Climate, Vegetation and the Complex History of Pinus sylvestris during the Holocene, in Wester Ross, Northwest Scotland.

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Abstract

The close proximity of the North Atlantic to Wester Ross makes the vegetational, climatological and anthropogenic history of the region particularly interesting. This study aims to reconstruct these three aspects of the Holocene from peat deposits at Meall Bad á Chrotha. Such studies are vital if we are to improve our understanding of current climate change and prepare for the future implications it may cause.

The warming climate of the Lateglacial and early Holocene saw the terrestrialisation of the lake which occupied the site and subsequent establishment of mixed *Betula*/*Corylus* woodland. The continued rise in summer temperatures led to the Mid-Holocene Thermal Maximum and the associated replacement of mixed woodland by *Pinus sylvestris*. It is likely that dense pine forest occupied the area by 7000 cal BP, encouraged by the drier conditions and possibly the occasional, small-scale wild or domestic fire. The lack of evidence for Mesolithic-Neolithic activity within the record is fairly unsurprising due to the high altitude and exposed nature of the site; however, there is evidence for early activity elsewhere on the South Erradale Peninsula.

The 'Mid-Holocene Pine Decline' is a distinctive part of the vegetational history of Meall Bad á Chrotha most probably occurring in response to a substantial reduction in summer temperatures. This climatic deterioration known as the 'Holocene Thermal Decline' and subsequent wet shift can be linked to an ice rafted debris event which occurred in the North Atlantic around 4200 years ago (Bond et al, 1997). Minor rises in *Plantago lanceolata* and charcoal from the mid-late Holocene most likely represent the intensification of land use in the valley below as communities exploited the newly open landscape, and were perhaps responsible for the continued fall in arboreal pollen. The temperature continues to fall below early Holocene values with a rise in water-tolerant communities and peat-forming conditions towards present day.

Table of Contents

List of Tables

List of Figures

Abbreviations

BP: Before Present

Cal BP: Calibrated or calendar years before present

MBAC: Meall Bad á Chrotha

MTWA: Mean temperature of the warmest month.

r^2 : Coefficient of determination – The proportion of total variance explained by the regression.

RMSEP: Root mean squared error of prediction – RMSEP measures the predictive abilities of a train set.

TLP: Total Land Pollen

WA-PLS: Weighted averaging partial least squares regression

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Chapter 1: Introduction

1.1 The Importance of Reconstructing Past Environments and Holocene Climate Research

The subject of climate change is one of the most relevant topics of today; however, in order to understand how climate change can affect the world we live in, it is vital to understand how past climate change has transformed and shaped the landscapes around us today (Le Truet et al, 2007). It is clear that human beings are having a discernible impact on global climate and profound changes are underway. Gauging the severity, likely long term effects and possible consequences of these changes that we are inducing can be determined through the reconstruction of past environments (Oldfield and Iversen, 2003).

Holocene climate research was initially driven by interest and curiosity about the past. However, now there is a major environmental concern pushing the need for this research forward: Global Warming. This research is now essential in providing information about natural climate fluctuations since the last glacial maximum (Birks, 2007). The Holocene is characterized by intense climatic fluctuations (Mayeski et al, 2004); however, there is still much work to be done if the cause and consequence of these fluctuations is to be fully understood. During the 21st century, it is anticipated that greenhouse gas levels will rise to levels that have not been witnessed in millions of years resulting in climate change. Indeed we can see the changes to climate already occurring (NOAA, 2014). Recent data suggests that the first half of 2014, from January-June was the third warmest on record. The average land and ocean combined surface temperature for the globe, for the period of January to June for the 20th century was 13.5°C (56.3°F). The average for the same period in 2014 was 0.67°C (1.21°F) warmer than the 20th century average (NOAA, 2014). June 2014 has set a new record for the highest combined global average land and ocean surface temperature reaching 0.72°C (1.30°F) over the average of 15.5°C (59.9°F) for the 20th century (figure 1 and figure2 on page 3) (NOAA, 2014). In order to prepare for any climate variability within the next 50-100 years it is essential that observations in past climate are carried out to see how previous climatic fluctuations affected the environment.

Research into past climatic events and ocean-atmosphere-ice feedbacks mechanisms which in particular involve the North Atlantic Thermohaline Circulation is vital as the increases of atmospheric CO_2 is predicted to, in turn, increase the levels of freshwater input into the

North Atlantic through a surge in the melting of ice sheets (Anderson, 1997). This would trigger a shift in the North Atlantic Circulation which would lead to a fundamental climatic event. A significant change in the North Atlantic Thermohaline Circulation is closely linked to the Younger Dryas climatic event. Discharge of Laurentide ice along with an increase of melt water from around the North Atlantic could have triggered this shift in the North Atlantic Thermohaline Circulation. This shift consequently created an episode of near glacial conditions due to a decrease in northward heat transport (Anderson, 1997). The Intergovernmental Panel on Climate Change described such an event as one of the potential 'surprises' which could be the result of the unpredictable, nonlinear responses of the climate system to anthropogenic perturbations (IPCC, 1996).

"Climate has changed on all time scales throughout Earth's history. Some aspects of the current climate change are not unusual, but others are. The concentration of CO_2 in the atmosphere has reached a record high relative to more than the past half-million years, and has done so at an exceptionally fast rate. Current global temperatures are warmer than they have ever been during at least the past five centuries, probably even for more than a millennium. If warming continues unabated, the resulting climate change within this century would be extremely unusual in geological terms. Another unusual aspect of recent climate change is its cause: past climate changes were natural in origin, whereas most of the warming of the past 50 years is attributable to human activities." (Jansen et al, 2007).

Variations in the concentrations and radiative forcing associated with the three dominant atmospheric greenhouse gases carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O) were minor and were *"likely mostly due to natural processes"* compared to the continuous level of increase over the past century and is "*very likely unprecedented in at least the past 16kyr"* (Jansen et al, 2007). Data collected from ice cores show that variations in the last 650kyr in CO_2 remained between 180 and 300ppm while CH_4 remained between 320 and 790ppb. Therefore, "*it is very likely that the current atmospheric concentrations of CO_2 (379ppm) and CH_4 (1,744ppb) exceed by far the natural range of the last 650kyr"* (Jansen et al, 2007). The concentrations of CO_2 and Antarctic temperatures fluctuate simultaneously indicating a very close connection between the carbon cycle and climate (figure 3, page 5). *"It is likely that earlier periods with higher than present atmospheric CO_2 concentrations were warmer than present"*. This can be seen on a millennium scale warming events (Pliocene - 5 to 3 Ma) and events that last a few hundred thousand years (Palaeocene-Eocene Thermal Maximum, 55Ma) (Jansen et al, 2007). *"It is very likely that* *the global warming of 4°C to 7°C since the Last Glacial Maximum occurred at an average rate about 10 times slower than the warming of the 21st century".* Climate models suggest that due to fluctuations in greenhouse gas concentrations and ice sheet dynamics during the Last Glacial Maximum, which occurred about 21Ka, temperatures were 3°C to 5°C lower than modern day (Jansen et al, 2007).

