*, 2004). The variables concerned, which include the temperature at which the water condenses, the amount of precipitation that has previously occurred, the isotopic value of the water vapour source and the extent to which the water has travelled over land (Dansgaard, 1964; Rozanski *et al.*, 1993; Araguás-Araguás *et al.*, 2000), are all strongly linked to storm system dynamics (Burnett *et al.*, 2004). As previously discussed, Baldini *et al.* (2010) have demonstrated that $\delta^{18}O$ of precipitation in Ireland is heavily influenced by air mass history and atmospheric circulation. Therefore, in the context of the western seaboard of Atlantic northwest Europe variation in the $\delta^{18}O$ of *Sphagnum* α-cellulose could reasonably be expected to reflect changing intensity and migration of the westerly storm belt. However, a precise mechanism for the interpretation of $\delta^{18}O$ records of this type has not been developed, although oxygen isotope variability in modern waters collected from three sites in northwest Europe found that lower $\delta^{18}O$ values were found during the generally cooler, wetter winter months, in contrast to the higher values experienced in the warmer, drier summer months (Daley *et al*., 2010). Extrapolation of this pattern to the decadal to centennial timescales of the peat record would imply that lower $\delta^{18}O$ values should be interpreted as being indicative of prevailing cooler and wetter conditions. Intuitively this would imply a strengthening of the westerly storm belt, based on the variables outlined above, capable of determining the isotopic composition of meteoric precipitation. If variation in $\delta^{18}O$ of *Sphagnum* α-cellulose does reflect changes in rainfall composition (Daley *et al*., 2010), then isotopic records could be expected to demonstrate at least a degree of inter-site agreement and coincidence with other regional isotopic records.

Strong visual and statistical correlation between a composite BSW and $\delta^{18}O$ record from Walton Moss led Daley *et al*. (2010) to hypothesise a common climatic driver for the two variables over centennial scales. Figure 8.15 presents the $\delta^{18}O$ records from each site plotted with the respective testate amoebae WTD reconstruction (i.e. prior to regional tuning and stacking and thus maintaining their original chronologies), in an effort to ascertain whether the $\delta^{18}O$ correlates with BSW. No significant relationship is observed between the $\delta^{18}O$ and WTD reconstructions from either site, suggesting that $\delta^{18}O$ of *Sphagnum* α-cellulose does not simply reflect changes in atmospheric circulation, assuming the testate amoebae record is representative of regional climate. Daley *et al.* (2010) suggest that periods of poor correlation could be explained, at least in part, by a sporadic presence of *Sphagnum*, but clearly this explanation cannot be offered here. Of course, sampling resolution complicates any assessment of correlation between the $\delta^{18}O$ records and other palaeoecological data from these sites, as isotopic analyses were conducted at lower resolution owing to financial and time constraints. However, poor correlation is shown in all instances, including those that express more comparable sampling resolutions, supporting the conclusions made above.

Interestingly at Sluggan Moss, a major isotopic excursion towards lower $\delta^{18}O$ values occurs coincidentally with the wet phase in the palaeoecological data recorded at both sites, commencing c. 5400 cal yr BP. Concurrent shifts of this type, in the isotopic and palaeoecological data, would imply a significant variation in the prevailing atmospheric conditions, with lower $\delta^{18}O$ values indicating a strengthening of the westerlies, subsequently resulting in wetter prevailing conditions in the region, and would thus be consistent with the mechanisms proposed by Daley *et al*. (2010). Furthermore, at Fallahogy Bog, a broadly coincident shift in $\delta^{18}O$ data begins c. 5400 cal yr BP, again consistent with palaeoecological data indicating a shift to wetter conditions, but does not possess the same duration at the one recorded at Sluggan Moss. In addition, this excursion consists of just one data point and should therefore be cautiously interpreted.

An isotopic shift of similar magnitude and duration commences approximately 800 years later at Sluggan Moss, but is not associated with an equivalent wet shift in the palaeoecological data. The resolution of the palaeoecological record does not change between these two periods but the isotopic excursion consists, again, of only one data point, albeit with a low associated error. As a result, interpretation of this second excursion, and its relationship with the palaeoecological record, must be treated with caution. In addition, this excursion is not a regional feature, with the record from Fallahogy Bog showing the opposite trend shifting towards higher δ^{18}O values. An additional shift, broadly coherent between both sites but of lower magnitude, occurs c. 3500 cal yr BP. These shifts do not correspond with palaeoecological evidence of wetter conditions, and it is difficult to assign climatic significance to this feature.

Figure 8.15 Comparison of δ^{18}O record (solid) with testate amoebae-inferred WTD reconstructions (dashed) from a) Sluggan Moss and b) Fallahogy Bog.



The second major regional shift to wetter conditions, commencing 3100 cal yr BP, as recorded in the testate amoebae records from both sites, does not appear to result in a significant isotopic shift at Sluggan Moss. The isotope record at Fallahogy Bog does not extend this far. In addition, the two sites' δ^{18}O records demonstrate poor correlation, despite being less than 30km apart, suggesting the absence of a regional record of atmospheric circulation change. Errors associated with the age-depth models could, in some instances, explain poor correlation between two regional records. However, a significant amount of chronological tuning would be required to align these records, and given the presence of a common age marker in Hekla 4 and the relatively high resolution of ^{14}C dates in both sequences makes this an unlikely explanation for this isotopic disparity.

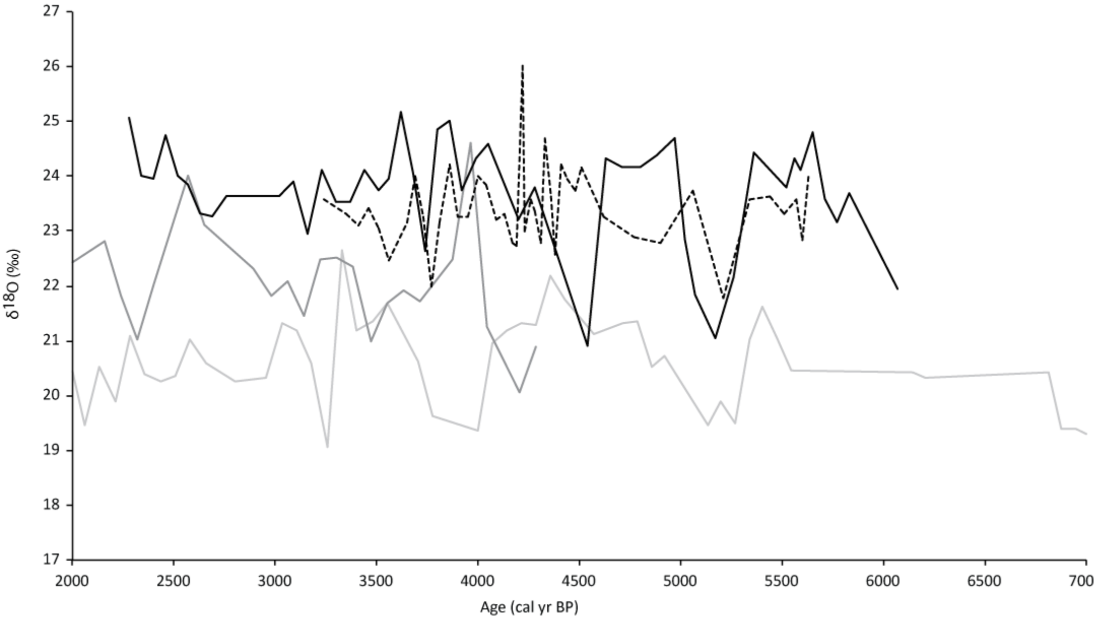
Furthermore, the northern Irish sites bear little resemblance to other sites from the vicinity, in central Ireland and northern England (Fig. 8.16). Again, some isotopic variation between sites would be expected given the slight difference in geographical location, and the subsequent influence of different topographical and atmospheric circulation parameters. However, significant changes in the prevailing climatic influence of the North Atlantic could reasonably be expected to register isotopically at all sites in the region, if atmospheric circulation is the main driver. Apart from some potential correlation at 5450 cal yr BP between Sluggan Moss and Walton Moss, support for a regionally coherent $\delta^{18}O$ signal is weak, calling into question the use of $\delta^{18}O$ from *Sphagnum* α-cellulose as a palaeo-circulation proxy. Age control on the sequences from Walton Moss and Raheenmore is varied. The Irish site has just 4 ^{14}C ages through the period of overlap with Sluggan Moss, the longer sequence of this study, whereas the isotope record from Walton Moss has 12 ^{14}C ages. Whilst such chronological disparity does not make reliable correlation easy, it is clear that a significant amount of chronological tuning would not rectify the lack of correlation, such is the level of $\delta^{18}O$ incoherence. This suggests that isotopic variation is not being directly driven by climatic changes, casting doubt on a simple palaeoclimatic interpretation.

Early peat-based isotopic studies found magnitudes of isotopic change several times larger than expected (Brenninkmeijer *et al*., 1982; Dupont and Brenninkmeijer, 1984; Dupont and Mook, 1987; van Geel and Middeldorp, 1988). As these early studies incorporated bulked peat into the analyses, they were heavily influenced by fluctuations in the proportions of vascular plants present. Daley *et al* (2010) suggest that analyses run on a single genus should yield a more robust estimate of the isotopic composition of precipitation. This study achieved levels of isotopic variation comparable to those found by Daley (2007) and Daley *et al* (2010) in a number of sites in northwest Europe. However, as Figure 8.16 demonstrates the isotopic results from the two sites in this study achieve consistently higher average isotopic values than two sites in central Ireland and northern England (Daley, 2007; Daley *et al*., 2010). Whilst some variation in the average value for $\delta^{18}O$ would be expected, and could be explained by disparity in the distance the meteoric water has travelled over land, prior to precipitation, consistently lower values such as these require additional explanation.

Woodley *et al* (2011) present a novel methodology enabling the simultaneous measurement of stable isotopes of carbon and oxygen using high-temperature pyrolysis, thus reducing the required sample size. This method significantly reduces the amount of sample required for analysis and therefore permits significant time, as well as financial, savings to be made. This method is now widely accepted and, as a result, was adopted in this study. However, previous studies in this field (e.g. Daley, 2007; Daley *et al*., 2010) followed the earlier methodology, which involved simple combustion in place of the high-temperature pyrolysis. Woodley *et al*. (2011) ran a large number of tree-ring α-cellulose samples using both techniques, finding that the results were statistically indistinguishable, after adjustment to correct for the formation of carbon monoxide in the reaction chamber. The results presented here have been corrected according to this protocol but the apparent offset persists. Others have observed this phenomenon (Daley, pers. comm.), but currently no explanation can be offered.

In addition, Daley (2007) and Daley *et al* (2010) isolated *Sphagnum* leaves for stable isotopic analysis, compared to the stems used in this study (cf. Loader *et al*., 2007; Moschen *et al*., 2009; Tillman *et al*., 2010). A statistically significant isotopic offset has been found between stems and branches for both δ¹³C (c. 1.5 – 2 ‰; Loader *et al*., 2007; Moschen *et al*., 2009; Tillman *et al*., 2010) and δ¹⁸O (c. 0.9 ‰; Moschen *et al*., 2009). It would, therefore, be reasonable to expect a significant offset to exist between leaves and stems, potentially explaining some of the offset between sites from different studies seen in Figure 8.16. To the author's knowledge, no comprehensive study has taken place assessing the offset between leaves, alongside stems and branches.

Figure 8.16 Comparison of δ¹⁸O record from Sluggan Moss (solid; this study), Fallahogy Bog (dashed; this study), Raheenmore (dark grey; Daley, 2007) and Walton Moss (light grey; Daley, 2007, Daley *et al*., 2010).



However, as the offset between the sequences in this study, and those produced by Daley (2007), are likely to be the result of methodological issues, it can be reasonably assumed that these effects are uniformly distributed through the record, and so the timing and direction of changes could still be used to compare the records. There appears to be some agreement between the Sluggan Moss and Walton Moss isotopic records during the first major excursion seen at the Irish site, commencing c. 5500 cal yr BP. Whilst the magnitude of the 'event' is far greater at Sluggan Moss (c. 3 ‰) than at Walton Moss (c. 1.5 ‰), the timing and duration are markedly similar. Beyond this, however, comparability is poor between all records, and so any sporadic correlation must be interpreted cautiously.

8.3.2 δ¹³C data

Sphagnum lack stomata and, therefore, cannot regulate carbon uptake. Instead, δ¹³C in *Sphagnum* is controlled principally by environmental variables such as light intensity (Farquhar *et al*., 1989; Ménot and Burns, 2001), temperature (Ménot and Burns, 2001; Jedrysek and Skrzypek, 2005) and moisture (Price *et al*., 1997; Ménot-Combes *et al*., 2004). The effects of temperature and light conditions can be considered negligible on a local scale at a given time, and subsequently variation in *Sphagnum* δ¹³C can reasonably be expected to reflect changing BSW conditions (Rice and Giles, 1994, 1996; Price *et al*., 1997; Ménot and Burns, 2001; Lamentowicz *et al*., 2008; Loisel *et al*., 2009, 2010; van der Knaap *et al*., 2011). Based on this interpretation it could be suggested that *Sphagnum*

δ^{13}C records should reflect, to some extent, hydrological changes recorded at each site, as revealed by other BSW proxies, such as testate amoebae, plant macrofossil and/or humification analyses derived from the same sequences. Depleted (i.e. more negative) δ^{13}C values would be expected during periods of reduced water resistance, caused by a drier bog surface, and enriched (i.e. less negative) values when water resistance increases, as a result of increased BSW; a pattern demonstrated by Loisel *et al*. (2010) in records from Northern Quebec.

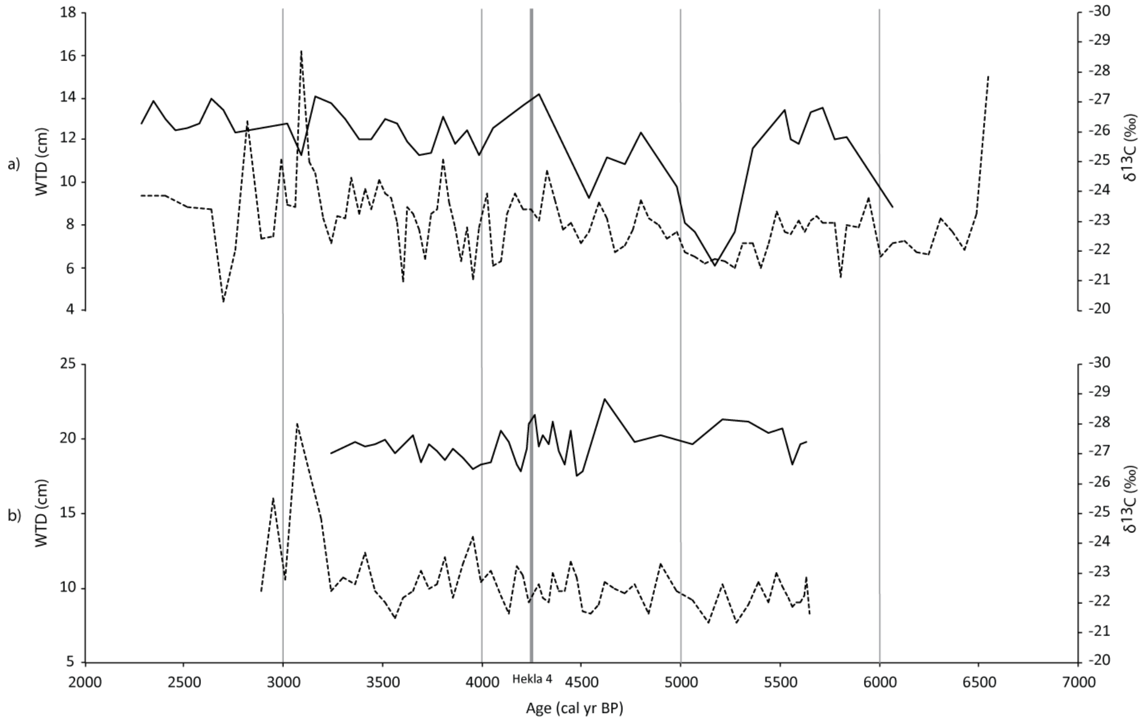
Figures 8.17 presents the δ^{13}C records from each site plotted with the respective testate amoebae WTD reconstruction, in an effort to ascertain whether the δ^{13}C correlates with BSW conditions, as influence by WTD. This relationship was not found here, with the only marked agreement shown between δ^{13}C values and WTD reconstructions during the major wet period commencing c. 5450 cal yr BP at Sluggan Moss (Fig. 8.17a). δ^{13}C data and the WTD reconstruction from Fallahogy Bog express no correlation (Fig. 8.17b).

As δ^{13}C values from both sites express little or no correlation with WTD reconstructions, explanations must be sought for their disparity. Loisel *et al*. (2010) offer a number of potential explanations for periods of disagreement between their WTD reconstructions and δ^{13}C values, which can be discussed for this record. Firstly, as in this study, Loisel *et al*. (2010) compared δ^{13}C values to a single palaeohydrological proxy, testate amoebae. As a result, error ranges of the subsequent WTD reconstructions were presented as possible, if only partial, explanation for disagreement between the two records. Given the level of disagreement between the δ^{13}C and WTD reconstructions, this potential explanation seems unlikely.

A number of other environmental variables have been proposed to account for a lack of correlation between δ^{13}C values and the BSW record. Hygrophilous *Sphagnum* species, such as *Sphagnum* section *Cuspidata*, are by definition more sensitive to desiccation than hummock or lawn mosses, as a result of poorer capillary water transport systems and water holding capacity (Titus and Wagner, 1984; Rice, 2000; Gauthier, 2001). It has been demonstrated that desiccation increases discrimination against the heavier isotope (Williams and Flanagan, 1996; Ménot and Burns, 2001) and is more likely to occur in pool or hollow environments (Rydin, 1985).

Loisel *et al*. (2010) imply that depleted δ^{13}C values, recorded by species indicative of pool environments are influenced by a combined effect of temperature and moisture on their metabolic activity. These species may be more susceptible to water table changes and are likely to produce δ^{13}C values of greater range than other species. Indeed, at Sluggan Moss, the three periods of most pronounced variations in the δ^{13}C record occur when *S*. s. *Cuspidatum* is either dominant (c. 5250 cal yr BP), or present (c. 6200 cal yr BP; 4500 cal yr BP). At these times, however, δ^{13}C values become enriched (i.e. less negative), indicating a wetter environment, with no evidence for the desiccation of these species. Whilst evidence of wetter conditions would be expected in the δ^{13}C record during periods of *S*. s. *Cuspidatum* presence, the magnitude of these changes could be explained by an increased propensity of these species for greater isotopic ranges, as suggested by Loisel *et al*. (2010). *S*. s. *Cuspidatum* is not present in the studied sequence at Fallahogy Bog.

Figure 8.17 Comparison of $\delta^{13}C$ record (solid) with testate amoebae-inferred WTD reconstructions (dashed) from a) Sluggan Moss and b) Fallahogy Bog.

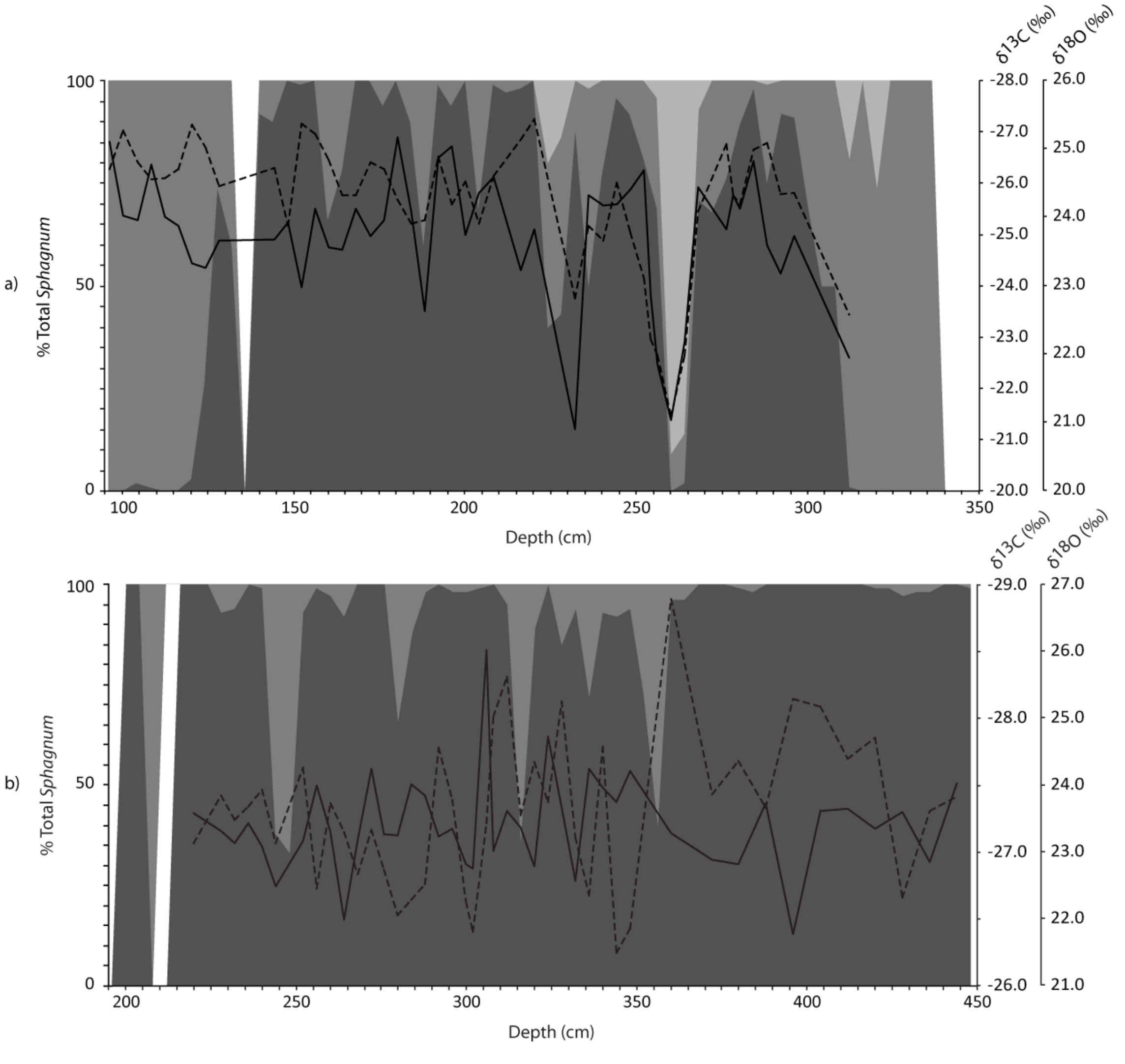


It has also been suggested that the $\delta^{13}C$ record, as recorded in *Sphagnum*, could be affected by an uptake of dissolved inorganic carbon (DIC) by the plants from the surrounding water (Ménot and Burns, 2001). Similarly, it has been demonstrated that submerged *Sphagnum* species may also assimilate carbon from methane (CH_4), via symbiotic relationship with methanotrophic bacteria (Raghoebarsing *et al.*, 2005). Whilst these two processes could significantly interfere with any potential climate signal, as derived from the $\delta^{13}C$ record, the precise bio-mechanisms required for their operation are relatively poorly understood, and no further evaluation of their potential role can be undertaken using the data from this study.

Whilst it could be argued that there is potential for a 'species signal' to exist as a result of these mechanisms, it has been shown that effects of species-specific morphological features on the carbon assimilation and isotopic composition, are far less than that of moisture (Rice, 2000; Loisel *et al.*, 2009). Furthermore, comparison of the isotopic record for both $\delta^{13}C$ and $\delta^{18}O$ with the composition of the *Sphagnum* component at Sluggan Moss (Fig. 8.18a) demonstrates that, for example, low values for both isotopes persist beyond the *S.* s. *Cuspidata* incursion during the wet period c. 5300 cal yr BP. Whilst there may appear to be a weak visual correlation between the isotopes and the presence of *S. austinii*, especially in the first half of the record, the relationship is poorly constrained and subsequently over-interpretation must be avoided. Furthermore, this comparison at Fallahogy Bog reveals no correlation between $\delta^{13}C$ and $\delta^{18}O$ and the composition of the *Sphagnum* component (Fig. 8.18b).

Correlation between $\delta^{13}C$ and $\delta^{18}O$ at Sluggan Moss, together with the palaeoecological record, during the apparent wet phase 5500 – 5000 cal yr BP, is suggestive of a common driver under wetter conditions. This basic relationship would be consistent with the interpretation of $\delta^{18}O$ as a proxy for atmospheric circulation (Daley *et al.*, 2010) and therefore strength of the westerlies, and δ^{13}C as a proxy for BSW, as the two variables are inherently linked. However, why this agreement should only operate under extreme wet conditions is unclear. Fallahogy Bog, as discussed earlier, appears to be a drier bog than Sluggan Moss, which could explain by only poor agreement between the two isotopes exists there. The coincident isotopic excursions witnessed at Sluggan Moss also call into question the affect of any DIC and CH$_4$ uptake on the δ^{13}C record under extreme wet conditions, as neither the δ^{13}C variations are consistent those in the δ^{18}O record, the latter of which should not affected by DIC or CH$_4$.

Figure 8.18 Comparison of δ^{13}C (dashed) and δ^{18}O (solid) with the composition of the *Sphagnum* component (*S. austinii*, dark grey; *S.* s. *Acutifolia*, mid-grey; *S.* s. *Cuspidata*, light grey) from a) Sluggan Moss and b) Fallahogy Bog.

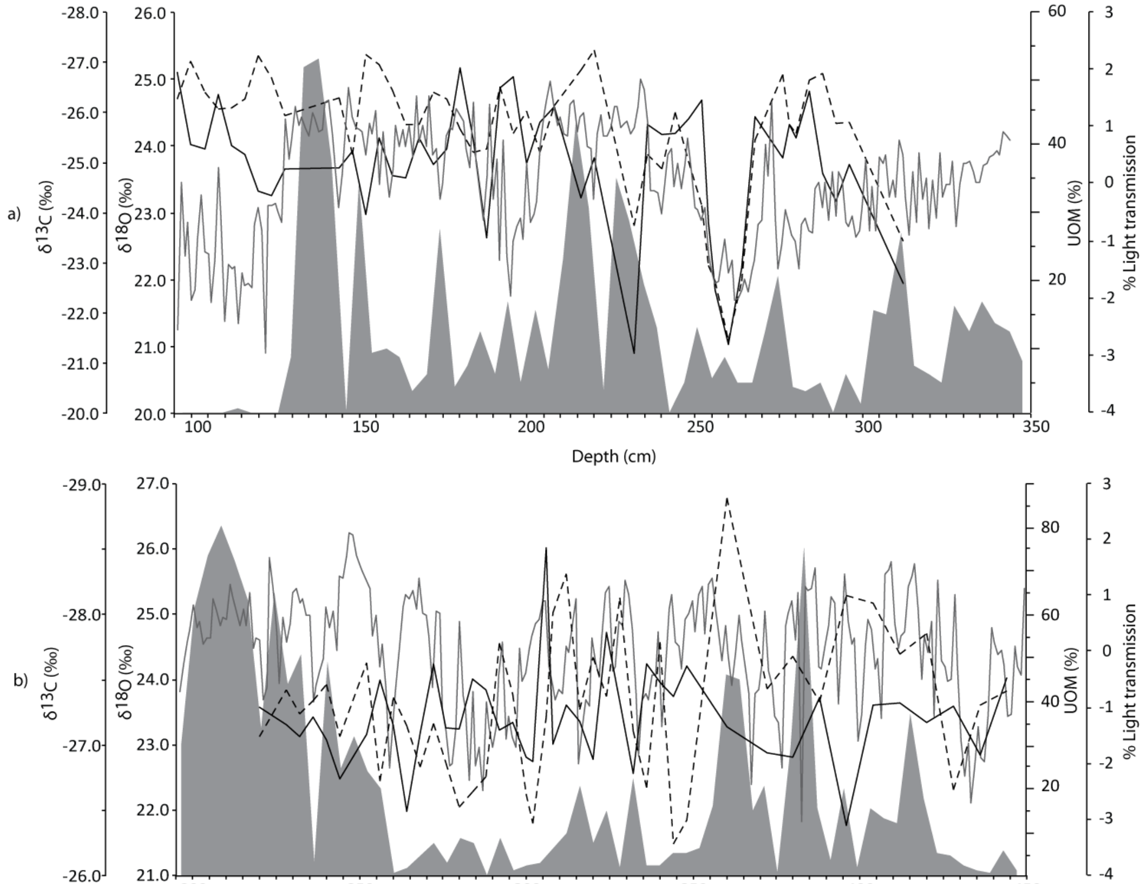


It is worth noting that Loisel *et al.* (2010) followed a slightly different preparation for isotopic analysis of *Sphagnum* in that they did not extract α-cellulose from their samples, choosing instead to run analyses on dried samples. The benefits of α-cellulose have already been discussed and Daley *et al.* (2009, 2010) maintain that this preparation should be accepted as the standard where possible. Loisel *et al.* (2010) describe the early-stage diagenetic processes, which could alter multi-component *Sphagnum* δ^{13}C values, as labile organic compounds such as carbohydrates (i.e.

including cellulose) are preferentially removed over time. As this study isolated α-cellulose for analysis, any down-core trend caused by preferential decomposition of plant sugars was avoided.

In addition, it has been suggested that long-term anaerobic decomposition could potentially modify *Sphagnum* δ^{13}C values (Loisel *et al.*, 2010; Tillman *et al.*, 2010). Loisel *et al.* (2010) compared their δ^{13}C record with UOM, as a proxy for intensified decay, finding that periods of high percentage UOM also exhibited δ^{13}C depleted values, thus implying that a preferential loss of δ^{13}C-enriched compounds through intensive aerobic decomposition has not occurred. Here, Figure 8.19 compares the isotopic record from both sites with variations in decay indices, UOM and humification (% light transmission) data over time, with the latter possibly a better proxy for long-term anaerobic decomposition. These data demonstrate no consistent relationship with either isotope, at either site, suggesting that long-term anaerobic decomposition has made no significant impact on isotopic composition of *Sphagnum* α-cellulose (cf. Tillman *et al.*, 2010).

Figure 8.19 Comparison of δ^{13}C (dashed line) and δ^{18}O (solid line) records with the proportion of UOM (grey histogram) and degree of humification (grey line) from a) Sluggan Moss, and b) Fallahogy Bog.



The final factor worthy of consideration, as suggested by Loisel *et al.* (2010), is species-specific resistance to decay. It has been demonstrated that pool or hollow species are more easily decomposed than their hummock equivalents (Johnson and Damman, 1991). Loisel *et* al. (2010) suggest that this, coupled with their earlier hypothesis that hollow species are likely to become more enriched in $\delta^{13}C$ than hummock species, relatively speaking, could interfere with any isotopic climatic signal recorded in peat sequences.

However, the earlier comparisons between variation in the composition of the *Sphagnum* component and the isotopic composition of *Sphagnum* α-cellulose demonstrate that no significant or consistent relationship can be gleaned from these data, supporting previous assertions that any species-specific effects that might exist are negligible (Rice, 2000; Loisel *et al*., 2009). Although, these studies also suggest that for $\delta^{13}C$ moisture is the chief driver, a proposal for which these data can provide little support.