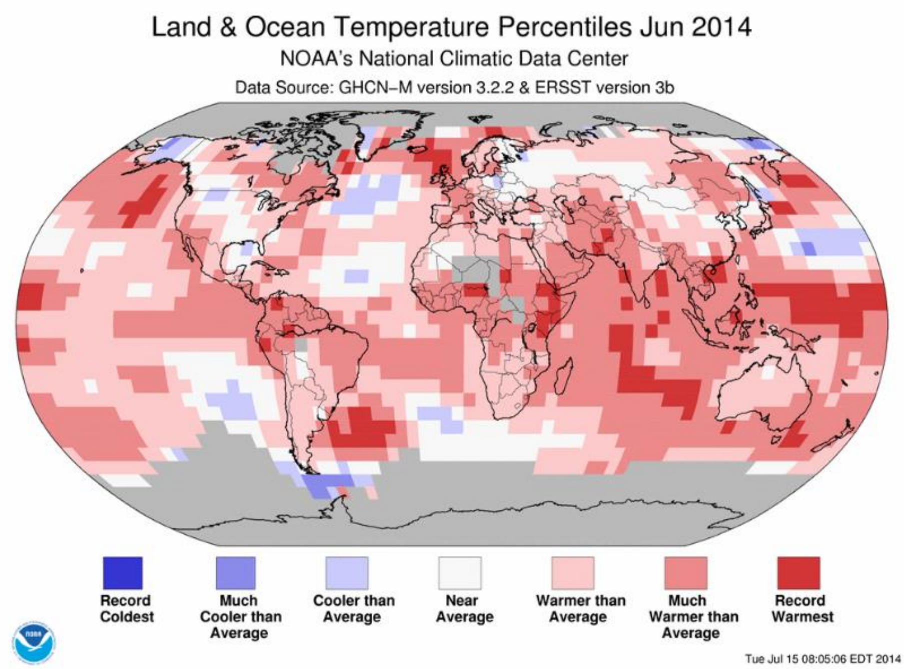


Figure 1. The combined global land and ocean temperature percentiles for June 2014 (NOAA, 2014).

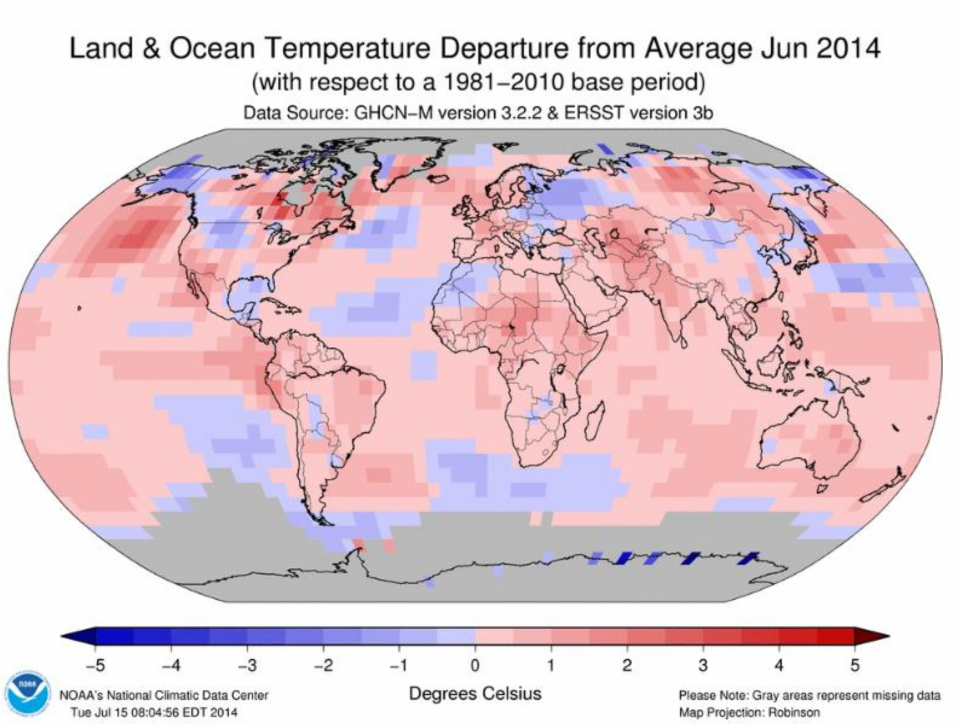


Figure 2. The departure from average for the land and sea temperature combined for June 2014 (NOAA, 2014).

Climate proxies allow us to reconstruct past climate when long term, widely distributed temperature observations are unavailable. Climate information, such as precipitation and temperature, can be recorded by different proxies in different ways, and at different time scales (Chambers et al, 2012). Tree rings, borehole temperatures and pollen abundance are the three most commonly used climate proxies. The density and width of tree rings and their ability to decipher seasonal temperatures through their accuracy of dating to the year make this proxy the most frequently and extensively used (Li et al, 2010). However, when centennial or longer term climatic interpretation is needed, tree ring proxies are restricted by the technique used to eliminate the non-climatic variations from the tree ring series (Cook et al, 1995; Briffa and Melvin, 2008). In comparison to tree rings, borehole temperatures are used to interpret longer time scales. NH continental temperature has been reconstructed for the past 500 years using a borehole depth temperature profile (Huang et al, 2000; Harris and Chapman, 2001; Chapman et al, 2004). The variations in temperature are recorded and preserved by the profile as the heat from the surface is diffused down into the earth (Li et al, 2010). Pollen is often used to fill the missing links between borehole temperatures and tree rings as it is sensitive to multi-decadal climate variability (Li et al, 2010).

"Palaeoenvironmental data indicate that regional vegetation composition and structure are very likely sensitive to climate change, and in some cases can respond to climate within decades."

(Jansen et al, 2007).

We can reconstruct the long-term development of peat-forming wetlands by taking a core through the peat and identifying plant remains; building a picture of vegetation change during mire formation. If the ecological requirements of species are well known, past changes in vegetation can be used to reconstruct past environments (Ellis, 2005). Pollen analysis is a key technique for longer term climatic and vegetation history, as climate change can easily be deduced from pollen-analytic information (Faegri and Iversen, 1989). The basic method is to define the ecologic niche(s) of the taxon, however pollen analysis as a single proxy can be unreliable as the vegetation lags behind in its response to climate change (The Committee on Surface Temperature Reconstruction for the Last 2,000 years, 2006). In theory, a multi-proxy approach is preferable to allow comparison between proxies, however in practice, this is rarely executed owing to the large amount of time and funding required (Blundell et al, 2004).