8.3.3 Summary of isotopic data

Overall, the isotopic data generated in this study cannot support or reject either hypothesis for the principal drivers of each respective stable isotope. If $\delta^{18}O$ were driven primarily by changes in atmospheric circulation, as proposed by Daley *et al*. (2010), then it would be reasonable to expect two records in close proximity, such as Fallahogy Bog and Sluggan Moss, to possess distinct similarities. This is clearly not the case. In addition, these $\delta^{18}O$ data also show no significant relationship to other $\delta^{18}O$ records in the region. Similarly, if $\delta^{13}C$ were primarily driven by changes in BSW, as proposed by Loisel *et al*. (2010), then it would be reasonable to expect each isotopic record to possess distinct similarities with their corresponding site-specific BSW/WTD records. This is not a pattern observed at either site in this study. These data would therefore suggest that other factors are of influence, either sporadically or consistently, in the composition of *Sphagnum* isotope composition, beyond those proposed by Daley *et al*. (2010) and Loisel *et al*. (2010).

At Sluggan Moss, an apparently a wetter site than Fallahogy Bog, agreement exists between both the isotopes during a period of extreme wetness, commencing c. 5450 cal yr BP. This mutual excursion persists through changes in *Sphagnum* species and, therefore, cannot be attributed solely to species-specific effects. This relatively isolated agreement implies that different mechanisms are dominant in their influence on the isotopic record at different times and under varying environmental conditions, calling into question the efficacy of the technique as a palaeoclimatic proxy, at least under the constraints of our current understanding. Finally, no significant agreement exists at Fallahogy Bog between either isotope of the WTD reconstruction, further complicating interpretation of the isotopic signal.

van der Knaap *et al.* (2011) produced similarly unconvincing results in a peat-based, multi-proxy study of the last 1000 years from a site in the subalpine Swiss Alps. Periodic correlations were found between *Sphagnum*-derived $\delta^{13}C$ records and testate amoebae-inferred WTD reconstructions, and between the $\delta^{13}C$ and $\delta^{18}O$ record. Whilst the former relationship was presented as evidence for a potential link between $\delta^{13}C$ and WTD, the overall conclusions reached were that more research was needed to clarify the mechanism by which carbon is incorporated into the *Sphagnum* plant. It is also worth noting that α-cellulose was not extracted here, due to limitations of available sample material. This limitation also meant that a maximum of two isotopic measurements were made per sample, with many providing enough material for just one measurement. In addition, an overly complex sampling regime was adopted, involving a convoluted subdivision of different types of stems and branches, according to colour and perceived structure, coupled with various techniques to remove offset and combine samples into a time-series. Together the issues presented by the van der Knaap *et al.* (2011) study epitomise the problems associated with the technique. Lamentowicz *et al.* (2008) reached similar conclusions to van der Knaap *et al.* (2011) and Loisel *et al.* (2010) within regard to the relationship between $\delta^{13}C$ and BSW. However, this study adopted a third variation in sampling regime, where *Sphagnum* stems mainly of a single species were isolated, dried and the bulk organic matter of the material was subject to stable isotopic analysis, without the extraction of α-cellulose.

A lack of consistency between methodological approaches with regard to the subject of the analyses (i.e. *Sphagnum* α-cellulose vs. whole *Sphagnum* stem material) and methodology (i.e. high temperature pyrolysis vs. low temperature pyrolysis/combustion) severely hampers the interpretation of findings and any subsequent comparison between studies. As shown by the comparison of this study's isotopic records with the other proxy data, as well as with other regional peat-based isotopic records, a comprehensive evaluation of the laboratory techniques involved and the development of refined and standardised *Sphagnum*-specific methodology is urgently required. In addition, it is generally recognised that a full understanding of the biomechanics involved in the generation of the isotopic composition in *Sphagnum* has not yet been developed, and subsequently requires further investigation. At present, the uncertainties surrounding the processes involved in deriving and driving $\delta^{18}O$ and $\delta^{13}C$ data in *Sphagnum* records severely reduces their potential for robust palaeoclimatic reconstruction.

Chapter 9 – Discussion II

9.1 Introduction

The second and third research questions of this study are concerned with: i) whether a 4.2 kyr event can coherently be characterised in the region of Great Britain and Ireland; and, ii) whether peat-based stable isotopic analysis can contribute to our understanding of the nature and/or cause of this event. This chapter will consider the isotopic and palaeoecological proxy data generated by this study, building on the methodological discussion of Chapter 8, before contextualising it alongside key records from the Great Britain and Ireland, and North Atlantic realm.

Hypothesised causes of the global 4.2 kyr event are centred around changes in the earth's ocean-atmosphere circulation system (see Chapter 2). A complex interaction between a number of sub-systems (e.g. NAO, westerlies etc.), many of which are strongly linked to the North Atlantic, appear to be the most plausible driving causes of the event. Based on evidence from previous studies (e.g. Hughes *et al*., 2000; Barber *et al*., 2003) it would be reasonable to expect a period of wetter conditions encompassing the period 4200 cal yr BP in Great Britain and Ireland. Whilst plant macrofossil records from the region (e.g. Hughes *et al*., 2000; Barber *et al*., 2003) have demonstrated preliminary evidence for an event around this time, no attempt has yet been made to examine the period as comprehensively as this study, in terms of sampling resolution, number of proxy techniques and chronological resolution.

9.2 A 4.2 kyr event in the peatland data of Great Britain and Ireland?

The importance of obtaining multiple records within a region is emphasised by Charman *et al*. (2006), as there are inherent uncertainties associated with any single proxy record from any given site. However, as Charman *et al*. (2006) also acknowledge, problems arise when attempting to interpret and reconcile multiple records from a single region, with many of these issues are associated with the chronological differences between records. To mitigate against both these issues, each proxy record was firstly assessed for evidence of regional coherence, and therefore a likely climate signal. As the only proxy that expressed demonstrable regional coherence, the testate amoebae records were tuned and stacked (cf. Charman *et al*., 2006; Blundell *et al*., 2008; Swindles *et al*., 2010a) to produce a regional, mid-Holocene palaeoclimate curve.

Inspection of the stacked WTD record from Sluggan Moss and Fallahogy Bog (Fig. 9.1a) reveals that, whilst indicating some moderate increases in wet conditions, the period c. 4250 – 3750 cal yr BP, presents no convincing evidence to support previous assertions by Anderson *et al*. (1998), Hughes *et al*. (2000) and Barber *et al*. (2003) that this was a significant period of prevailing wet conditions in Great Britain and Ireland. WTDs do become higher at points within this period, but these are not distinguishable from variability elsewhere in the studied sequences. Other stacked testate amoebae data from elsewhere in northern Ireland (Fig. 9.1b; Swindles *et al*., 2010a) and also in northern Great Britain (Fig. 9.1c; Charman *et al*., 2006), when potential ^{14}C dating errors are taken into account, demonstrate a larger magnitude of change during this period, with water tables rising notably through the period associated with the 4.2 kyr event. Regrettably, neither of these records persist back beyond c. 4500 cal yr BP, and cannot, therefore, be used to contextualise this event within the broader mid-Holocene period. Again, this highlights a paucity of data of this type from the mid-Holocene. It is possible, that the stacked record developed within this study is muted by the co-dominance of *A. flavum* and *D. pulex*. A fourth stacked but untuned palaeoecological record, from Scotland (Langdon and Barber, 2005), is presented in Figure 9.1d. Whilst this record displays some broader correlation with the other three stacked records, agreement during the 4400 – 4000 cal yr BP period is negligible. However, the resolution of the Scottish record is poor and therefore of limited use here.

Figure 9.1 Comparison of stacked water table records from Great Britain and Ireland 2 – 7 ka. All normalised and plotted against standard units. a) Northern Irish stacked (2 records) testate amoebae-inferred WTD reconstruction (this study); b) Northern Irish stacked (2 records) testate amoebae-inferred WTD reconstruction (Swindles *et al.*, 2010a); c) Northern British stacked (12 records) testate amoebae-inferred WTD reconstruction (Charman *et al.*, 2006); d) Scottish stacked (7 records) multi-proxy palaeohydrological reconstruction (Langdon and Barber, 2005). Light grey shading shows, for reference, the period 4250 – 3750 cal yr BP, following preliminary evidence of a shifts to cooler/wetter conditions recorded in Great Britain and Ireland by Anderson *et al.* (1998) Hughes *et al.* (2000) and Barber *et al.* (2003). Dark Grey bands refer to periods 5500 – 5000 and 3200 – 2300 cal yr BP discussed in Section 9.5.

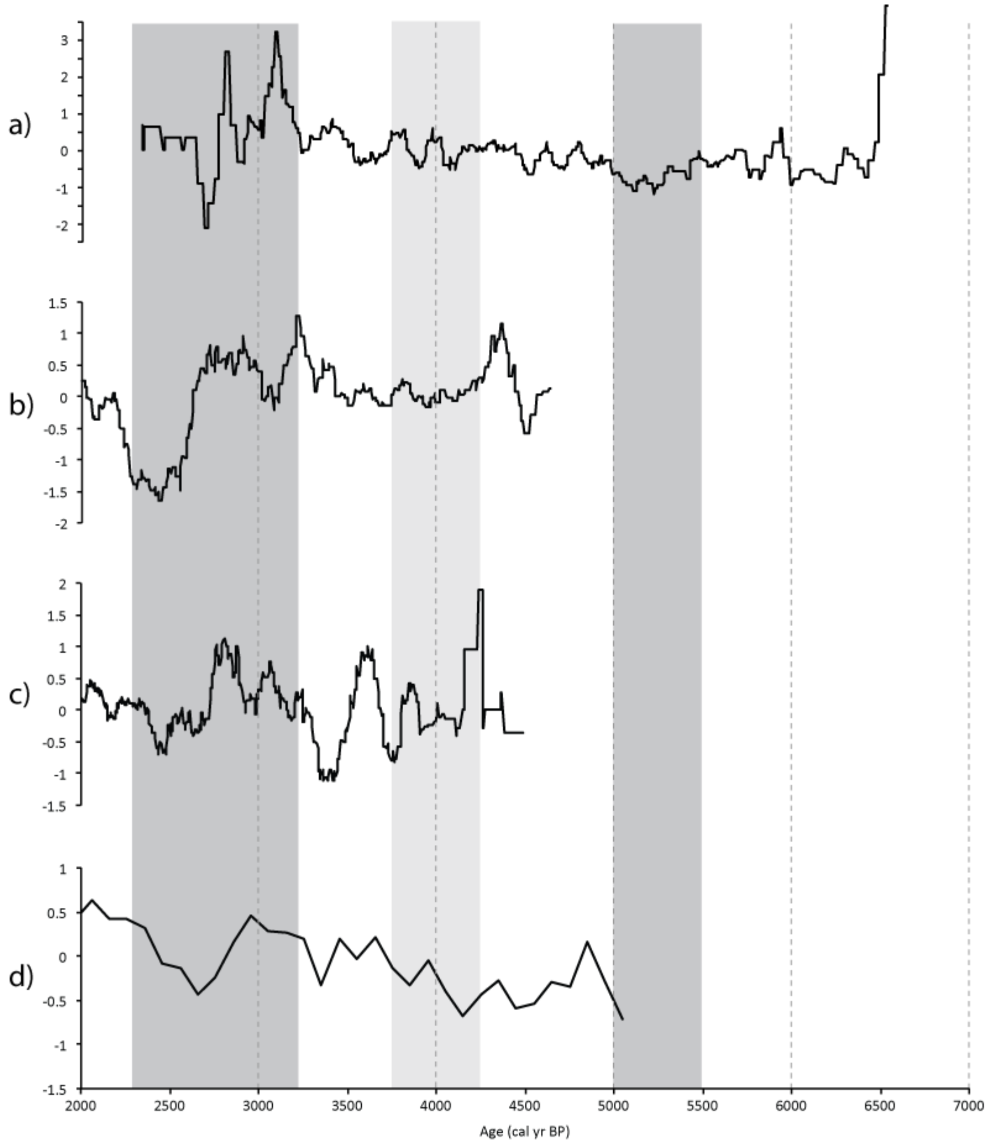


Figure 9.2 Comparison of this study's stacked water table record with peat-based palaeoecological studies from Great Britain and Ireland. All normalised and plotted against standard units. a) Northern Irish stacked (2 records) testate amoebae-inferred WTD reconstruction (this study); b) Walton Moss, northern England, plant macrofossil DCA BSW reconstruction (Hughes *et al*., 2000); c) Butterburn Flow, northern England, testate amoebae-inferred WTD reconstruction (Mauquoy *et al*., 2008); d) Abbeyknockmoy Bog, central Ireland, plant macrofossil DHI BSW reconstruction (Barber *et al*., 2003); e) Mongan Bog, central Ireland, plant macrofossil DHI BSW reconstruction (Barber *et al*., 2003); f) Bolton Fell Moss, northern England, plant macrofossil DHI BSW reconstruction (Barber *et al*., 2003). N.B. Records d), e) and f) were digitised from hard copies as original data were unavailable (Barber, pers. comm.). Shading as Figure 9.1.

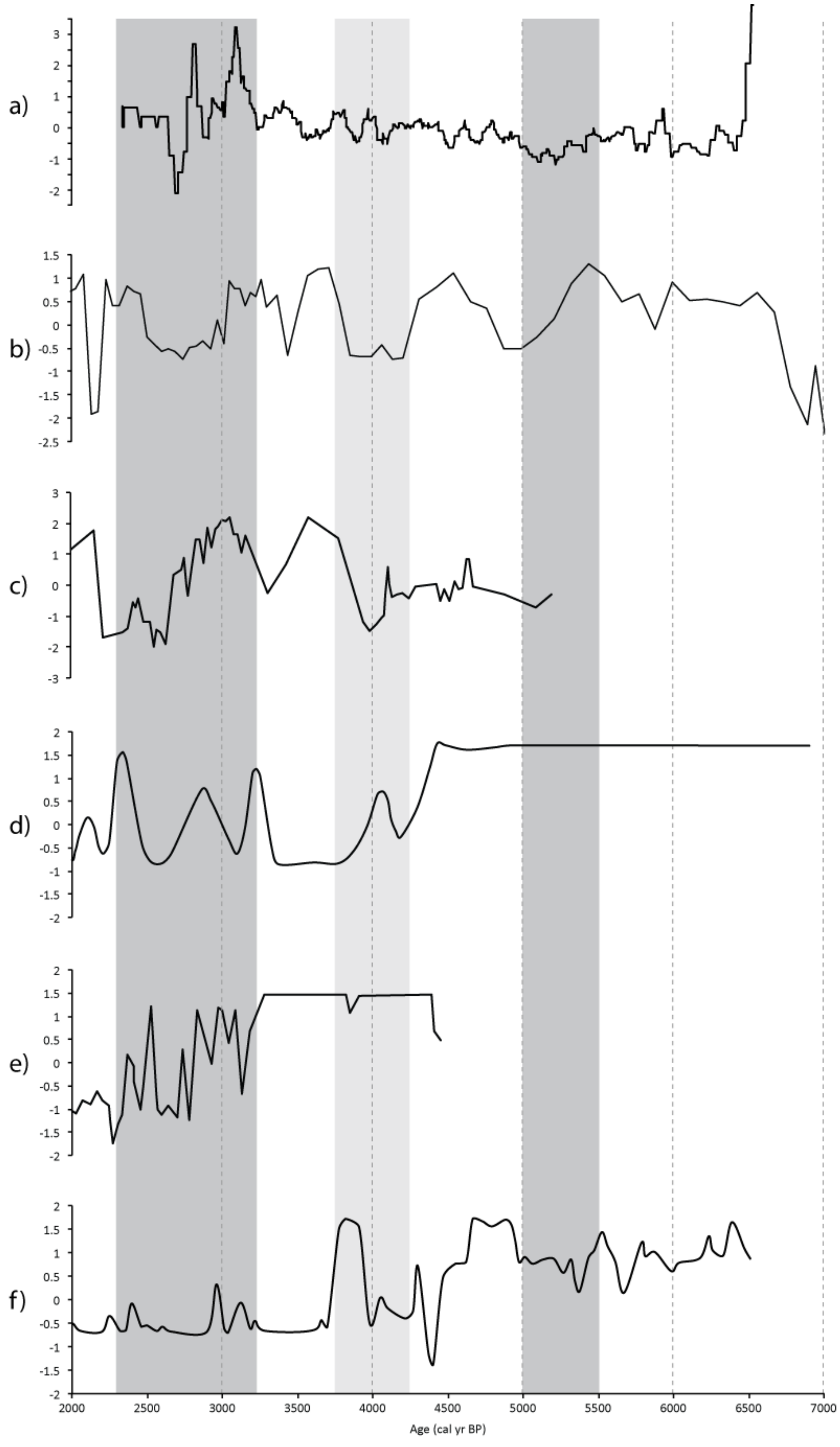


Figure 9.2 presents a number of well-cited, plant macrofossil records from Great Britain and Ireland, in an effort to establish regional coherence. Two sites in northern England, Walton Moss (Fig 9.2b; Hughes *et al*., 2000) and a site in central Ireland, Abbeyknockmoy Bog (Fig. 9.2d; Barber *et al*., 2003) show clear agreement during the 4.2 kyr event period, with coincident shifts towards, or a prolonging of, wetter conditions, within reasonable dating uncertainties. An additional record from central Ireland, Mongan Bog (Fig. 9.2e; Barber *et al*., 2003), displays poor agreement but, again, resolution in this record is poor. Finally, two additional sites in northern England, Butterburn Flow (Fig 9.2c; Mauquoy *et al*., 2008), and Bolton Fell Moss (Fig. 9.2f, Barber *et al*., 2003), demonstrate more signal complexity during the period. However, sampling and chronological strategies at both these sites are questionable. Sampling resolution at Butterburn Flow is extremely varied, and whilst dating resolution is good at Bolton Fell Moss during the 2000 – 7000 cal yr BP period presented here, all of the 9 ^{14}C dates were calculated from bulked samples, which included material from up to 8 vertical cm. The chronological issues associated with this approach are discussed at length in Chapter 6. As a result, the inconsistencies between these records and others in the region could potentially be attributed to errors associated with the dating strategy. It is worth noting that other, unpublished palaeoecological datasets exist from Ireland (e.g. Daley, 2007; Amesbury, 2008) but are not of sufficient resolution during this study's focus to warrant inclusion here.