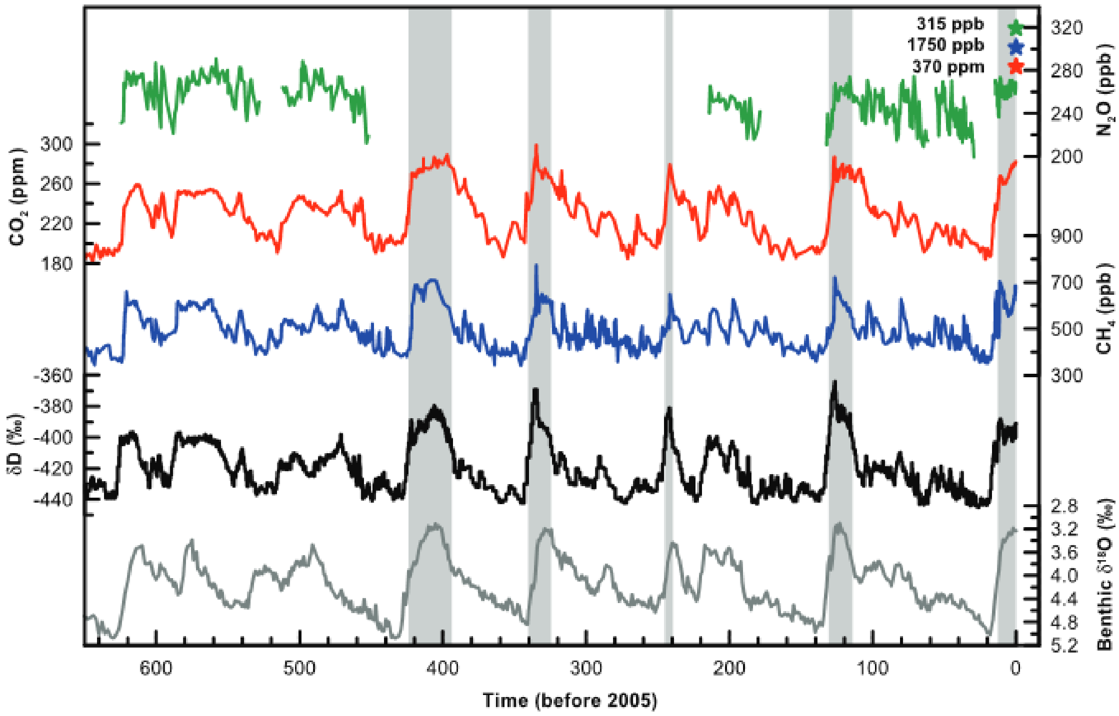


Figure 3. Variations of deuterium (δD; black) which a local temperature proxy, and the concentrations of atmospheric greenhouse gases (CO_2 (red), CH_4 (blue), N_2O (green)) collected from trapped air within Antarctic ice cores and recent atmospheric measurements for the last 650kyr (Jansen et al, 2007; Petit et al, 1999; Indermühle et al, 2000; EPICA community members, 2004; Spahni et al, 2005; Siegenthaler et al, 2005a, b). Previous interglacial warm periods is shown by the grey shading. The stack of 57 globally distributed benthic $\delta^{18}O$ marine records (dark grey), a proxy for global ice volume oscillations (Lisiecki and Raymo, 2005), is displayed for comparison with the data from the ice cores. Downward trends in the benthic $\delta^{18}O$ curve reflect increasing ice volumes on land. The shaded vertical bars are based on the ice core age model (EPICA community members, 2004), and the marine record is plotted on its original time scale based on tuning to the orbital parameters (Lisiecki and Raymo, 2005). The stars and labels show atmospheric concentrations at year 2000.

1.2 Introduction of Pollen Analysis and the Principles of Palynology

Dane Lennart Von Post introduced pollen analysis in 1916 and since then pollen has been regarded as one of the most important and widely used proxies in Quaternary Palaeoecology. The term palynology was introduced to cover a wider aspect, taking into account not only pollen but other resistant organic walled microfossils. The term ‘pollen analysis' is reserved for the original concept of fossil pollen in a geologic context (Faegri and Iversen, 1989).

The introduction of pollen analysis signified the importance of mires which are now regarded as invaluable natural archives of past environmental conditions. Mires can preserve pollen, along with non-pollen palymorophs, in a remarkable state of preservation for thousands of years. As well as inferring climate from vegetation records, peat humification and changes in the straigraphy of the peat reflect past climate variations. Cooler and wetter climatic conditions can be interpreted from a decrease in the level of peat decomposition (Blackford and Chambers, 1995). Much of our knowledge about climate, vegetation and human activity is derived from peat deposits.

The species composition of the peat forming vegetation reflects any variations in the levels of precipitation and temperature (Chambers et al, 2012). The primary aim of a pollen analytical study is to reconstruct past vegetation; however the ability to interpret the conditions under which this vegetation grew is also an important aspect of any palynological study. If there is little or no human interruption, vegetation will reach a stage of equilibrium with the climate and the environment in which it grows (Roberts, 1998); therefore climatic variable reconstructions can be achieved by applying palynological vegetation histories to modern data on plant-climate interactions (Birks, 1981).

1.3 Research Questions, Objectives and Aims

The aim of this research is to reconstruct the past vegetation of Meall Bad á Chrotha, Wester Ross, which in turn will allow the reconstruction of past climates through modern data on plant-climate reconstructions therefore improving our understanding of the impacts climate variations can have on the environment. For centuries there has been much speculation about the cause of the fluctuations in *Pinus sylvestris*, therefore studies such as this are necessary to try and gain an understanding of how and why these fluctuations occurred as pine woodlands are one of Scotland's most striking and unique forest types. Pine is also a good indicator of any hydrological change, therefore a reconstruction of its complex history will aid the reconstruction of any hydrological shifts. Furthermore, a reconstruction of vegetation and charcoal allows the interpretation of any human activity that may have occurred in the area during the Holocene and what influence such activity had on past environments and the landscape that we see today.

Research Questions:

- What might have led to the final pine decline in the mid-late Holocene after c.4000 cal BP? Was the decline more likely to be related to human or physical impacts such as climate change, increased oceanicity or a volcanic event i.e Hekla?
- Are there any regional or local variations as to when the sudden expansions and declines, in particular the final decline, in *Pinus sylvestris* occur? And is there any evidence at Meall Bad á Chrotha of a two-phase decline similar to the Eilean Subhainn and Glen Carron studies?
- What evidence is there, such a *Plantago lanceolata* or Cerealia, to suggest that Neolithic and later peoples were present in Wester Ross? If so, what influence would these peoples have had on the vegetation?
- Do the records give evidence for an increase in fire activity during the late Holocene?

Chapter 2: Literature Review

2.1 The Late Devensian (Lateglacial) in Scotland

2.1.1 The Wester Ross Re-advance

Robinson and Ballantyne (1979) provide evidence which suggests a re-advance of the Late Devensian ice sheet or Dimlington ice sheet (McCormack, 2011) which predated the Loch Lomond Advance or the Lateglacial Stadial, in the Northwest highlands, naming it the Wester Ross Re-advance. Throughout Wester Ross there are extensive ice marginal moraines which indicate the former ice limit (Figure 4).

This re-advance, which occurred during the retreat of the Late Devensian ice sheet and the beginning of the Lateglacial Interstadial, can be seen in five separate moraines throughout Wester Ross. The Red Point Moraine is an uninterrupted >10km ridge ending at the coast of Opinan. There are also extensive moraines at Applecross, Gairloch, Aultbea and An Teallach (figure 5.1-5.5). A minimum age of 14,920 cal BP for the moraines in Wester Ross has been given after the radiocarbon dating of organic material from Loch Droma (Kirk and Goodwin, 1963).

There are a number of different dates that have been given for the Wester Ross Re-advance. 19,400 cal BP was the mean age given by Everest et al (2006) for the Re-advance moraines in Torridon, further south, after the use of cosmogenic isotope dating. Although, sandstone boulders from south Torridon indicate, through exposure dating carried out by Ballantyne et al (2009) that the re-advance occurred around 15,820 cal BP. Further exposure dating, also carried out by Ballantyne et al (2009) dated the moraine linked to the Red Point moraine at the Northwest extremity of Baosbheinn as having a mean exposure age of 15,690 cal BP. For moraines in the northerly Assynt region, Bradwell et al, (2008) recorded an exposure age of 15,970 cal BP, statistically similar to the Red Point exposure age recorded by Ballantyne et al (2009).

The interruption of the retreat of the Late Devensian ice sheet is not supported by any evidence of climatic cooling (Coope, 1975). This has created much debate over the years about the nature, extent and pattern of the Wester Ross Re-advance. Many authors believe that the Lateglacial Interstadial was ice free in Northwest Scotland (Sissons, 1967, 1977a; Lowe et al, 1994). On the other hand, cosmogenic exposure ages from moraine ridge boulders on the Wester Ross Re-advance moraine complex (McCormack, 2011; Bradwell et al, 2008; Ballantyne et al, 2009) indicate that ice persisted within Wester Ross between 15,000 and 16,000 years ago.