However, based on comparisons of the data generated in this study with the additional records presented in Figures 9.1 and 9.2, it can be suggested that it is difficult to identify a coherent regional climatic event occurring in Great Britain and Ireland during the period associated with the global '4.2 kyr event' described in Chapter 2. Whilst some records appear to demonstrate a coherent shift during this period, suggestive of a common climatic driver operating in the North Atlantic region during the period, most records do not. Records that are suggestive of a 4.2 kyr event signal in the region (e.g. Hughes *et al*., 2000; Barber *et al*., 2003) are based solely on plant macrofossil records, which this study has shown do not always provide a strong climatic signal. The stacked WTD record developed in this study is one of the records that does not exhibit this trend, despite it being a representation of regional climate change (see Section 8.2.3.3). Figures 9.1 and 9.2 also serve to highlight further the disparity between the potential and the reality of long-core peat based palaeoclimate studies in Ireland, in terms of mid-Holocene, multi-proxy palaeoclimate studies (see Section 4.2.2).

9.3 A 4.2 kyr event in the climate records of the North Atlantic?

The suggestion that Ireland be used as a 'testing ground' for techniques examining the nature of climatic change in the North Atlantic region (Blundell *et al*., 2008) holds true within the rationale of this study's regional focus (see Section 4.2.2). A novel technique for the stable isotopic analysis ($\delta^{18}O$) of *Sphagnum* α-cellulose has been presented by Daley *et al.* (2010) as a method capable of tracking changes in the isotopic composition of meteoric precipitation, and therefore a proxy for atmosphere-oceanic circulation variation. In addition, $\delta^{13}C$ analysis has been presented as an additional proxy for BSW (Loisel *et al*., 2010), alongside the 'traditional' suite of plant macrofossil, testate amoebae and humification analysis.

The isotopic data developed during the course of this study are presented in Figure 9.3b-c for evaluation alongside this study's stacked water table record (Fig. 9.3a) and other key records from the Ireland (Fig. 9.3d-e) and, later, the North Atlantic region (Fig. 9.4). Discussion will focus initially on the period c. 4250 – 3750 cal yr BP, before broader correlations during other periods of significant climatic change are discussed in Section 9.5.

As previously discussed, neither set of isotopic data generated here appear to possess a coherent signal either as proxies for atmospheric circulation or BSW. As a result of this ambiguity, no event associated with a global '4.2 kyr event' can be identified from these data. In addition, comparison of the $\delta^{13}C$ and $\delta^{18}O$ data with the stacked testate amoebae water table reconstruction from northern Ireland further supports the lack of coherent climatic signal. However, relatively few isotope records are available for comparison from the region, particularly from Ireland. Figure 8.16 presented two additional peat-based isotopic records from central Ireland and northern Great Britain, finding no significant correlation (although see Section 8.3.1). To fully assess the use of $\delta^{18}O$ from *Sphagnum* cellulose as a proxy for past atmospheric-oceanic interactions, it would be useful to compare with lacustrine and speleothem isotopic records, to test the proposed mechanisms involved.

A speleothem-based $\delta^{18}O$ record from southwest Ireland is one of few available isotopic records from the region (McDermott *et al*., 2001). The record, from Crag Cave (CC3), is presented in Figure 9.3d and demonstrates little coherence with either the stacked palaeoecological or $\delta^{18}O$ data from Sluggan Moss and Fallahogy Bog. Interestingly, however, there is some agreement between the $\delta^{18}O$ data from Sluggan Moss and CC3 during the period c. 5300 – 4300 cal yr BP. This agreement occurs during what are by far the largest isotopic excursions in the peat record, and also broadly coincident with what is arguably the wettest period recorded by the palaeoecological data here, 5500 – 5000 cal yr BP. However, significant coherence does not exist anywhere else between the Sluggan Moss record. Also, little coherence can be said to exist between the Fallahogy Bog and Crag Cave $\delta^{18}O$ data, beyond the single sample excursion c. 5400 cal yr BP described previously (see Section 8.3.1). The data from CC3 were adjusted by Fairchild *et al*. (2006) to correct for analytical error c. 8400 – 8200 cal yr BP. Such corrections were not necessary on the section of record presented here. It is worth mentioning that the resolution on the peatland-derived $\delta^{18}O$ record is far lower than that of the speleothem, which could complicate interpretation. Even aside from this, the apparently intermittent relationship between the two $\delta^{18}O$ records is difficult to explain, especially given the lack of a relationship with the equivalent record from Fallahogy Bog. As a result, it would be unwise to place too much importance in this transitory correlation.

Interpretation of the northern Irish bog oak record, as presented in Figure 9.3e in terms of their changing number over time, differs between studies (Turney *et al*., 2005; Charman, 2010; Swindles and Plunkett, 2010). Ultimately, the ability of oak trees to encroach on to the bog surface is governed by changes in past hydrological conditions. During dry periods, water tables are lowered and the bog surface can be populated by trees (Pilcher *et al*., 1996). Turney *et al*. (2005) inspected the data, and describe cyclic pattern, which they interpret as evidence for changes in moisture delivery to Ireland over time. Subsequently, peaks in bog oak populations are interpreted as being representative of 'dry' periods, within a changing the precipitation-evaporation balance. In addition, average age analysis (cf. Leuschner *et al*., 2002) to examine whether the observed 'cycles' had any effect on oak population dynamics, found that troughs in tree population coincided with peaks in the average age of the concurrent populations. Turney *et al*. (2005) interpret this as being indicative of recruitment failure, whereby saplings were prevented from establishing themselves under pervasive wetter conditions, allowing older members of the population to persist, before dying themselves, thus explaining the subsequent marked drop in mean age (Leuschner *et al*., 2002). Turney *et al*. (2005) attempted to qualify this hypothesis further, through comparison of the bog oak time-series with a BSW reconstruction from southeast Scotland (Langdon *et al*., 2003). It was suggested from this that higher water tables in the Scottish record corresponded well with peaks in tree population, and lower water tables with troughs.

However, records of BSW from Great Britain and Ireland, more extensive than that used by Turney *et al*. (2005), were presented by Charman (2010) and demonstrate that although coherent patterns of multi-centennial moisture variability occur across the region, these do not coincide with troughs in bog oak population. The dating precision of these records is in the region of ± 50 years (2σ) and subsequently chronological uncertainty can reasonably be discounted as an explanation. Alternatively, it is suggested by Charman (2010) that the interpretation of the bog oak record should be adjusted. Charman (2010) argues that it is not the peaks and troughs in the tree population records that should be interpreted as indicators of dry and wet conditions, respectively, but rather the rising and falling limbs on the curve. For example, during dry conditions, oak trees could be expected to encroach on to the bog, creating a rising limb in the population data. Likewise, periods of increasing wetness should coincide with periods of bog oak population decline, and the consequently a falling limb in the data. Furthermore, increases in mean age should be associated with declining populations, as recruitment slows or even stops. When Charman (2010) reassigns wet phases based on this new interpretation, the bog oak and BSW record demonstrate a greater level of agreement.

Like both sites in this study, many of the bogs that have produced trees for the northern Irish dendrochronology are located in the Lower Bann and Main river valleys (Pilcher *et al*., 1996). In fact, Sluggan Moss has contributed greatly to the bog oak and pine records (Brown *et al*., 1986; Pilcher *et al*., 1995b) and subsequently presents an excellent opportunity to directly test, at least in part, the relationship between BSW and the tree-ring population. However, simple visual comparison between the records struggles to find any consistent correlation between the bog oak record (Figs. 9.3e) and the stacked WTD record (Fig. 9.3a). As a result, there do not appear to be declines in bog oak population coincident with rises in WTD, thus not providing further support for Charman's (2010) interpretation of the tree record. However, the chronological uncertainty associated with the radiocarbon dating technique employed in peat-based studies would make it very difficult to say with any degree of certainty that any WTD decline occurred entirely coincidently with a decline in bog oak population, with troughs occurring at the same time (cf. Turney *et al*., 2005); or whether it preceded it, with bog oak population declining (i.e. the falling limb or the trough) in line with increasing wetness (cf. Charman*,* 2010). In any event, the sporadic nature of this potential correlation must be interpreted with caution. In addition, Swindles and Plunkett (2010) plotted a compiled late-Holocene northern Irish water table reconstruction, derived from the testate amoebae records of two sites (Swindles *et al*., 2010a), against the bog oak data, demonstrating that no consistent pattern exists between the two. Swindles and Plunkett (2010) do not comment on Charman's (2010) interpretation, but instead suggest that further work must be done to fully understand the palaeoecology of the bog oaks before they can be used as reliable palaeoenvironmental proxies. Evidence presented here supports this suggestion and, subsequently, given the degree of uncertainty associated with the records, they should not be used to infer the timing and nature of any climatic event associated with the global '4.2 kyr event'.

Figure 9.4 goes on to present a number of key records from the North Atlantic realm, to be considered alongside the isotopic and stacked WTD records present in Figure 9.3a-c. Figure 9.4a shows the percentage of hematite-stained grains found in a series of four stacked marine records from the North Atlantic (Bond *et al*., 2001). This record has been demonstrated to possess a c.900 and c.500 year cycle, often coupled as a quasi-periodic '1500 year' periodicity, reflecting the southerly migration of cold, ice-bearing waters from the northern North Atlantic. It has been suggested that the 4.2 kyr event corresponds well with IRD event 3, and comparison between this study's stacked testate amoebae water table reconstruction suggests little agreement between the period of change c. 4250 – 3750 cal yr BP and the latter half of IRD event 3.

Figure 9.4b presents oxygen isotope measurements from the planktonic foraminiferal species *Neogloboquadrina pachyderma* (dex) from two marine cores from the Norwegian continental margin. Correlation with nearby instrumental time series, suggest that these data primarily reflect summer temperature conditions of near surface (c. 50 m) waters, and can therefore be used as a proxy for North Atlantic SST. A downward trend commencing c. 4300 cal yr BP and returning to previous SSTs c 3750 cal yr BP, encompassing a short-lived increased in SST c. 4000 cal yr BP (Sejrup *et al*., 2011), demonstrates a two-step pattern which is similar to that seen in the stacked WTD record at this time. However, as discussed previously, the magnitude of the changes c. 4250 – 3750 cal yr BP in the stacked WTD record are no larger or smaller than the general variability within the studied sequence (see Section 8.4.1), and so any climatic interpretations here must be cautious.

Figure 9.4c presents variation in the concentration of sea salt in the GISP2 ice-core (O'Brien *et al*., 1995), interpreted as a proxy for the latitudinal extent of the polar vortex, which influences the atmospheric loading of aerosols over the ice sheet (Turney *et al*., 2005). In theory, if δ^{18}O of *Sphagnum* α-cellulose reflects changes in atmospheric-oceanic circulation, it should possess some similarity to the GISP2 Na record. However, as previously discussed, this is not the case at either site. There is, however, little agreement between this record and the stacked record presented in Figure 9.2a, c. 4250 – 3750 cal yr BP.

In a discussion of the Irish bog oak record, Swindles and Plunkett (2010) challenge the conclusions reached by Turney *et al*. (2005) regarding the role of solar forcing in North Atlantic climatic change, based on the northern Irish bog oak record. Following additional comparison of the dendrochronological record with proxy archives of solar activity and ice-rafting in the North Atlantic, Turney *et al*. (2005) had suggested that millennial to centennial scale climatic change in the region is not driven by a linear response to changes in solar activity, as suggested elsewhere (e.g. Chambers and Blackford, 2001; Blaauw *et al*., 2004a; Swindles *et al*., 2007a).

Figure 9.4d presents ^{14}C production rate (atoms/cm^2 per second) data, derived by way of the ocean-atmosphere box diffusion model, as applied to the INTCAL98 atmospheric ^{14}C dataset (Pearson *et al*., 1986; Stuiver *et al*., 1998). This is taken as an established proxy for variations in solar activity (Oeschger *et al*., 1975; Stuiver and Braziunas, 1993; Bond *et al*., 2001; Hu *et al*, 2003; Turney *et al*., 2005; Swindles *et al*., 2007a). The debate surrounding the role played by solar forcing in the palaeohydrological record derived from peat bog records is extensive (e.g. Hong *et al*., 2000; Chambers and Blackford, 2001; Blaauw *et al*., 2004a; Turney *et al*., 2005; Plunkett, 2006; Hughes *et al*., 2006; Swindles *et al*., 2007a; Plunkett and Swindles, 2008). However, simple visual comparison of this study's stacked palaeoecological record with this record of solar activity presents is inconclusive, with no consistent relationship apparent, instead expressing only sporadic coherence. As a result, these data cannot contribute to this debate any further.

Finally, inferred Greenland temperatures are presented in Figure 9.4e (Alley, 2000), demonstrating declining temperatures from approximately the beginning of the period c. 4250 cal yr BP and recovering to previous levels by c. 3750 cal yr BP, demonstrating little coherence with stacked WTD record, c. 4300 – 3750 cal yr BP.

Elsewhere in the studied sequence, no explicit relationship can be said to exist between the δ^{18}O data generated as part of this study, and records of wider climatic change in the North Atlantic. As a result, no firm support can be given to the assertions of Daley *et al.* (2010), that the technique be used as a proxy for atmospheric-oceanic circulation change. Again, at Sluggan Moss there is a moderately persistent, but transient agreement between isotopic and palaeoecological records, but this is not sufficient evidence upon which to confirm the presence of a genuine climatic signal. This lack of coherence with other regional records further supports the conclusions of Section 8.3.3 that methodological limitations and lack of mechanistic understanding currently reduce interpretive power of isotopic analyses of *Sphagnum* within palaeoclimatic studies.

Figure 9.3 Palaeoclimatic records from the Ireland, 7000-2000 cal yr BP. a) Northern Irish stacked (2 records) testate amoebae-inferred WTD reconstruction (this study); b) δ^{13}C records from Sluggan Moss (solid) and Fallahogy Bog (dashed) (this study); c) δ^{18}O records from Sluggan Moss (solid) and Fallahogy Bog (dashed) (this study); g) δ^{18}O from speleothem CC3, Crag Cave, southwest Ireland (McDermott *et al*., 2001); h) Northern Irish dendro-dated bog oak population (Turney *et al*., 2005). Shading as Figure 9.1.

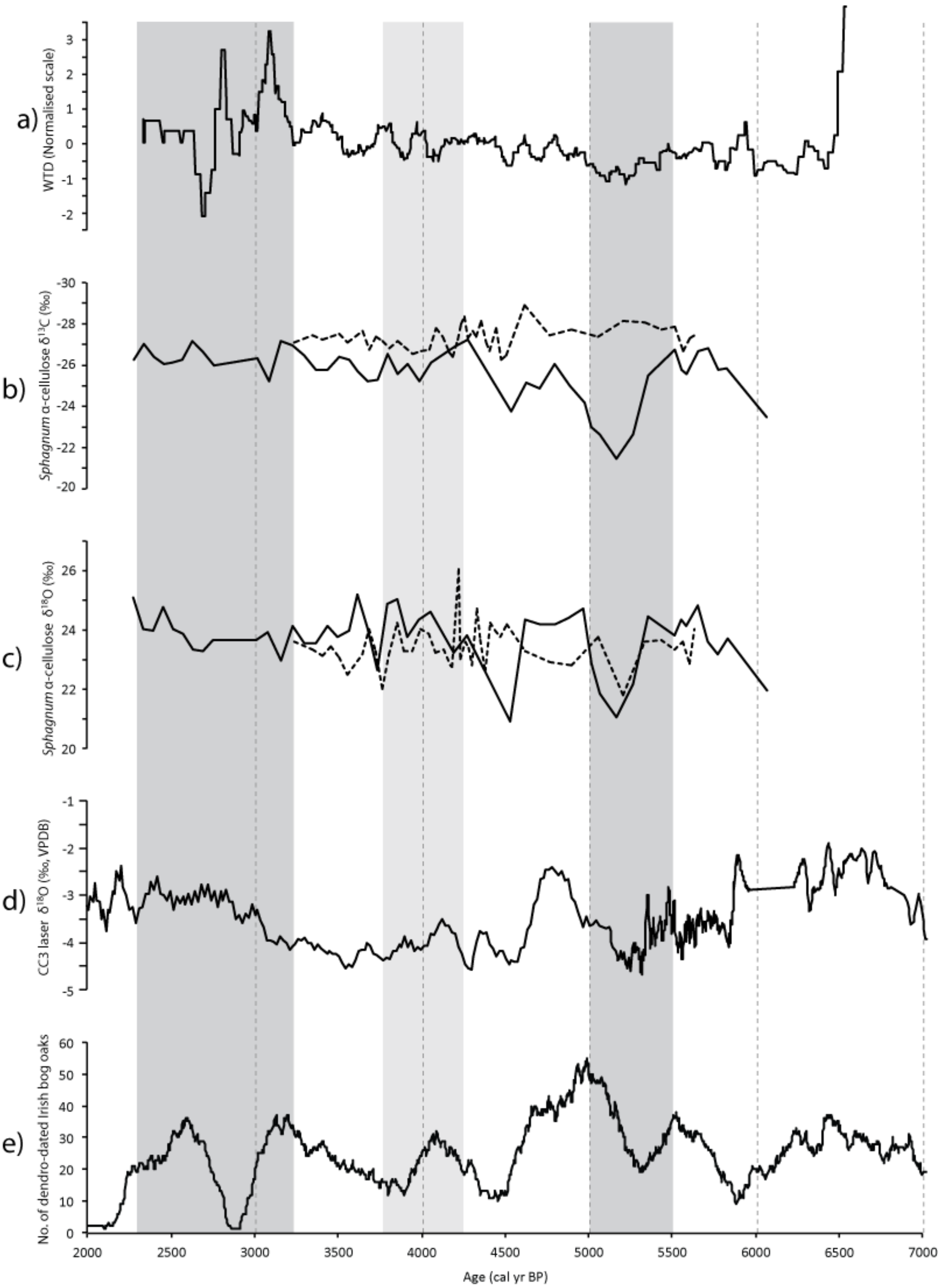
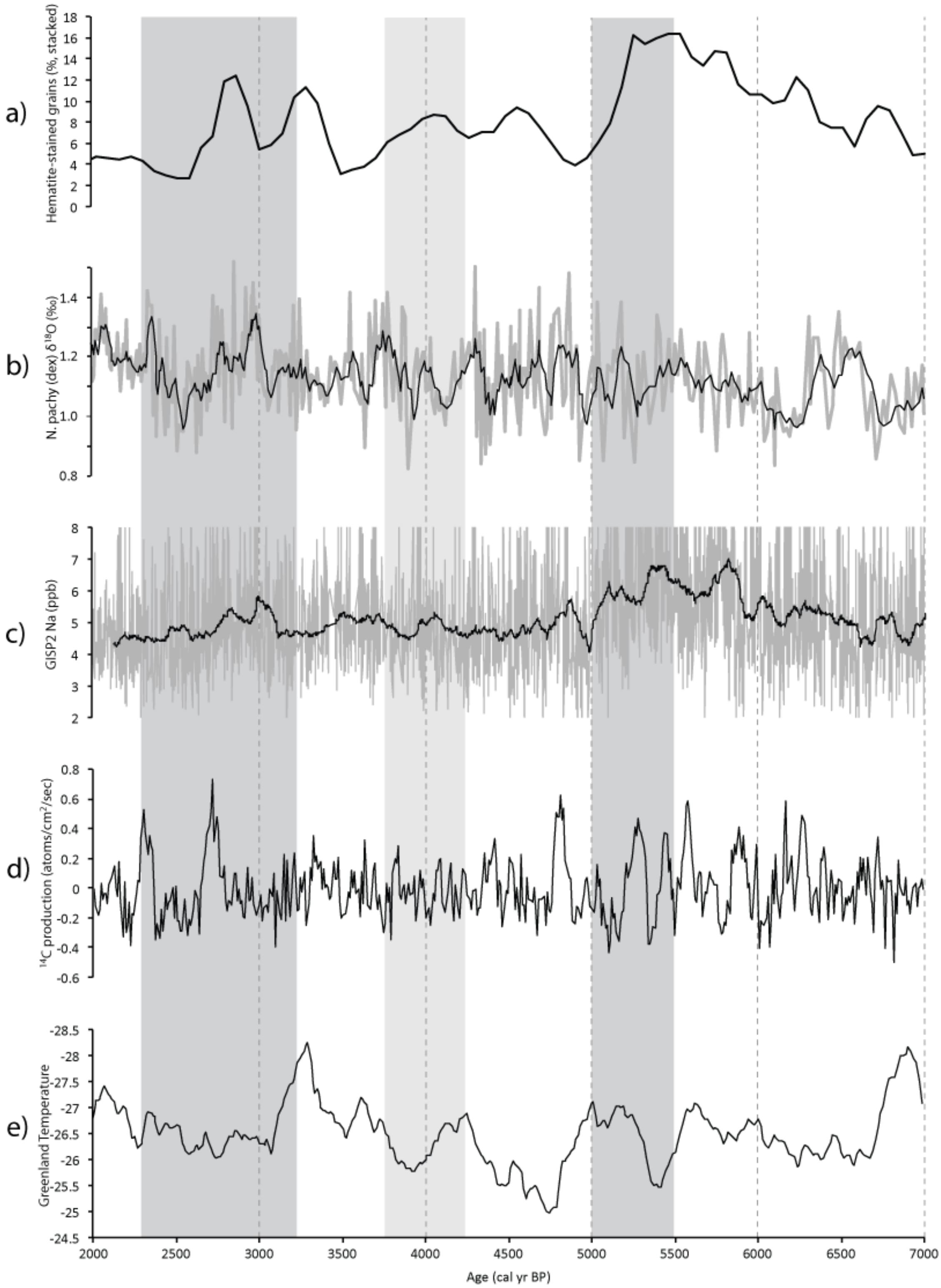


Figure 9.4 Palaeoclimatic records from the North Atlantic, 7000 – 2000 cal yr BP. a) North Atlantic hematite-stained grains record (stacked cores: MC52, V29191, MC21 and GGC22) (Bond *et al*., 2001); b) P1003 *N. pachy*. (dex) sea level adjusted δ^{18}O record from two marine cores from the Norwegian continental margin (Sejrup *et al*., 2011); c) Sodium (Na) content of the GISP2 ice-core (O'Brien *et al*., 1995); d) ^{14}C production rate (Pearson *et al*., 1986; Stuiver *et al*., 1998); e) Temperature over Greenland (GISP2) (Alley, 2000). Shading as Figure 9.1.



9.4 Summary

Palaeoecological evidence from northern Ireland, developed in this study, demonstrate little evidence for the existence of a prolonged phase of significantly wetter and/or colder prevailing climatic conditions associated with the 4.2 kyr event. Some comparable records from the region (e.g. Hughes *et al*., 2000; Barber *et al*., 2003; Charman *et al*., 2006; Swindles *et al*., 2010a) exhibit this phase, but others do not (e.g. Barber *et al*., 2003; Langdon and Barber, 2005; Mauquoy *et al*., 2008). These data therefore demonstrate the difficulties associated with characterising the global 4.2 kyr event, as manifested in Great Britain and Ireland. The former records exhibit some coherence with other records from the North Atlantic region, however, including North Atlantic SST and IRD presence, as well as Na concentrations and temperature reconstructions from the Greenland ice cores. As outlined in the discussion above, this suggests a period of oceanic and atmospheric change in the region, which manifested itself in the certain peatland records of Great Britain and Ireland as a period of cooler and/or wetter conditions. However, this signal fails to register in many peatland records, despite their demonstrating climatic sensitivity (e.g. this study). Where wet phases do exist, they are poorly temporally constrained and do not represent the largest excursions in the mid-Holocene period (see Section 9.5). As a result, the 4.2 kyr event's use as a new Middle-Late Holocene Boundary GSSP (Walker *et al*., 2012; see Section 2.4) may not be justified in the northwest European region.

In addition, isotopic evidence from this study does not present convincing evidence of a '4.2 kyr event' signal, with serious doubts cast over the use of $\delta^{18}O$ and $\delta^{13}C$ from *Sphagnum* α-cellulose as proxies for atmospheric circulation change and BSW. The fact that a regional climatic record is demonstrated in the testate amoebae data (see Section 8.2.3.3), and correlation can be made with other palaeoecological records of North Atlantic atmospheric and oceanic changes, further suggests that the stable isotope evidence presented here do not demonstrate either regional atmospheric circulation change or BSW changes (cf. Daley *et al*., 2010; Loisel *et al*., 2010). There is some agreement between the $\delta^{18}O$ and palaeoecological data from Sluggan Moss c. 5500-5000 cal yr BP suggesting a significant change at this time. However, this transient agreement may be driven by factors other than climate during particularly wet periods, further highlighting the need for further study into the biomechanics involved in the generation of the *Sphagnum* α-cellulose isotopic signal.

9.5 Other periods of abrupt change?

The above palaeoecological data, developed within this study, have shown that the initial temporal focus of this study, the period associated with the 4.2 kyr event, may not yield the largest mid-Holocene climatic change in Great Britain and Ireland. Rather, these data provide evidence for the existence of two additional periods of climatic change during the mid-Holocene: a significant shift to wetter and/or colder conditions c. 5500 – 5000 cal yr BP, and a period of complex climatic change 3250 – 2300 cal yr BP, marked on figures in this chapter by dark grey bands.

9.5.1 3250 – 2300 cal yr BP

The most recent of these two periods sees a sequence of two successive shifts from dry to wet conditions, although characterised by an overall downward shift towards wetter conditions c. 2800 cal yr BP, preceded by a notably drier period. This ‘double-dip’ pattern can be clearly seen in all three of the higher-resolution stacked records from northern Ireland (this study; Swindles *et al.*, 2010a) and northern Great Britain (Charman *et al.*, 2006) (Fig. 9.1). Disparity in the timing of these changes may be accounted for by the error margins of their ^{14}C chronologies, and the relatively small number of records used in the northern Irish studies.

The initial shift to drier conditions c. 3250 cal yr BP is particularly prominent in the testate amoebae records from Sluggan Moss and Fallahogy Bog, representing the largest magnitudinal shift in either record. The shift, whilst not recorded in either site’s plant macrofossil or humification data, possesses such a degree of chronological coherence between sequences that it can be confidently attributed to a significant climatic change. However, Swindles *et al.* (2010a) found evidence for a number of widespread summer ‘drought’ phases in two late-Holocene peat records from northern Ireland, the earliest of which was dated to 1150 – 800 BC (i.e. 3100 – 2750 cal yr BP). This may, on first inspection, appear to contradict the above assertion of regional consistency. However, closer consideration of the dating strategy employed by Swindles *et al.* (2010a) reveals the chronologies possess only 2 ^{14}C ages in the c. 1000 cal yrs either side of this ‘event’, compared to 4 and 5 at this study’s Sluggan Moss and Fallahogy Bog, respectively. Whilst more dates cannot always be taken as being representative of a more robust chronology, it was felt that the structure of the dating strategy within this project provided a more comprehensive chronological reconstruction of this time period (see Chapter 6). In addition, as is the case with both sites in this study, the drought phases identified by Swindles *et al.* (2010a) were followed immediately by significant shifts back towards wetter conditions. The similarity in this pattern, coupled with the potential for dating error associated with the Swindles *et al.* (2010a) sites, leads to the reasonable conclusion that the events are, in fact, the same, with the perfectly reasonable assumption made that chronological error can account for such disparity in timing. Subsequently, this temporal correlation indicates the presences of a significant climatic change around this time, and further confirms the climatic nature of the regional signal produced by the testate amoebae data in this study.

The succeeding wet shift, commencing c. 3100 cal yr BP persists until c. 2600 cal yr BP, but is punctuated by a short-lived dry period centring c. 2800 cal yr BP. When likely chronological discrepancies are taken into consideration, this pattern is extremely consistent with many BSW records from Great Britain (e.g. Walton Moss, Hughes *et al.*, 2000, Fig. 9.1b; Butterburn Flow, Mauquoy *et al.*, 2008, Fig. 9.1c) and central Ireland (e.g. Abbeyknockmoy, Barber *et al.* 2003, Fig. 9.1d; Mongan Bog, Barber *et al.*, 2003; Fig. 9.1e), as well as stacked records from northern Great Britain (Charman *et al.*, 2006) and Scotland (Langdon and Barber, 2005), despite the latter’s lower resolution. If the potential chronological issues regarding the Swindles *et al.* (2010a) stacked record, outlined above, are extended through this period, then good agreement also exists here. At many of these sites, the wet phase persists until c. 2300 cal yr BP. Within this study’s stacked record (representing only Sluggan Moss data by this point), however, water tables appear to return to levels comparable with the sequence mean c. 2600 cal yr BP, although, these levels are still significantly below the preceding dry period. In addition, there is a possibility for the climatic signal to be muted slightly by the co-dominance of the two testate species, as previously discussed.

Overall, this collective body of evidence further reinforces previous assertions, made elsewhere, that this period was one of one of major climatic change in both hemispheres (e.g. van Geel *et al.*, 1996, 1998, 2004; Speranza *et al.*, 2000, 2002; Blaauw *et al.*, 2004a; Mauquoy *et al.*, 2004b; Chambers *et al.*, 2007; Swindles *et al.*, 2007a; Tipping *et al.*, 2008). Extensive reviews of the so-called '2.8 kyr event' can be found in van Geel *et al.* (1996) and Tipping *et al.* (2008). Indeed, further examination of Figure 9.4 reveals that the 3250 – 2300 cal yr BP period highlighted here is characterised by significant variation in a number of other records in the North Atlantic region and appears to be broadly concurrent with IRD event 2 (Fig. 9.4a). In addition, SST in the North Atlantic (Fig. 9.4b) and GISP Na concentrations (Fig. 9.4c, as a proxy for the latitudinal extent of the polar vortex) have been shown to rise and fall in a fashion very similar to the pattern witnessed in the stacked palaeoecological record (Fig. 9.3a) at this time. Greenland estimated temperature also display declines into this period. These records reinforce conclusions based on terrestrial evidence outlined above, that the period represented one of considerable climatic change, focussing on a decline to wetter and/or colder conditions just after 3000 cal yr BP.