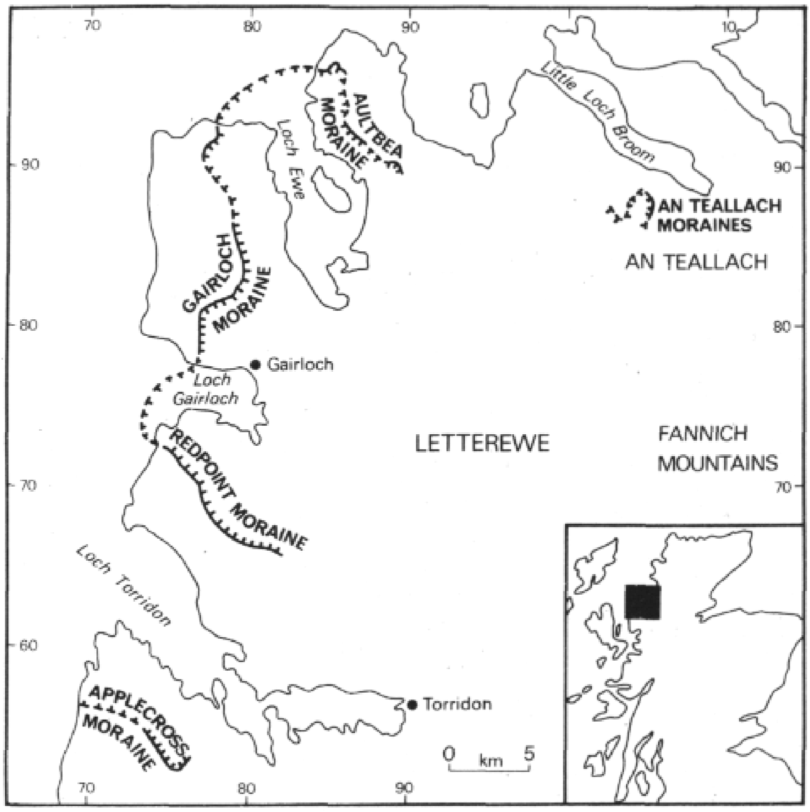


Figure 4. Former ice margins, within Wester Ross, other than those of the Loch Lomond Re-advance are shown by the serrated line. The glacier ice lay on the serrated side of the line. The dashed serrated line represents interpolated ice margin positions (Robinson and Ballantyne, 1979).

Sissons and Dawson (1981) argued on glaciological grounds that a re-advance is more plausible than a standstill. On either side of the Red Point moraine, changes in orientation of striae indicate a re-advance (Robinson and Ballantyne, 1979); however it is as yet unconfirmed by stratigraphic evidence and is of unknown magnitude. Although, the presence of an end moraine demonstrates active retreat of the last ice sheet in this area.

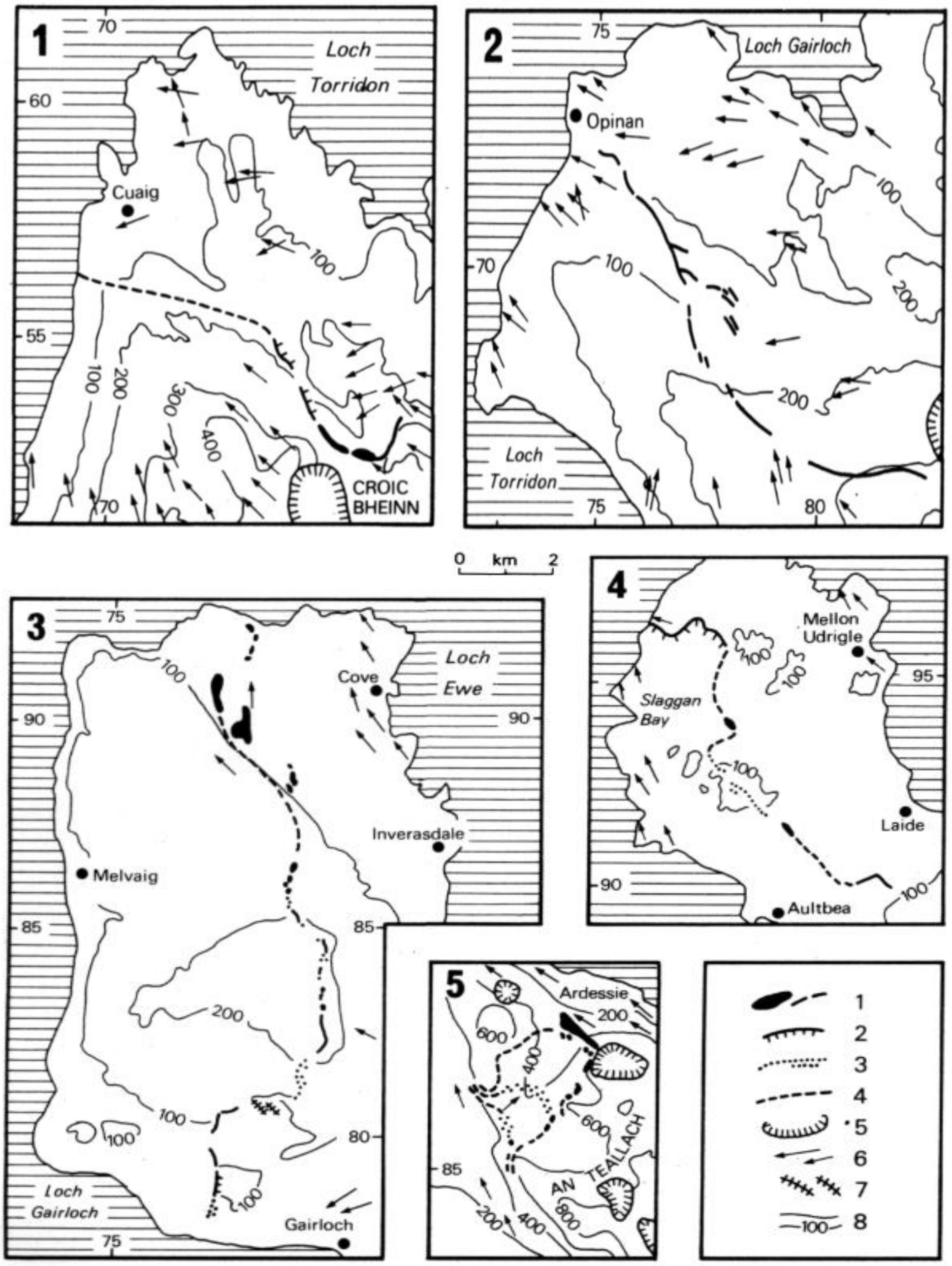


Figure 5.1 – 5.5. 5.1 Applecross Moraine; 5.2 Redpoint Moraine; 5.3 Gairloch Moraine; 5.4 Aultbea Moraine; 5.5 An Teallach Moarine. **Key:** 1 – Moraine ridges; 2 – Drift limit; 3 – Belt of boulders; 4 – Interpolated ice margin; 5 – Loch Lomond Advance galciers; 6 – Striae; 7 – Esker; 8 – Contours at 100m 0r 200m intervals (Robinson and Ballantyne, 1979).

2.1.2 The Windermere Interstadial (Bølling-Allerød) (15,000 – 13,000 cal BP)

The Lateglacial was first recognised in lake sediments in 1901, in Allerød, Denmark. Birch remains were found in the organic mud between arctic alpine flora remains in the upper and lower clay layers (Joosten, 1995). A return to cold conditions which interrupted the warm episode during deglaciation (Allerød) can be interpreted from this sequence. The introduction of pollen analysis in 1916 supported this Lateglacial sequence and subsequently the Jessen-Godwin pollen Zonation scheme for Northwest Europe was formed (Table 1).

The Windermere Interstadial which signifies the ending of the glacial period has been subdivided into three chronozones (Mangerud et al, 1974). Figure 6 shows all the calibrated dates and the different periods during the Interstadial. The earlier Bølling period (Ib in Jessen-Godwin pollen zones (Godwin, 1975), with rapid warming and a thermal maximum radiocarbon dated between c.15,000 – 14,000 cal BP, was dominated by open tundra and dwarfs shrubs. It is thought a rise of 9-10^{o}C in the mean July temperature would have brought temperatures up to a maximum of 17^{o}C (Barton, 1999; Atkinson et al, 1987). Although, the moderate oceanicity of the climate created more continental style conditions as winter temperatures stayed low despite the rapid rise in summer temperatures (Coope et al, 1977). The later phase dated from c.13,800 – 13,000 cal BP, Allerød (II), is characterised by warm conditions although cooler than the Bølling, with park tundra and extensive birch woodland. These two chronozones are separated by the Older Dryas (Ic) between c. 14,000 – 13,800 cal BP. The Older Dryas is a colder period interrupting a general trend of cooler conditions; this can be seen in the pollen record by a short lived decline in dwarf shrubs. The pollen record during zone II suggests that the Lateglacial climate optimum occurred during the Allerød chronozone, however fossil Coleoptera (beetles) have been used to create temperature reconstructions which indicate that the Lateglacial thermal maximum actually occurred at 15,000 cal BP and that a cooling trend was in fact underway by c.14,700 cal BP (Bishop and Coope, 1977; Coope, 1977; Coope and Joachim, 1980; Atkinson et al, 1987).

2.1.3 The Loch Lomond Stadial (Younger Dryas) and Lateglacial Holocene Transition (13,000 – 11,700 cal BP)

The Loch Lomond Stadial signifies the final phase of substantial glacial conditions to affect Great Britain at the end of the Pleistocene, before the abrupt and rapid warming that characterised the beginning of the Holocene. The Loch Lomond Stadial is dated between 13,000 and 11,700 cal BP (Mangerud et al, 1974) and immediately follows an episode of increased warmth during the Lateglacial Interstadial. During the Lateglacial Interstadial rapid deglaciation meant the release of freshwater into the North Atlantic through the melting of ice sheets at high latitudes. In particular, the deglaciation of the Laurentide Ice Sheet and the discharge of fresh water from Lake Agassiz (Broecker, 2006) along with an increase in melt water from around the North Atlantic could well have triggered a large-scale change in the North Atlantic Thermohaline Circulation (Anderson, 1997). The resultant decrease in northward heat transport produced a significant near glacial episode which is known as the 'Loch Lomond Stadial' or 'Younger Dryas' (zone III (Godwin, 1975)) as it is known in continental Europe. The extent of the Loch Lomond Stadial is becoming more evident; it is now suggested that it was hemispheric and global (Lowell et al, 1995; Peteet, 1995). There is a strong correlation between a number of glacial re-advances and the cool conditions of the Stadial. In Scotland, the Stadial was characterised by an 8-10^{o}C drop in mean annual temperature (Hubbard, 1999; Isarin and Renssen, 1999) and marks a shift back to open tundra along with an increase in arctic and alpine taxa, such as *Salix herbacea* and *Artemsia*, which are more adapted to the colder conditions and a decrease in dwarf shrub heaths (Table 1).

2.1.4 Anthropogenic Activity during the Late Upper Palaeolithic and Evidence for Ahrensburgian Technology

In Wester Ross, evidence has been found which suggests human activity in the late upper Palaeolithic. The finding of numerous tanged points provides the main evidence to suggest human activity in the area (Edwards and Mithen, 1995; Finlayson and Edwards, 1997; Livens, 1956; Mercer, 1980; Mithen, 2000; Morrison and Bonsall, 1989; Wickham-Jones and Firth, 2000). These findings are further supported by pollen and charcoal sequences (Edwards and Mithen, 1995), faunal remains (Lawson and Bonsall, 1986) and bone and antler artefacts (Morrison and Bonsall, 1989). The majority of tanged points found in Scotland are stray finds with no clear place of origin or little known history. Mainland European and north German Ahrensburgian group tanged points have been used as a comparison to try and identify the provenance of the tanged points found; a point found at Shieldaig in Wester Ross has many similarities to the Ahrensburgian points (Ballin and Saville, 2003). A site was excavated near Shieldaig, Loch Torridon, Wester Ross, by Michael Walker in 1973. Quartz, bloodstone and flint made up some of the 6,000 artefacts discovered at the site. Many microliths were found within the predominantly Mesolithic collection, along with two Neolithic arrowheads and one tanged point. The earliest clear indicator of human presence in Scotland comes from the dating of a carbonized hazel nut shell found with a narrow-blade microlith assemblage at Cramond, Edinburgh (Ashmore, 2001; Denison, 2001). The shell was dated between 10,425 cal BP and 10,250 cal BP.

During the end of the Upper Palaeolithic period higher ground was ice covered, tundra vegetation was present in only the most adaptable areas, terrestrial fauna was greatly reduced and polar air temperature and water conditions prevailed in the Northwest of Scotland (Lowe and Walker, 1986; Peacock and Harkness, 1990). Temperatures during the winter were on average 20-30°C lower than present temperatures (Isarin et al, 1998). Cold adapted mammals could possibly have inhabited these areas, perhaps only in the summer months (Ballin and Saville, 2003). However, it is very improbable that there was any human presence this far north in Scotland during the time these severe periglacial conditions prevailed (Tolan-Smith, 1998). Barton (1997) suggested that any late glacial hunters present at this time may well have left Britain completely. The subsequent climatic improvement which followed the Loch Lomond Stadial left a period of several hundred years when any hunters in Northwest Scotland demonstrated Ahrensburgian technology (Ballin and Saville, 2003).

Late Glacial Chronozones (Mangerud et al, 1974)	Estimated Ages ^{14}C yr BP	Jessen-Godwin Pollen Zones (Godwin, 1975)	Vegetational Characteristics	Climatic Characteristics
Younger Dryas	11,000-10,000	III	Open tundra, increase in arctic/alpine taxa including *Salix herbaceae* and *Artemsia*. Decrease in arboreal pollen, increase in cold adapted shrub and herb taxa.	Return to glacial conditions.
Allerød	11,800-11,000	II	Shift from open-park tundra communities, birch established throughout North West Europe.	Warm conditions but cooler than Bølling.
Older Dryas	12,000-11,800	Ic	Brief decline in dwarf shrubs including *Juniper*.	Phase of colder conditions during general cooling trend.
Bølling	13,000-12,000	Ib	Open tundra and some dwarf shrubs following deglaciation.	Rapid warming and thermal expansion.

Table 1. Lateglacial Chronozones (Mangerud et al, 1974) and Jessen-Godwin Pollen Zones (Godwin, 1975).

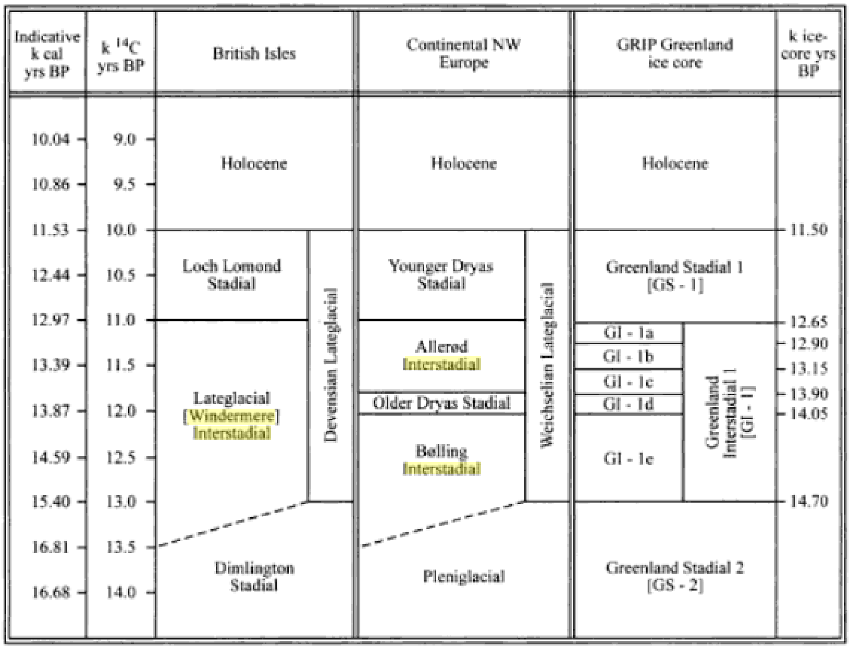


Figure 6. Subdivisions of the Lateglacial in the British Isles, Continental Northwest Europe and Greenland. The dates are based on ^{14}C and the Greenland ice cores. The ages are calibrated according to Reimer et al (2004) (Brenchley, 2006).