Many studies have attributed these changes to fluctuations in solar activity, dominated by a reduction c. 2800 – 2710 cal yr BP (e.g. van Geel *et al.*, 1996, 1998; Speranza *et al.*, 2000, 2002; Blaauw *et al.*, 2004a; Mauquoy *et al.*, 2004b). However, in Ireland, it has been suggested that whilst records exhibit evidence for a similar response to this solar forcing, the manifestation is delayed by approximately 100 years (Swindles *et al.*, 2007a; Plunkett and Swindles, 2008). It is suggested that this is perhaps the result of the island's proximity to the ocean, as a potential modulating factor in the process of solar forcing response. The multi-proxy data presented in these studies is, however, dated using a relatively small number of AMS and conventional ^{14}C dates, tephra horizons and with age depth models constructed using simple linear interpolation through age estimate midpoints. The issues associated with these techniques are discussed at length in Chapter 6, and subsequent age-depth models frequently possess sizeable age errors, thus limiting the confidence with which robust assessments of intra-record correlation can be made. Two events of reduced solar activity, indicated by higher ^{14}C production rates (Fig. 9.4d), do occur broadly within the 3250 – 2300 cal yr BP period. However, while the dating strategy outlined in Chapter 6 was constructed to afford the most robust chronological information possible to this study, it would still be difficult to assess whether variation in solar activity had any bearing on the palaeoecological changes seen here, on the basis of these data. As with the period c. 4300 - 3750 cal yr BP, neither δ^{13}C (Fig. 9.3b) nor δ^{18}O (Fig. 9.3c) record from either site in this study exhibit a significant change during this period. Furthermore the complex period of change, characterised by an initial shift to drier conditions, c. 3250 cal yr BP, appears to precede any major changes in solar activity.

9.5.2 5500 – 5000 cal yr BP

At both Sluggan Moss and Fallahogy Bog a wet shift, beginning c. 5500 cal yr BP, is recorded by all three proxies, suggesting a significant period of climatic deterioration in the region. These conditions persist for approximately 300 years at Sluggan Moss. At Fallahogy Bog, however, this wet shift is interrupted by a short-lived dry period c. 5250 cal yr BP, lasting c. 100 years. Palaeoecological evidence suggests that 'recovery' to previous BSW conditions occurs coincidentally at both sites culminating at the end of the period c. 5000 cal yr BP. The variation in the manifestation of this 'event' at both sites suggests that Sluggan Moss was generally a wetter site than Fallahogy bog, where the effects of such climatic determinations are more likely to persist, uninterrupted for longer periods of time. This suggestion is supported by comparison of the raw palaeoecological evidence from each site, with humification, plant macrofossil and testate amoebae data from Fallahogy Bog all indicative of drier conditions than their equivalent from Sluggan Moss (see Sections 7.2 and 7.3). As previously mentioned, the $\delta^{13}C$ (Fig. 9.3b) and $\delta^{18}O$ (Fig. 9.3c) records from Sluggan Moss respond in phase during this period, both with one another and broadly with the palaeoecological data. As these isotopic excursions reflect neither a regional isotopic shift (i.e. comparing $\delta^{18}O$ records from Sluggan Moss and Fallahogy Bog) nor a record of BSW (i.e. $\delta^{13}C$ does not consistently correlate with palaeoecological data), they are not deemed suitable for inclusion in any wider interpretation (see Section 8.3).

The period 5600 – 5000 cal yr BP was a period of global climatic change, which had significant impact of the human communities across a range of environments (Magny, 2004; Magny and Haas, 2004; Caseldine *et al.*, 2005; Staubwasser and Weiss, 2006). A comprehensive review of the evidence for a major widespread climate reversal at this time is presented by Magny and Haas (2004), and so no attempt is made here to replicate this. In summary, however, the period is considered to represent an inter-hemispheric event during which time cooler conditions prevailed. Various mean annual temperature decrease estimates are presented in Magny and Haas (2004) ranging from 0.75°C in Finland (Heikkilä and Seppä, 2003) to 1.5°C in the European Alps (Bortenschlager, 1977; Haas *et al.*, 1998) and Greenland (Dahl-Jensen *et al.*, 1998). Similar to the 4.2 kyr event, the period 5600 – 5000 cal yr BP is characterised as being drier in central Asia, northern Africa, eastern North America and the southern Mediterranean, with wetter conditions prevailing in west-central Europe and South America (Magny and Haas, 2004).

Unlike the period 3250 – 2300 cal yr BP, during which time climatic changes have often been attributed to solar forcing (see Section 9.5.1), no single mechanism has been offered as a driving force behind the changes recorded 5600 – 5000 cal yr BP. As with the '4.2 kyr event', study of the period is hampered by a series of chronological and theoretical issues. Differences in dating precision, accuracy and resolution, together with varying sample resolution of the proxy records involved hinders our ability to define the exact timing and nature of the event. As a result, Magny and Haas (2004) present a range of possible mechanisms for the change, including changes in orbital forcing, solar activity and ocean-atmospheric circulation. Whilst trying not to replicate this lengthy discussion, it is important to note that a number of common themes exist between the proposed component mechanisms for this period of change and the 4.2 kyr event, including: 1) a weakening of the African and Asian monsoonal systems, linked to cooler SSTs, thus reducing the thermal gradient between land and ocean (Damnati, 2000; Morrill *et al*., 2003); and 2) a migration of the westerlies in response to a stronger thermal gradient between low and high latitudes (Magny *et al*., 2001; Mullins and Halfman, 2001; Lamy *et al*., 2001). These similarities, characterised by suggestions of major changes in ocean atmospheric circulation, allow the rationale for the study of such events in Ireland and the North Atlantic region, as presented earlier (see Chapters 3 and 4), to remain valid for this period.

As with the previous two periods discussed here, the palaeoecological change c. 5500 – 5000 cal yr BP, appears broadly consistent with the latter half of IRD event 4 (Fig. 9.4a), which also happens to be the largest of the three included in the time frame presented in Figure 9.2. As with the period c. 3250 – 2300 cal yr BP, the GISP2 Na concentration, and temperature estimate records (Figs. 9.4c, e) all exhibit good agreement with the palaeoecological data from northern Ireland, 5500 – 5000 cal yr BP. SST in the North Atlantic (Fig. 9.4b), exhibits a less clear relationship, but it could be argued that a downward trend exists in the data similar to, but smaller than that seen c. 3250 – 2300 cal yr BP. Overall, these data considered together indicate a period of significant change in the North Atlantic realm, as recorded in a wide range of climatic archives.

In addition, other Irish records have identified this period as a particularly significant climatic event, potentially resulting in considerable societal impact. Caseldine *et al*. (2005) present data from Achill Island, on the far west coast of Ireland in County Mayo. Here, an especially dry period c. 5800 – 5200 cal yr BP, recorded by humification data and palynological evidence, is concurrent with hypothesised Neolithic settlement in the area (Cooney, 2000). Perhaps of most archaeological importance during this period was the setting out of extensive field systems at Céide, to the northeast of Achill, in Co. Mayo (Caulfield, 1978; Molloy and O'Connell, 1995).

In records from across Achill, this dry period is succeeded by evidence of an extreme inwash event, unique in the 9000 year record, occurring between 5300 and 5050 cal yr BP. Caseldine *et al*. (2005) state that this event was categorically extreme in its magnitude and suggest it was caused by an exceptionally large storm, or series of storms, with rain heavy enough to redistribute sediment over a wide area. At Céide, the field systems were abandoned rapidly c. 5200 cal yr BP and then lay unoccupied as blanket peat expanded across the region (O'Connell *et al*., 1988; Molloy and O'Connell, 1995). Interestingly this period also correlates with the Middle-Late Neolithic transition, which Baillie and Brown (2002) describe as the product of significant disruption to human activity. In addition, Baillie and Munro (1998) also demonstrated that one of the narrowest growth rings in the northern Irish dendrochronological record, potentially indicative of an extreme climatic event, occurs 3195 BC (5145 cal yr BP). Whilst the precise correlation of all these events is beyond the ability of ^{14}C dating, it is clear that the period represents one of considerable environmental and societal change in Ireland.

In summary, palaeoecological data from Sluggan Moss and Fallahogy bog between c. 5500 – 5000 cal yr BP, alongside records from Achill Island, western Ireland, appear to have recorded an event of significant climatic and societal significance, operating on a hemispheric, and perhaps even global scale (Magny and Haas, 2004). However, Caseldine *et al*. (2005) describe the difficulty in obtaining a robust climate signal for this period, as bogs in the region had generally not yet reached ombrotrophic status. Clearly this event is worthy of further study in Great Britain and Ireland, with the potential of multi-proxy peat-based palaeohydrological record from ombrotrophic raised bog, before this study, rarely exploited beyond 4500 cal yr BP (e.g. Charman *et al*, 2006; Swindles *et al*., 2010a). However, of the few long records available in northwestern Europe, many demonstrate a shift to wetter conditions c. 5300 cal yr BP (e.g. Aaby, 1976; Tipping, 1995; Hughes *et al*., 2000; Barber *et al*., 2003; Langdon *et al*., 2003). Indeed, two sites presented earlier, Walton Moss (Fig. 9.2b; Hughes *et al*., 2000) and Bolton Fell Moss (Fig. 9.2f; Barber *et al*., 2003), demonstrate this shift, further reinforcing the potential of the region's peatlands for the study of such mid-Holocene climatic events.

9.5.3 Summary

Aside from the period c. 4300 - 3750 cal yr BP, the stacked palaeoecological records developed during this study exhibit evidence for two other periods of regionally consistent climatic change, typically characterised by shifts towards wetter conditions. The periods 3250 – 2300 and 5500 – 5000 cal yr BP are both consistent with changes in a number of key proxy archives from the North Atlantic region, indicative of significant regional, hemispheric or even global climate change. This validates the potential of ombrotrophic raised bog records from northern Ireland as subjects for multi-proxy, high-resolution palaeoecological studies, as they are potentially: climatically sensitive, regionally representative and able to provide chronologies of the highest precision possible for records of their type. It is proposed that applying the palaeoecological, stable isotopic and chronological methodologies used in this study, to other sites in Great Britain and Ireland would yield high-quality data, capable of furthering our understanding of both mid-Holocene climatic events in the region, and the development of novel techniques, such as the stable isotopic analysis of *Sphagnum* α-cellulose.

9.6 Methodological appraisal

A further benefit of the methodological approach chosen to address the research questions, proposed at the beginning of this study, is an opportunity to assess how peat records can be best employed to identify and/or characterise mid to late-Holocene climatic events. To achieve this, a methodological appraisal will now be conducted, based on both the findings of this study, and our current understanding of the palaeoecological, geochemical and chronological techniques employed here.

9.6.1 Palaeoecological techniques

9.6.1.1 Plant macrofossils

This study encountered a number of potential problems associated with the extraction of a climate signal from plant macrofossil records (Section 8.2.1). These issues centred around: i) climatic complacency, often caused by the dominance of a single eurytopic taxa (i.e. *Sphagnum austinii*) (Blundell and Barber, 2005; Hughes *et al*., 2006); ii) a general practice of grouping non-Section *Sphagnum* taxa at Section level, particularly with regard to section *Acutifolia* taxa which, as a group, possess a variety of hydrological preferences; iii) the limitations of analytical tools, such as DCA and DHI.

Clearly, nothing can be done about a plant macrofossil record's complacency to climatic variations, which merely demonstrates the value of the multi-proxy approach in such situations. However, more can be done to maximise our ability to identify a climatic signal, should one exist in the record. Firstly, wherever possible *Sphagnum* species should be identified to species level. This can be achieved with identifications based on the distinctive stem leaves (Daniels and Eddy, 1990). Admittedly, such material is often not habitually or abundantly available, but any extra information as to the species that is likely present can aid palaeohydrological reconstructions. The subdivision of *Sphagnum* section *Acutifolia*, for example, would enable analytical tools, such as DCA, which assign a unimodal hydrological distribution to each species, to be of more reconstructive use.

The unimodal distributions adopted by DCA, however, severely hamper the effectiveness of this tool, with many species possessing far more complicated hydrological preferences than this overly simplistic model. The non-metric multidimensional scaling (NMDS) technique (Minchin, 1987; Virtanen *et al*., 2006), presented for use in palaeoecological contexts by Daley and Barber (2012), could potentially provide a more ecologically realistic approach. This ordination technique is capable of dealing with non-linear species responses (Faith *et al*., 1987; Minchin, 1987), and is also able to bypass the problem of data being skewed by rare species, which is inherent in DCA, using applying a series of dissimilarity measures to the data. However, the technique is still novel in this context, and a comprehensive review of its application must be conducted.

The original DHI (Dupont and Brenninkmeijer, 1984; Dupont, 1986) has been subtly modified a number of times, to provide a more accurate distribution of species weightings (Barber *et al*., 2003; Daley, 2007; Mauquoy *et al*., 2008; Daley and Barber, 2012). However, it could be argued that an 8-pt scale is not capable of fully representing an accurate hydrological distribution of bog plants' ecological preferences. Whilst Daley and Barber (2012) found the technique to perform favourably in statistical tests, it is felt that there is still significant room for improvement, assigning each species weighting based on extensive field measurements and the adoption of a more representative number of index points (i.e. 50-100). As a result, further development of this technique, could render it a useful tool in the reconstruction of past changes in BSW. However, if the overarching issue, as appears to be the case in the data presented in this study, is that non-climatic factors are driving species changes over and above climatic influence then no amount of statistical development will remedy this problem.

9.6.1.2 Peat humification

The potential limitations of the humification technique have already been discussed (see Section 3.4.2) and autocorrelation between humification and plant macrofossil records, most likely owing to the presence of a species signal in light transmission data (Yeloff and Mauquoy, 2006). Whilst the extent to which this affects the humification signal is largely unknown (Payne and Blackford, 2008), the potential for a relationship is demonstrated well in this study. However, the technique remains a relatively quick and simple way to produce high-resolution, contiguously sampled datasets and should not be dismissed out of hand. Hughes *et al*. (2012) present a method whereby *k*-values of fresh plant material (cf. Overbeck, 1947; i.e. light transmission values of alkali extracts from live samples) can be used to correct the impact of botanical variation on transmission values over time. Hughes *et al*. (2012) applied the technique to records from Newfoundland, eastern Canada, with success, finding that these adjustments were often sufficient to change the timing and number of decadal to centennial-scale 'events', as well as millennial to multi-millennial trends, recorded in data. The technique, however, is again still in its infancy with further efforts required to extend its applicability to other regions of the globe possessing different bog surface assemblages. However, where plant species assemblages exhibit minimal variability humification analyses are still likely to function as a valid climatic proxy. For example, the peatlands of southern Patagonia instances are heavily dominated by *Sphagnum magellanicum*, with little botanical variation exhibited in records developed thus far (e.g. Pendall *et al*., 2001; Mauquoy *et al.,* 2004; Chambers *et al*., 2007), and would therefore avoid the effects of any 'species signal'. In light of the potential limitations outlined throughout this study, it could be suggested that the technique is most suited for use in blanket peatlands where plant macrofossils and testate amoebae are likely to be too decayed for functional analyses and species diversity may be reduced. In such circumstances, degree of humification analyses could still provide a vital proxy for past changes in BSW (e.g. Nilssen and Vorren, 1991; Blackford and Chambers, 1995; Chambers *et al.*, 1997).