2.2 Early Holocene in Scotland (11,700-6000 cal BP)

2.2.1 Vegetation and Climate

The warming climate of the early Holocene saw a reorganisation of the composition and distribution of vegetation throughout Scotland. Two significant shifts in vegetation occurred during this period: the expansion of birch-hazel woodland and its subsequent replacement by *Pinus sylvestris*.

Birks (1984) reconstructed the vegetation record from at An Druim, near Eriboll, in the extreme Northwest of Scotland. *Juniperus* and *Empetrum* and other dwarf shrub heaths signified the beginning of the Holocene. There were high levels of *Betula* recorded as tree birch expanded rapidly as temperatures rose exceeding birch requirement of 10^{o}C annual temperature (Birks, 1984). The rapid expansion of *Corylus avellana* about 9800 cal BP formed mixed birch-hazel woodland that is characteristic of many Scottish pollen sites within the early Holocene. Low levels of *Ulmus* and *Salix* persisted. However, in contrast to other pollen sequences produced from north and Northwest Scotland, at An Druim, there was no significant rise in pine recorded; well no rise sufficient enough to meet the 20% Total Land Pollen threshold set out by Bennett (1984) to indicate local pine growth. However, within the Eriboll area pine stumps do exist fuelling the debate about the absence of pine pollen and presence of pine macrofossils.

Birks (1972b) also collected pollen data from Loch Maree, Northwest of Kinlochewe in Wester Ross. The record produced from Loch Maree shows a *Juniperus communis* dominated landscape around 10,800 cal BP, with lower percentages of *Empetrum* and herbs, such as *Rumex acetosa*. This vegetation was soon replaced by *Betula* at 10,200 cal BP, with small levels of *Corylus* present. Low but persistent levels of *Ulmus* and *Quercus* began to appear around 9400 cal BP, however in comparison to the data collected from An Druim the most significant aspect of this record is the early and rapid development of *Pinus sylvestris* radiocarbon dated to between 8900 and 8250BP; the earliest known incidences of pine in Scotland (Birks, 1972b, 1986). Loch Maree is of particular interest as it has large populations of *Pinus sylvestris* around its shores, a taxon that is very rare today in Northwest Scotland (Ratcliffe, 1977). The isoenzyme loci and monoterpene data collected from these pines indicate that they have “little genetic affinity between contemporary Scottish and continental European populations” (Kinloch et al, 1986). The record from Loch Maree supports the data produced by Kinloch et al (1986) suggesting the population of *Pinus sylvestris* around Loch Maree was and is specific and unique to this area. The origin of this premature influx of *Pinus* is still a mystery (Kinloch et al, 1986; Birks, 1989). Although, pine populations were present in areas of central and southern England as well as southern Ireland at the time. Birks (1989) proposes long-distance seed dispersal or glacial survival as the most likely hypothesis for this early occurrence of pine. By 8000 cal BP, pine dominated in widespread open woodland with high percentages of *Pteridium aquilinum* and *Calluna vulgaris* (Birks, 1972b, 1986).

The dates given for the rise in pine pollen throughout Wester Ross and northern Scotland appear to be irregular and asynchronous indicating that the timing of the pine expansion varied not only regional but also more locally (Anderson, 1996).

Pennington et al (1972) gave a date of 7424 cal BP for the pine rise at Loch Clair, near Kinlochewe in Wester Ross while they also dated the rise at Loch Sionascaig in Sutherland at 8707 cal BP. Anderson (1996, 1998) dated two cores from Glen Torridon and Glen Carron in Northwest Scotland (figure 7 show sites mentioned in the text). The pine rise at Glen Torridon was dated at 7490 cal BP, while the rise at Glen Carron was dated at 7580 cal BP. Despite the inconsistency in the dates of the pine rise, the records as a whole suggest that the majority of Wester Ross was densely forested by *Pinus sylvestris* by about 7000 thousand years ago (Anderson, 1996). Anderson (1998) proposed that the fairly dry bog conditions indicated from C:N (Carbon: Nitrogen) and light transmittance data for the early to mid-Holocene were responsible for this pine expansion. The continued warming trend throughout the early Holocene led to the Holocene thermal maximum and the dry and warm conditions necessary for pine growth (Anderson, 1998). A reduction in waterlogged conditions encouraged pine growth and expansion (Bell and Tallis, 1974).

However, Tipping et al (2007) suggests that at Loch Farlary, the rise in *Pinus sylvestris* was independent from any hydrological fluctuations. The expansion of *Pinus sylvestris* around the Loch Farlary occurred fairly late at 7780-7720 cal BP, in comparison to the expansion throughout the rest of the region of 9500 – 8500 cal BP (Tipping et al, 2007). The topographic limits of the glens surrounding Loch Farlary may well have controlled the spread of the taxon (Bennett, 1986). Competition from *Corylus* may have confined *Pinus sylvestris* to the peaty, damper and more acidic soils while *Corylus* occupied the better soils (Bennett, 1984; Davies, 1999), establishing a fairly open community landscape with its successful competitors, *Betula* and *Calluna*. By 7500 cal BP the record supports immediate local growth of pine at Farlary with growth after 7600 cal BP encouraged by dry soil conditions (Tipping et al, 2007). The time of the colonisation of pine, however was independent from any precipitation variation as the surface of peat at Farlary had a significantly reduced level of wetness for the past 500-600 cal years previous to the pine colonisation. It is most likely that the timing of when pine first reached the east was determined by the earlier establishment of pine woodland on the west coast and the distance from seed source (Tipping et al, 2007).

It has been suggested that the colonisation and establishment of Scots pine woodlands was encouraged by an intensification in the fire regime. Indeed, the removal of competitors through burning would allow pine seedlings to establish and invade the newly open surface. Within the British Isles the only combustible native tree taxa is *Pinus sylvestris* (Rackham, 1986), with the taxa making up many communities which are fire-dependant (Agee, 1998). However, the debate as to whether these fires were caused by human interference or are of natural origin is ongoing. There is a distinct lack of data to suggest that any fires which promoted the expansion of pine during the early Holocene were due to anthropogenic activity such as the clearance of land through small scale burning of vegetation. Anderson (1998) argues that by 7000 years ago the majority of Wester Ross and Northern Scotland was covered in dense pine woodland. Any communities present in Scotland at this time would be unable to produce such a widespread and large scale event such as the pine rise. If fires did help establish the pine woodlands which covered much of Scotland from the early to the mid Holocene then they were most probably spontaneous wildfires due to the increased aridity caused by early Holocene warming or lightning strikes (Durno and McVean, 1959; McVean, 1963; Rackham, 1986). The suggestion that such fires were of natural origin is supported by Tipping (1994, 1996) and Tipping and Milburn (2000). They argue that the shift to wetter conditions between 6540-6200 cal BP coincided with the reduction in charred particles at 6400 cal BP; therefore, the arid conditions and the subsequent dry mire surface during the pine rise were likely to have been the contributing factor in the fire regime as opposed to anthropogenic activity.