9.6.1.3 Testate amoebae

Of the data collected as part of this study, the testate amoebae records were the only ones that produced a coherent record of regional climatic change. The technique, however, was not without its limitations, which, in this instance, principally concerned: i) the representation of the co-dominant *Difflugia pulex* in modern surface samples used to develop European transfer functions; ii) the domination of records by a small number of eurytopic taxa (e.g. *Amphitrema flavum*, *D. pulex*). However, in spite of these issues, testate amoebae are often found to provide more consistent hydrological reconstructions, where other techniques exhibit climatic complacency (e.g. Blundell and Barber, 2005); a suggestion supported by this study. In addition, it has been suggested that testate amoebae are capable of recording short-term environmental variability (Sullivan and Booth, 2007), unlike plant macrofossil records which, it has been demonstrated, have a propensity to lag behind, or even mute, climatic signals (e.g. Blundell and Barber, 2005; Hughes *et al*., 2006). Whilst Warner *et al*. (2007) suggest that a greater understanding of the organisms' relationship with micro-environmental variables is needed to assess the role played by short-term variability (including seasonality) in testate amoebae ecology, it is clear that the proxy possess an immense potential to achieve the ambition set out by Blackford (2000) to provide quantitative climatic reconstructions of past climatic change (see Charman *et al.* (2012) for first attempts at this). From this, coupled with their capacity to respond rapidly to short-term environmental variability, it could be suggested that testate amoebae-based reconstructions, as demonstrated in this study, may represent the best opportunity to document the timing and nature of abrupt climate change events in peat-based palaeoecological studies.

9.6.1.4 Summary

The limitations of each individual palaeoecological proxy, many of which are demonstrated in this study, highlight the importance of a multi-proxy approach to peat-based palaeoclimate studies. By incorporating the proxies discussed above, opportunities are presented to compare, contrast and cross-validate between proxy records. It could be suggested that single-proxy reconstructions should no longer be considered robust records of past climatic variation. In addition, this study has demonstrated that, in some instances, certain proxies perform better than others in providing a coherent climatic signal. As a result, the importance of comparing multiple, multi-proxy records from each studied region cannot be over emphasised. This approach allows for the more comprehensive assessment of regional climate change, mitigating against the effects of site-specific local variability. This ability is of particular importance in the examination of abrupt climate change events, so as to avoid misidentification of a 'regional' event or, conversely, 'false negative' conclusions, formed by examining evidence from a single site and/or single proxy record.

In the context of this study, testate amoebae records were found to perform best, but this cannot be assumed to always be the case. Indeed, it is likely that different proxies function more efficiently under differing conditions (e.g. humification during wet conditions, Charman *et al*., 1999; testate amoebae during dry conditions, Booth and Jackson, 2003). In addition, the study of changing botanical composition through a peat sequence can also yield information that is vital to subsequent climatic interpretations, such as the onset of ombrotrophy and, subsequently, the ability of proxy records to reflect a robust climate signal. Finally, attempts to refine and develop existing techniques are encouraging (e.g. humification *k*-values, Hughes *et al*., 2012; plant macrofossil NMDS/DHI, Daley and Barber, 2012) and further justify each techniques continued use in multi-proxy studies.

Recent findings by Swindles *et al*. (2012a) suggest that peat-based climate reconstructions may be affected by autogenic ecohydrological feedbacks. Swindles *et al*. acknowledged that some shifts in reconstruction WTD may reflect climatic signals, but this may not universally be the case, and further work into the effect of autogenic and allogenic processes on peatland palaeo-WTD is urgently required. However, this serves to highlight the importance of the conclusions of Chambers *et al*. (2012) in a recent review of proxy-climate indicators in peat that not all proxies and techniques have universal applicability. Multi-proxy studies can continue to have relevance if they are developed, as was the aim of this study, within the context of a regional framework of sites, with which comparative results can be compared and assessed for evidence of allogenic climate signals (cf. Payne and Blackford, 2008), amidst noise caused by autogenic ecohydrological responses. This study demonstrates that robust regional climate signals can be extracted from data displaying considerable climatic complacency. In fact, Swindles *et al*. (2012a) too suggest that studies, which compare multiple sites with high-resolution chronologies, are fundamental for helping to identify ‘real’ climate events. This study possesses greater chronological and sampling resolution than the example (i.e. Swindles *et al*., 2007a) cited by Swindles *et al*. (2012a).

9.6.2 Stable isotopic techniques

Unlike early-Holocene abrupt climatic events, which are largely associated with the deglaciation of continental ice-sheets at the close of the last glacial (e.g. Barber *et al*., 1999; Clark *et al*., 2002), periods of mid-Holocene abrupt climatic change have frequently been linked to significant changes in ocean and atmospheric circulation (e.g. Magny and Haas, 2004; Marchant and Hooghiemstra, 2004; Booth *et al*., 2005). Subsequently, based on the suggestions of Daley *et al*. (2010), the $\delta^{18}O$ analysis of *Sphagnum* α-cellulose has immense potential for the analysis of past abrupt climatic events. However, this study provided no evidence that these data reflect a coherent regional precipitation signal, as proposed by Daley *et al*. (2010), despite taking place in a region where $\delta^{18}O$ of precipitation is closely linked to changes in atmospheric circulation (Baldini *et al*., 2010). Similarly, Loisel *et al*. (2010) assert that $\delta^{13}C$ analysis of *Sphagnum* material can be interpreted as a proxy for past changes in BSW. However, the data presented here, again, provide no support for this statement. This suggests that whilst theoretically sound, the application of these proxy methodologies requires further critical investigation before they can be routinely adopted as proxy-climate indicators.

Subsequently, this study identifies a number of problems that must be overcome in order to improve the palaeoclimatic potential of these techniques. Firstly, and perhaps most importantly, a methodological standard for the stable isotopic analysis of peat-based material must be developed as a matter of urgency. Daley *et al*. (2010) present a convincing case for the extraction and analysis of *Sphagnum* α-cellulose. However, this approach is not currently adopted by all workers (e.g. Loisel *et al*., 2010) and results will remain difficult to interpret until methodologies are unified. Similarly, Woodley *et al*. (2011) present an additional methodology, centred around a high-temperature pyrolysis gas chromatography (GC)/isotope ratio mass spectrometry (IRMS) methodology, as opposed to the standard combustion GC/IRMS approach previous practiced, which, can significantly reduce associated temporal and financial constraints. Again, however, this technique has yet to be presented as part of a standardised methodology.

Investigation into the isotopic offset between different components of *Sphagnum* plants (i.e. leaves, branches stems) suggests that offsets do exist and that the isolation of a single component is beneficial to the production coherent records. It could sensibly be suggested that isolating *Sphagnum* stems represents the most logical, time saving option (cf. Loader *et al*., 2007; Moschen *et al*., 2009; Tillman *et al*., 2010), but is still not a practice universally adopted (e.g. Daley *et al*., 2010). Similarly, studies appear to suggest that effects of offsets between *Sphagnum* species are less than that of moisture changes (Rice, 2000; Loisel *et al*., 2009) and that, therefore, species do not require identification prior to analysis.

Until a standardised methodology is offered and widely adopted, understanding of the precise nature of both the $\delta^{18}O$ and $\delta^{13}C$ signal from peatlands is unlikely to improve. Coherent and convincing accounts of the likely mechanisms involved have been provided (e.g. Daley *et al.*, 2010; Loisel *et al.*, 2010), but these relationships must be replicated at a range of other sites. In addition, other issues and factors associated with the stable isotopic analysis of peat material have been raised (Tillman *et al.*, 2010) including: i) variation in species growth rates; ii) poorly understood biochemical pathways in plant tissues; iii) signal preservation; all of which must be addressed in order to better our understanding of the technique's biomechanics. Whether these issues can be overcome to an extent whereby a signal with palaeoclimatic potential can be isolated from these data remains to be seen, however. In terms of sampling strategies, the application of the technique is also governed by the availability of *Sphagnum* for analyses. During periods of *S. austinii* dominance, for example, sampling strategy is largely determinable by the analyst, owing to the generally good preservation of the taxon. However, during other periods, optimal sampling resolutions may be difficult to obtain due to a lack or the poor preservation of suitable material.

Including the stable isotopic analysis of *Sphagnum* material alongside the established suite of palaeoecological techniques outlined above presents an opportunity compare, contrast and cross-validate the isotopic records, which is likely to aid our understanding as to their climatic significance. If robust relationships can be developed, this inclusion would provide a method of assessing the relationship between shifts in BSW (δ^{13}C, palaeoecological proxies) and changes in atmospheric circulation and patterns of precipitation (δ^{18}O), essential for assessing the likely causes of mid-Holocene abrupt climate change events, and also increasing our understanding of changing peatland dynamics over time. Charman *et al.* (2009) suggest that with the advent of isotopic data potentially capable of recording changes in atmospheric circulation (cf. Daley *et al.*, 2010), it may be most effective to interpret BSW records from northwestern Europe as indices for the strength and position of westerly airflow, rather than concentration on temperature and/or precipitation. Given that it is likely that the proportionate role of the two variables has changed over time, this seems a sensible solution. Again, much of this depends on whether a robust climatic signal can be extracted from isotopic data of this sort.

9.6.3 Chronology and 'wet shifts'

To assess the regional manifestation of a given climatic event, the importance of constructing multiple, multi-proxy records for a given region is emphasised above. However, the efficacy of inter-site comparisons hinges largely on the ability to develop a robust chronology. Errors associated with ^{14}C dating and the subsequent age models are frequently >100 cal yrs and so the existence of precisely dated tephra isochrons is of immense chronological value to studies looking to assess for evidence of regional climatic change. Somewhat surprisingly, tephra analysis still does not appear to be as routine as one might expect, given the rapidity and low costs associated with the initial ashing process (e.g. Daley*,* 2007; Daley *et al.*, 2010). Tephrochronology is limited by the spatial distribution of past eruptions, but in locations where horizons can reasonably be expected (e.g. Ireland, Scotland, northern England), tephra analysis must unquestionably be undertaken. In studies with limited access to ^{14}C dates, poor chronologies may well be the limiting factor in the examination of the timing of abrupt climate change events.

A number of methods for assigning 'wet shifts' to multi-proxy palaeoecological data are presented in the literature. Hughes *et al*. (2006) suggest that we should assign wet shifts at any point where two or more proxies are in agreement. However, due to autocorrelative nature of humification in relation to the plant macrofossil records, this is problematic. There is also no guarantee, as demonstrated in this study, that all proxies reflect a robust climatic signal. Loisel *et al*. (2010), on the other hand, focus on wet phases, rather than wet shifts, describing any period during which normalised BSW reconstruction values are below zero as a wet period, and vice versa, which could be considered somewhat arbitrary. It has also been suggested that, in some situations, different proxies can respond to climatic variability on varying time scales (Blundell and Barber, 2005), further complicating the issue of assigning wet shifts/phases.

It may also be noted that current literature is very focussed on wet shifts/phases, perhaps to the detriment of identifying dry phases (cf. Swindles *et al.,* 2010a), which should be assigned equal recognition, given their importance both climatically and archaeologically. As results from northern Ireland in both this study and Swindles *et al*. (2010a) suggest, it may be possible to examine dry phases, as well as wet phases, within the peat records of the region. Charman *et al*. (2007) suggest that water table reconstructions reflect changes in summer water deficit, and therefore dry phases can be interpreted as indicating prolonged periods of high frequency and/or greater magnitude of drought (Swindles *et al*., 2010a). Currently, peat-based palaeoclimatic literature is heavily dominated by the examination of cool/wet phases. However, it could be considered counter-intuitive to be look for wet events in a predominantly wet environment, such as Ireland. Caseldine *et al*. (2005) also found evidence for a significant dry period preceding the 5.2 kyr event signal discussed in Section 9.5.2. Combined, this Irish data suggests that overtly incorporating the study of dry shifts, and their regional commonality, into peat-based palaeoclimatic studies may be advisable and rewarding, helping to further isolate the wet shifts that have been previously identified.

The lack of a standard, objective methodology for identifying wet shifts and/or phases can potentially create problems for the comparison of regional records. The tuning and stacking approach adopted in this study provides a method whereby records from the same region can be combined in order to establish a more robust regional signal before comparison with other palaeoclimatic records. A number of previous studies have combined multi-proxy data into composite BSW records (e.g. Charman *et al*., 1999; Blundell and Barber, 2005; Langdon and Barber, 2005; Hughes *et al*., 2006; Daley and Barber, 2012). The problems associated with this technique have already been discussed, and as the lack of intra-site agreement between proxies in this study shows, the combination of all three palaeoecological proxies would often only mute or remove a climatic signal, as all proxies cannot be guaranteed to reflect a climate signal. The construction and subsequent tuning and stacking of composite BSW records (cf. Langdon and Barber, 2005) should be discouraged.

Blaauw *et al*. (2007a, 2008) present a more objective method of assessing for synchronicity between proxy records, later applied by Charman *et al*. (2009). Rather than assessment by eye, time scales of both proxies remain independent and the probability of increased wetness and/or decreased temperature during a given time-window is calculated for each record. Subsequent probabilities of synchronous change are then calculated. However, this approach was not employed here for a number of reasons. Firstly, Charman *et al*. (2009) found that defined peaks of probable wetness increases were generally only achievable in an extremely well dated sequence (c. 20 ^{14}C over c. 2750 cal yrs) with sequences of lower chronological resolution (c. 8 ^{14}C over 3500 cal yrs) producing poorly defined climatic deteriorations. The chronologies of the sites in this study, whilst better than many typical peat studies, were not deemed sufficiently resolved to produce peaks in BSW change defined sufficiently so as to enable robust assessment of synchronicity. Secondly, the technique is still in its scientific infancy and is not yet applied routinely. If an accessible and well-defined methodology is presented, the 'time-window' technique may well present the most robust chronological tool for the formulation of regional records and assessment of regional palaeoclimatic coherence.

9.6.4 The examination of abrupt climate events in peat records

The ombrotrophic raised bogs of northwestern Europe possess considerable potential for peat-based palaeoclimatic reconstructions, and the examination of mid to late-Holocene climatic events. The findings of this study, however, identify a series of issues, which if addressed could optimise such investigations. Firstly, studies should employ a multi-site, multi-proxy palaeoecological approach to enable the inter- and intra-site comparison of proxy records to assess for the presence of a robust regional climatic signal. Subsequently, the presence of abrupt climatic events can be confirmed as regionally significant occurrences, and not merely locally specific variability. The efficacy of different palaeoecological proxies appears to vary from site to site and even through time and, ideally, no one proxy should be relied upon more than another.

Clearly, chronology is an issue in studies looking to assess the timing and nature of abrupt climatic events. However, the chronological potential of the rapidly accumulating raised bog systems of northwest Europe, facilitated by an abundance of material for ^{14}C dating and the presence of numerous precisely dated tephra horizons. As a result, these two techniques should be practiced habitually in an effort to provide the most robust chronology possible. Ideally, the chronologies of records intended for comparison and/or combination should be of comparable resolution and associated error.

Whilst stable isotopic analysis of peat material could potentially significantly aid understanding of the mechanisms responsible for abrupt climatic change (i.e. changes in atmospheric circulation as recorded by δ^{18}O), based on the isotopic data of this study, current methodologies and understanding are not at a level suitable for palaeoclimatic interpretation. The use of δ^{13}C as a proxy for BSW is subject to the same limitations. However, their continued exploration based on a standard methodology will hopefully elucidate current issues and allow more routine use in future.

Recent NERC-funded peat-based palaeoclimatic projects (e.g. PRECIP, PATAGON) adopt many of these suggestions, including the development of high-resolution, multi-proxy, multi-site and multi-isotope studies with comprehensive ^{14}C dating and tephrochronology. Based on the evidence considered in this study, the approach adopted by these projects could be considered the optimal methodology, not just for the analysis of abrupt climate change, but also for wider peat-based studies of Holocene climatic change. It is also vital that refined and novel techniques for data analysis and interpretation, many of which have been discussed during the course of this section, are developed and adopted when possible.

9.6.5 The manifestation and definition of 'events' in the palaeoclimatic archive – a conceptual challenge

It seems that a signal, which could feasibly be referred to as an 'event' (i.e. a short-term event, of significant magnitude and a definite start and finish) around 4.2 kyr event, is only really witnessed with high-levels of replication in certain regions, such as the tropics and mid-latitudes (Staubwasser and Weiss, 2006). If the global 4.2 kyr event is interpreted as a manifestation of a complex set of teleconnections and interactions between a number of ocean-atmosphere circulation systems, as appears a likely scenario, Arz *et al.* (2006) question why other multi-centennial variations did not lead to a response comparable to the 4.2 kyr event. The answer could lie in regional feedback mechanisms and the thresholds within them. The dominance of an 'event' signal in climatically sensitive, potentially semi-arid regions, and a 'trend' everywhere else, suggests that 'tipping points' are playing a role in the 4.2 kyr event, in these regions at least. The Red Sea, for example, is known to amplify environmental signals (Arz *et al.*, 2003, Siddall *et al.*, 2003).