Lageard et al (1991) use a threshold of 3% as an indicator of local pine growth, in comparison to Bennett's (1984) 20%. Many authors use a higher threshold percentage to indicate local pine growth due to the significant quantities of pine pollen produced by individual trees, and the promotion of wind dispersal by the morphology of the pollen grains (Huntley and Birks, 1983). While Gear and Huntley (1991) work with a 30% threshold for pine pollen, Tipping (1989) argue that the Awe Valley in Western Scotland and other open Lateglacial landscapes will have a much higher pine pollen percentage through long-distance dispersal – in some cases as high as 40%. Bennett's theory of 20% TLP has since been reworked – an investigation done by Bennett in 1995 used 5% TLP as an indicator of local pine growth. This change of 15% came after Fossitt (1994) presented evidence to suggest that less than 5% of pine pollen was found in lake sediments along with pine stomata. Dendrochronological analyses done by Brown (1991) in Northern Ireland came to a similar conclusion. The analyses showed that pine was present to the South of Garry Bog, however there was only 5% TLP recorded for pine pollen at the same bog. Lageard et al (1991) demonstrate that there are substantial spatial variations (<50m) in the representation of pine pollen. There are a number of reasons that could be responsible for these variations such as, limited or retarded productivity of pollen, filtering impacts of local vegetation or defective tree growth due to differences in local hydrology. This study carried out at Meall Bad á Chrotha uses Bennett's 1984 threshold of 20% as an indicator of local pine growth, as it lies between the thresholds put forward by other authors, which range from 5% (Brown, 1991; Bennett, 1995, Fossitt, 1994) to 40% TLP (Tipping, 1989). Meall Bad á Chrotha is relatively open, therefore, lends itself to long-distance pollen dispersal, so a slightly higher percentage for pine pollen representation would be more reliable. Furthermore, dendrochronological or pine stomata analyses is not being carried out in this investigation, therefore, a higher threshold would again ensure reliability.

Figure 7. Map showing the main sites mentioned in the text: **1)** Loch Maree and Eilean Subhainn (Birks, 1972, 1996; Kerslake, 1982); **2)** Loch Sionascaig (Pennington et al, 1972); **3)** Badentarbet (Lamb, 1964); **4)** Loch Farlary (Tipping et al, 2008); **5)** Glen Carron (Anderson, 1996, 1998); **6)** Glen Torridon (Anderson, 1996, 1998); **7)** Loch Clair (Pennington et al, 1972); **8)** Isle of Skye (Selby, 2003, 2004); **9)** Reidchalmai (Tipping and McCulloch, 2003); **10)** Sand (Hardy and Wickham-Jones, 2003); Blue Circle represents Meall Bad à Chrotha.

2.2.2 The origins of *Pinus sylvestris* in Northwest Scotland

A large proportion of our knowledge about the invasion of tree taxa into Britain after glaciation comes from pollen records (Birks, 1989). In Britain, the natural populations of Scots Pine are restricted to the Scottish highlands. Out of all the conifers *Pinus sylvestris* is the most broadly distributed, while the populations in Scotland exhibit the outlier of the taxon (Vidakovic, 1991); 500km, at minimum, lie between the main distributions in Scandinavia and continental Europe, and the outlier of the taxon in the extreme Northwest of Scotland. *Pinus sylvestris* is comparable to other European tree taxa in regards to its expansion northwards from southern refugial populations after ice sheet retreat (Soranzo et al, 2000). The Alps, the Balkans and the Iberian Peninsula were localities with refugial populations, with the Balkans being responsible for the European recolonization of *Pinus sylvestris* (Huntley and Birks, 1983; Bennett et al, 1991). Ennos et al (1997) suggests that pine migrated through central Europe from the Balkans to southern Britain at 11,500 cal BP, with a second episode of migration which saw the recolonization of western Scotland around 8900 cal BP.

The origin of Scots Pine in the British Isles still encounters much debate. Reconstructing the paths and characteristics of spread is particularly problematic when it comes to *Pinus sylvestris* as this taxa has a tendency to deposit large amounts of pollen tens of kilometres away via the wind, therefore using pollen analysis as a sole way of reconstructing tree spread is unreliable. In the majority of origin studies, an extra dimension and set of information is added alongside pollen analysis through the use of genetic variation. Using present day populations of tree taxa such as *Pinus sylvestris*, the distribution of genetic variation can be determined (Sinclair et al, 1998).

From 20 natural populations in Scotland, 466 pine trees were sampled by Sinclair et al (1998) in order to determine their maternally inherited mitochondrial DNA (table 2). Cloned mtDNA genes were used as probes to identify three RFLP variants; two common (**a** and **b**) and one rare which was found at Glen Falloch (**c**) (figure 8). Mitotype **a** is existent in all Scottish populations of *Pinus sylvestris* however, three populations in the west are the only ones that exhibit mitotype **b**. Genetic differentiation for mtDNA, which migrates exclusively by seed is far greater ($F_{ST(m)}$ = 0.370) than for nuclear markers ($F_{ST(b)}$ = 0.028) which are dispersed by both seed and pollen (Sinclair et al, 1997). The presence of Mitotype **b** in the west of Scotland, and its absence in northern France and Germany indicates that *Pinus sylvestris* originated from both a single refugia in the west, most likely in western France or Ireland and from continental Europe via England. This suggests that populations of scots pine in Scotland today, originated from more than one refugium after glaciation (Sinclair et al, 1998). However, the *Pinus sylvestris* populations in Ireland cannot be sampled due to their extinction between 1000 and 2000 years ago, therefore it is impossible to correlate mitotype **b** to sources in Ireland (Birks, 1989). Nonetheless, it is still highly plausible that scots pine did originate from both continental Europe and a western refugium as the isozyme nuclear markers and the monoterpene data show a distinct genetic difference in the populations of the Northwest, in particular the population at Shieldaig, Wester Ross (Kinloch et al, 1986).

Population and number	Location	Elevation (m)	Number of each mitotype a	b	Total
1. Glen Einig	57°57′N, 4°46′W	100	13	0	13
2. Rhidorroch	57°54′N, 4°58′W	100	26	0	26
3. Strath Vaich	57°45′N, 4°47′W	300	13	0	13
4. Loch Maree	57°40′N, 5°25′W	20	4	0	4
5. Shieldaig	57°31′N, 5°37′W	100	38	7	45
6. Achnashellach	57°28′N, 5°17′W	150	25	0	25
7. Glen Strathfarrar	57°24′N, 4°43′W	150	30	0	30
8. Glen Affric	57°15′N, 5°04′W	250	26	0	26
9. Loch Hourn	57°10′N, 5°28′W	50	17	0	17
10. Glen Barisdale	57°04′N, 5°30′W	150	30	0	30
11. Glen Garry	57°04′N, 4°55′W	150	13	0	13
12. Glen Loy	56°55′N, 5°08′W	150	13	19	32
13. Conaglen	56°48′N, 5°21′W	200	16	0	16
14. Rannoch	56°41′N, 4°19′W	250	20	0	20
15. Doire Darach	56°32′N, 4°47′W	200	37	3	40
16. Glen Orchy	56°27′N, 4°53′W	150	24	0	24
17. Glen Falloch	56°22′N, 4°39′W	200	29	0	29*
18. Abernethy	57°14′N, 3°39′W	300	10	0	10
19. Ryvoan	57°10′N, 3°42′W	400	29	0	29
20. Glentanar	57°01′N, 2°53′W	250	24	0	24

*One additional individual of mitotype **c** found

Table 2. Populations of *Pinus sylvestris* in Scotland analysed for mtDNA variation, and numbers of each RFLP mitotype scored (Sinclair et al, 1998).