A number of records, however, seem to not show a specific 'abrupt' event, but rather a shift in the climate system, occurring c. 4ka BP. Such observations could represent shifts in certain aspects of the global climate system, which have resulted in 'abrupt' event-like response in certain proxy records. For example, in the titanium record of the Cariaco basin, which acts as a proxy for ENSO strength, Haug *et al.* (2001) document a shift to increased ENSO strength and variability that commences c. 4 ka BP. As already discussed, Johnsen *et al.* (2001) suggest that c. 4000 cal yr BP marks the point at which optimum Holocene warmth in Greenland terminates and temperatures commence a downward trend through the late-Holocene. A speleothem record from Oman, whilst not displaying evidence of an abrupt 'event' shows a long term drying trend, indicative of weakening in the Indian Summer monsoon system (Fleitmann *et al*., 2003). Gupta *et al*. (2003) identified a trend of monsoonal weakening related to North Atlantic SST minima, based on a marine record of upwelling off the Omani coast.

Whilst SST variation could account for some of the 4.2 kyr event's effects it cannot necessarily account for the rapidity of the change seen in some of the records (e.g. Cullen *et al*., 2000; Thompson *et al*., 2002). We must assume no linear relationship between SST and precipitation, based on studies conducted in regions, such as the South American tropics (Kane, 2002). Subsequently, this further hints at the existence of 'tipping points' in this relationship, affecting certain regions more than others. For example, the subtropical African climate has been shown to flip abruptly between humid and arid modes in response to summer radiation (the primarily forcing mechanism of the summer monsoon) passing a specific, critical value (deMenocal *et al*., 2000). It is perfectly feasible that similar situations could occur in other regions of the world, and their associated climate systems. Subsequently, this potential duality of the climatic signal at this time (i.e. event vs. trend) suggests that the 4.2 kyr 'event' could represent a threshold response to a wider climatic bifurcation in climate, which would validate its use by Walker *et al*. (2012) as a GSSP marker for the Mid-Late Holocene Boundary.

However, the suggestion of Walker *et al*. (2012) to use the 4.2 kyr event, as a new Middle-Late Holocene Boundary GSSP marker, to subdivide the Holocene epoch requires that we are able to identify it in a range of climatic records, and subsequently correlate between archives. This study's results demonstrate the difficulties associated with the identification of single events in peat-based palaeoclimatic records well. However, the problems associated with testing for synchronicity between a variety of climatic archives are numerous. Firstly, we have trouble defining precisely what section of each record represents the 'event' in question. For example, dramatic peaks in aeolian dust found in ice or marine cores (e.g. Fig. 2.2f, h, i) have reasonably easily defined 'start' and 'end' dates. However, for some other records (e.g. Fig 2.2e, g, j) defining the event is rather more difficult. This is because the signal produced by each proxy is the result of a variety of modulating factors, which record and/or respond to the same driving forces behind the event each in a different manner. In addition, we must consider the added unknown of tipping elements not only in the earth climate system, but also in ecosystem response, which is subsequently recorded in proxy and sedimentary records. For example, does a dust peak, such as those seen in Figures 2.2f, h and i represent the full duration 4.2 kyr event, or merely a tipping point within that at which dust was significantly mobilised. Equally, do some of the oxygen isotope records accurately record the duration of the event or do they merely 'sway' as forcing mechanisms wax and wane in their strength or influence? For these reasons, even if all records were based on absolute, annual chronologies, testing for synchronicity would still be impossible in its most literal sense. Parnell *et al*. (2008, p. 1873) state that "*if we cannot say what an event is we cannot always usefully discuss with others the uncertainty in the unique depth [in each record] and age that we propose to associate with it*". The inherent noise witnessed in the majority of proxy records only serves to confound problems of event definition. In addition, the manifestation periods of significant climatic change as 'events' in certain records, and 'trends' in others adds further difficulty to the task. Of course absolute chronologies will never be available and due to the natural variation in the nature of each record, they possess very different age-models, which have been constructed using a number of different approaches and methods. Subsequently, we may not be able to extract robust dates for the 'start', 'end' or 'duration' of an event, even if we could pinpoint them. However, the time-window approach discussed earlier (Blaauw *et al*., 2007a, 2008; Charman *et al*., 2009) may go some way in providing a more objective method of assessing for synchronicity between proxy records.

Chapter 10 – Conclusions

10.1 Specific conclusions from the research

In Chapter 1 a set of research questions were proposed, as follows:

1. Based on existing evidence, is it possible to characterise the timing and nature of a global '4.2 kyr event'?
2. Can a coherent '4.2 kyr event' signal be characterised in Great Britain and Ireland?
3. Can peat-based stable isotopic analysis contribute to our understanding of the nature and/or cause of the '4.2 kyr event'?

By addressing these questions during the course of this study, the following conclusions can now be reached:

- Evidence for the '4.2 kyr event' can be found on all seven of the earth's continents, centring broadly around c. 4000 cal yr BP. Evidence for the event is strongest, but not limited to, the lower latitudes. The broad spatial scale, severity of its effects in some regions, and manifestation in the archaeological record highlight the period as one of great palaeoclimatic importance and societal interest. The 4.2 kyr event was most likely caused by a complex set of interactions within the global ocean-atmosphere circulation system.

- The manifestation of the 4.2 kyr event in Great Britain and Ireland is unclear. Although some previous peat-based data suggested the existence of this event, this study found no compelling evidence support this, despite using methods and site selection criteria specifically designed with this purpose in mind. The high-resolution, multi-proxy palaeoecological and stable isotopic records from the North of Ireland provided the best opportunity to examine this event, in terms of the potential for climatic sensitivity and chronological constraint, in northwest Europe.

- Based on these data, novel methodologies for peat-based stable isotopic analysis cannot yet be confidently accepted as robust proxies for past changes in atmospheric circulation change and/or BSW. As a result, these analyses were not able to contribute to the understanding of the nature and/or cause of the 4.2 kyr event. It is suggested here that a lack of understanding of the biomechanics of stable isotope incorporation into *Sphagnum* plant tissues, together with the lack of standardised methodologies is significantly hindering the potential of the technique's application in peat-based palaeoclimatic studies.

- This study finds that owing to the climatic complacency of proxy records in certain instances, a multi-proxy, multi-site approach to the examination of Holocene climatic change, and the events within that variability, is essential. This approach enabled climatically insensitive records from both sites to be identified and excluded from further analyses. This study also found that by fully exploring the potential of both radiocarbon and tephrochronological dating techniques, inter-site comparisons can be far more robust and should be practiced to the greatest extent possible.

10.2 Recommendations for future research

- Caution should be exercised in the use of the 4.2 kyr event as a GSSP marker for the Middle-Late Holocene Boundary. The review presented in Chapter 2 confirms the assertions of Walker *et al*. (2012), with regard to the event's manifestation in a wide range of geomorphological, stratigraphical and archaeological records from many parts of the world. However, this study also demonstrates the difficulty in assigning an accurate and precise age to this event, due to variations in the quality of chronologies involved and the varying response of different climatic records to a common set of forcing mechanisms.

- The period c. 5500 – 5000 cal yr BP is clearly worthy of further palaeoclimatic investigation. Data presented in this study support assertions made previously that an event, of significant climatic and societal significance, occurred during this time not only in Ireland (Caseldine *et al*., 2005), but potentially operating on a hemispheric, and perhaps even global scale (Magny and Haas, 2004).

- The effects of non-climatic and/or allogenic forcing factors of BSW/WTD reconstructions were minimised in this study through careful site selection and intra-site proxy record comparison. However, certain palaeoecological proxy records (e.g. plant macrofossil and humification) still exhibit signs of climatic complacency. This suggests that the construction of composite BSW curves and the identification of wet shifts based upon multi-proxy climate curves should never be attempted without confirming first that each record used represents a robust regional climatic signal. This can be achieved through the inter-site comparison of proxy reconstructions.

- The development of standardised methods for the stable isotopic analysis of material from raised bog peat sequences is vital. Currently methodologies differ dramatically in terms of their chosen sample preparation and mass spectrometry technique. After extensive consideration of the literature, it is suggested that the methods followed here (i.e. isolation of *Sphagnum* stem material, extraction of α-cellulose and combined isotope analysis (i.e. $\delta^{13}C$ and $\delta^{18}O$) through high-temperature pyrolysis GC-IRMS should be accepted as the standard, owing to its ability to produce a robust reconstruction of the isotopic composition of plant tissues over time, whilst also minimising financial and time costs.

- Only after a standardised methodology is accepted within the community can results begin to be robustly compared and contrasted. Until then our understanding of the precise nature of the peat-based stable isotopic signal is likely to remain poor. Existing suggestions as to the nature of the signal (i.e. $\delta^{18}O$, changes in past atmospheric circulation; $\delta^{13}C$, changes in past BSW) cannot be confirmed by the data produced as part of this study. If, after robust intra-site comparisons, firm conclusions cannot be made, it could be suggested that the potential of the isotopic proxy is likely to be limited by our understanding of the biomechanics of isotopic fixation in *Sphagnum.* Subsequently, detailed studies into this topic must be undertaken, but it difficult to see how this problem might be remedied.

- An ability to correlate changes in BSW/WTD in the peat bogs of northwest Europe, and elsewhere, with major changes in atmospheric circulation is clearly of great palaeoclimatic benefit. It may also further our understanding as to the dominant drivers of raised bog hydrology. If stable isotopic analyses cannot produce such information then it is recommended here that other techniques (e.g. dust and sand influx; see de Jong *et al*., 2010 for review) should be extended in their use. Understanding the response of atmospheric moisture balance, as recorded by raised bog sequences, to periods of abrupt change brought about by significant atmospheric circulation has extremely important socio-economic implications, in terms of our ability to predict likely response to future periods of climatic change.

Appendix

Table A.1 Tephra geochemistry results from Sluggan Moss (italicised samples rejected due to poor totals)

Label	**Na_2O**	**MgO**	**Al_2O_3**	**MnO**	**FeO**	**K_2O**	**CaO**	**SiO_2**	**TiO_2**	**Total**
E_5_15	4.22	0.06	13.21	0.11	1.78	2.73	1.22	73.07	0.12	96.51
E_5_14	4.60	0.03	13.00	0.12	1.66	2.62	1.27	72.51	0.16	95.99
E_5_9	4.44	0.01	12.34	0.06	1.80	2.54	1.30	68.46	0.10	91.05
Avg.	4.42	0.03	12.85	0.10	1.75	2.63	1.27	71.35	0.13	94.52
SD	0.19	0.02	0.46	0.03	0.08	0.09	0.04	2.52	0.03	3.01
E_5_10	*4.20*	*0.03*	*11.20*	*0.06*	*1.71*	*2.32*	*1.12*	*62.43*	*0.06*	*83.12*
E_5_7	*3.61*	*0.01*	*11.00*	*0.09*	*1.65*	*2.21*	*1.07*	*62.08*	*0.13*	*81.86*
E_5_11	*3.42*	*0.01*	*10.64*	*0.09*	*1.50*	*2.16*	*1.02*	*59.98*	*0.13*	*78.95*
E_5_6	*3.66*	*0.05*	*10.54*	*0.04*	*1.45*	*2.13*	*1.05*	*59.02*	*0.06*	*78.00*
E_5_5	*3.39*	*0.01*	*9.84*	*0.06*	*1.32*	*1.96*	*0.98*	*55.45*	*0.11*	*73.12*
E_5_12	*2.19*	*0.02*	*6.66*	*0.11*	*1.02*	*1.35*	*0.72*	*39.17*	*0.05*	*51.28*
E_5_8	*1.85*	*0.01*	*6.41*	*0.00*	*0.92*	*1.31*	*0.58*	*37.34*	*0.03*	*48.45*
E_8	*2.1578*	*0*	*6.6495*	*0.1253*	*1.1745*	*1.3438*	*0.7415*	*38.7745*	*0.0256*	*50.99*
E_7	*3.637*	*0.0329*	*10.884*	*0.0763*	*1.5444*	*2.2512*	*1.068*	*61.007*	*0.0792*	*80.58*

Table A.2 Tephra geochemistry results from Fallahogy Bog

Label	**Na_2O**	**MgO**	**Al_2O_3**	**MnO**	**FeO**	**K_2O**	**CaO**	**SiO_2**	**TiO_2**	**Total**
41 / 1	4.4721	0.0168	12.3364	0.0535	1.7271	2.7193	1.3596	71.2709	0.1159	94.1437
42 / 1	4.9767	0.0159	12.7490	0.1266	1.8911	2.8327	1.3314	73.4277	0.2118	97.5629
43 / 1	4.7970	0.0243	12.7724	0.0040	1.9258	2.8218	1.3371	71.5046	0.0936	95.2806
44 / 1	4.8688	0.0107	12.5749	0.0927	1.9753	2.7598	1.2197	71.2839	0.0918	94.9295
45 / 1	4.7843	0.0146	12.6341	0.1105	2.0572	2.9334	1.2772	72.1878	0.0682	96.0924
46 / 1	4.4297	0.0101	12.0378	0.0670	1.8548	2.8236	1.3468	71.7668	0.0972	94.4337
48 / 1	4.4147	0.0218	12.2232	0.1075	2.1446	2.7846	1.3361	71.7955	0.0724	94.9250
49 / 1	4.7579	-0.0032	12.5483	0.0676	1.9984	2.8155	1.2360	71.9852	0.1290	95.6152
51 / 1	4.4896	0.0230	12.3797	0.0872	1.6824	2.8325	1.3113	71.2542	0.0738	94.1336
52 / 1	4.3765	0.0240	12.4368	0.1421	1.7344	2.7137	1.3132	71.3059	0.0477	94.0942
53 / 1	4.6160	0.0159	12.4239	0.1176	1.8474	2.7752	1.2537	71.5342	0.0778	94.6909
54 / 1	4.8109	0.0269	12.3455	0.0560	1.8423	2.8398	1.3916	72.1362	0.1254	95.5748
55 / 1	4.6621	0.0259	12.2681	0.0300	1.8907	2.7805	1.2597	71.7155	0.0299	94.6624
56 / 1	4.4868	0.0153	12.4274	0.0763	1.9592	2.9254	1.2575	71.9477	0.1653	95.2610
58 / 1	4.6087	0.0171	12.4431	0.0671	1.9280	2.7723	1.3739	72.6448	0.1687	96.0237
59 / 1	4.8017	0.0036	12.4739	0.0343	2.0991	2.9373	1.2636	72.0222	0.0884	95.7429
61 / 1	4.5478	-0.0070	12.1971	0.0873	1.9228	2.9512	1.2802	71.4105	0.1025	94.4994
63 / 1	4.6066	0.0097	12.1474	0.0328	2.1366	2.7730	1.3418	71.0792	0.0883	94.3081
64 / 1	4.5945	0.0137	12.4256	0.1143	1.9101	2.8690	1.3630	71.2295	0.1166	95.0023
65 / 1	4.8318	0.0203	12.5666	0.0859	1.7471	2.8231	1.4012	72.0615	0.0654	95.6279
68 / 1	4.6312	0.0279	12.1391	0.0826	1.9203	2.7976	1.2477	71.3171	0.0914	94.4891
71 / 1	4.8499	0.0183	12.5389	0.0031	2.0489	2.8767	1.4066	73.2798	0.1250	97.1471
72 / 1	4.4571	0.0160	12.4445	-0.0553	1.9442	2.8860	1.3033	71.1737	0.1043	94.4036
75 / 1	4.5703	0.0057	12.1665	0.1572	1.9520	2.7699	1.3438	71.0673	0.1184	94.1782
Avg.	4.6434	0.0153	12.4042	0.0728	1.9225	2.8256	1.3148	71.7667	0.1029	95.1176
SD	0.1683	0.0090	0.1913	0.0484	0.1238	0.0667	0.0546	0.6385	0.0399	0.9268

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