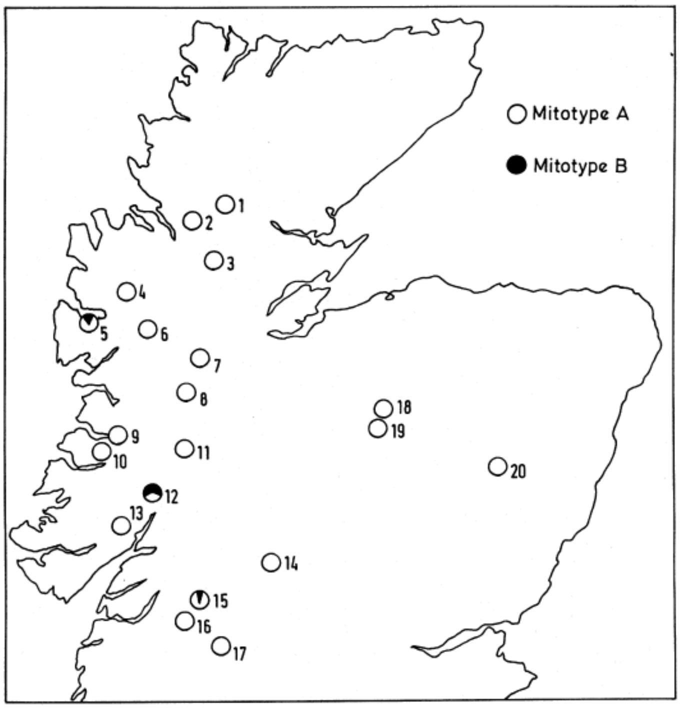


Figure 8. Distribution of mitotypes within 20 natural populations of *Pinus sylvestris* in Scotland (Steven and Carlisle, 1959; Sinclair et al, 1997). Table 2 corresponds to the numbers (Sinclair et al, 1997).

Further studies have indicated that populations in Wester Ross have an individual origin and distinctly different genetic markers that separate them from other native populations at the time of colonization; further supporting the hypothesis that the populations in the west originated from a western refugium. Forrest (1980) also provides evidence for the multiple origins of *Pinus sylvestris* during the post glacial. In Wester Ross, a collection of native pine populations showed a markedly lower amount of 3-carene in their cortical resin than those recorded anywhere else in Scotland, after the composition of their genetically determined resin monoterpene was analysed (Forrest, 1980). Kinloch et al (1986) produced isozyme variation data to provide evidence that a pine population at Shieldaig, Wester Ross, has the highest level of genetic differences when compared to other populations. One of the most significant problems with establishing the history of present day pine distributions is that interpopulation gene flow leads to the breakdown of genetic structure, however the population at Shieldaig still exhibited distinctive genetic markers as many as 50 generations later (Sinclair et al, 1998).

However, Soranzo et al (2000) has provided evidence which contradicts Sinclair et al's (1998) hypothesis that there are two separate gene pools within the *Pinus sylvestris* populations in Scotland. 747 trees were analysed from 23 populations throughout Europe using mitochondrial markers (*nad* 1, exon B/C) (table 3). This analysis revealed the existence of two haplotypes within Europe (table 4). One haplotype (*a*) was present in continental Europe and Scotland whereas both haplotypes *a* and *b* were present on the Iberian Peninsula; in Spain the *a* haplotype was present in the south and west and *b* was present in the Northeast region (figure 9). The *Pinus sylvestris* populations present on the Iberian Peninsula represent ancient gene pools as during the Quaternary period Spain stayed outside of the southern range of ice (Denton and Hughes, 1981), therefore pine remained throughout the glacial period (Huntley and Birks, 1983). There is one exception found in the data produced by Soranzo et al (2000) - one individual tree in Kubryk, Poland which was characterized by haplotype *b*. There have been three possible hypotheses put forward to explain the presence of haplotype *b* in Poland. It is possible that this variant does exist in extremely minor values throughout Europe. It is also possible that this *b* haplotype has been transported and deposited via long distance dispersal of cones or seeds from a *b* haplotype locality. Human activity cannot be ruled out as a possible cause for the presence of this individual *b* haplotype. Recently, the movement of plant material around Europe (Ennos et al, 1998) may well have impacted on the genetic diversity at the time when certain pine populations were established (Soranzo et al, 2000). Further testing will need to be carried out in order to determine the cause of the *b* haplotype in the Polish population. Soranzo et al (2000) tested three scots pine populations in Wester Ross, Coulin, Loch Maree and Shieldaig. No evidence was found to support Sinclair et al (1998) data that there were two definitive gene pools in Wester Ross. The contrasts between the underlying mutational process affecting the evolution of mitochondrial genome and their method of detection using different molecular assays may be the cause for the differing results. Furthermore, the presence of only one haplotype in western Scotland from Soranzo et al's (2000) study may be due to the sampling of only one genomic region.

Population	Country	Site key	Latitude	Longitude	Elevation (m a.s.l)	*n*
Rannoch	Scotland	RAN	56°41′ N	4°19′ W	230	47
Shieldaig	Scotland	SHI	57°31′ N	5°37′ W	15	47
Coulin	Scotland	COU	57°32′ N	5°21′ W	120	47
Loch Maree	Scotland	MAR	57°40′ N	5°25′ W	25	47
Glen Einig	Scotland	EIN	57°57′ N	4°46′ W	100	47
Glenmore	Scotland	GLE	57°10′ N	3°42′ W	265	47
Haguenau	France	HAG	48°50′ N	7°52′ W	Unknown	24
Kubryk	Poland	KUB	51°25′ N	17°20′ W	Unknown	24
Silene	Lithuania	SIL	55°45′ N	26°40′ W	Unknown	24
Spitzberg	France	SPI	48°36′ N	7°13′ W	Unknown	24
Zahorie	Czech Rep.	ZAH	48°48′ N	17°00′ W	Unknown	24
Puebla de Lillo	Spain	LIL	43°04′ N	5°15′ W	1550	30
San Zadornil	Spain	ZAD	42°50′ N	3°11′ W	1000	30
Borau	Spain	BOR	42°42′ N	0°35′ W	1550	29
Pobla de Lillet	Spain	POP	42°14′ N	1°58′ W	1100	30
Covaleda	Spain	COV	41°56′ N	2°48′ W	1550	27
Galve de Sorbe	Spain	GAL	41°15′ N	3°07′ W	1400	30
Valsain	Spain	VAL	40°49′ N	4°01′ W	1550	30
Navarredonde de Gredos	Spain	NAV	40°21′ N	5°07′ W	1550	30
Orihuela del Tremedal	Spain	ORI	40°31′ N	1°38′ W	1750	30
Gúdar	Spain	GUD	40°25′ N	0°41′ W	1700	30
La Cenia	Spain	CEN	40°45′ N	0°03′ E	1100	30
Baza	Spain	BAZ	37°22′ N	2°51′ W	2050	30

Table 3. Populations of *Pinus sylvestris* studied by Soranzo et al (2000).

Population	Total	a		b		Gene diversity ($\hat{H}$)
		n	Freq.	n	Freq.	
Rannoch	34	34	1.000	—	—	—
Shieldaig	46	46	1.000	—	—	—
Coulin	46	46	1.000	—	—	—
Loch Maree	47	47	1.000	—	—	—
Glen Einig	41	41	1.000	—	—	—
Glenmore	47	47	1.000	—	—	—
Haguenau	20	20	1.000	—	—	—
Kubryk	25	24	0.960	1	0.040	0.080
Silene	23	23	1.000	—	—	—
Spitzberg	24	24	1.000	—	—	—
Zahorie	15	15	1.000	—	—	—
Puebla de Lillo	30	30	1.000	—	—	—
San Zadornil	30	16	0.533	14	0.467	0.515
Borau	29	—	—	29	1.000	—
Pobla de Lillet	27	1	0.035	26	0.966	0.069
Covaleda	27	4	0.148	23	0.852	0.262
Galve de Sorbe	30	—	—	30	1.000	—
Valsain	29	24	0.828	5	0.172	0.296
Navarredonde de Gredos	30	17	0.567	13	0.433	0.508
Orihuela del Tremedal	30	11	0.367	19	0.633	0.481
Gúdar	30	1	0.033	29	0.967	0.067
La Cenia	30	—	—	30	1.000	—
Baza	30	27	0.900	3	0.100	0.186

Table 4. Haplotypes detected in *Pinus sylvestris* populations and diversity values (Soranzo et al, 2000).

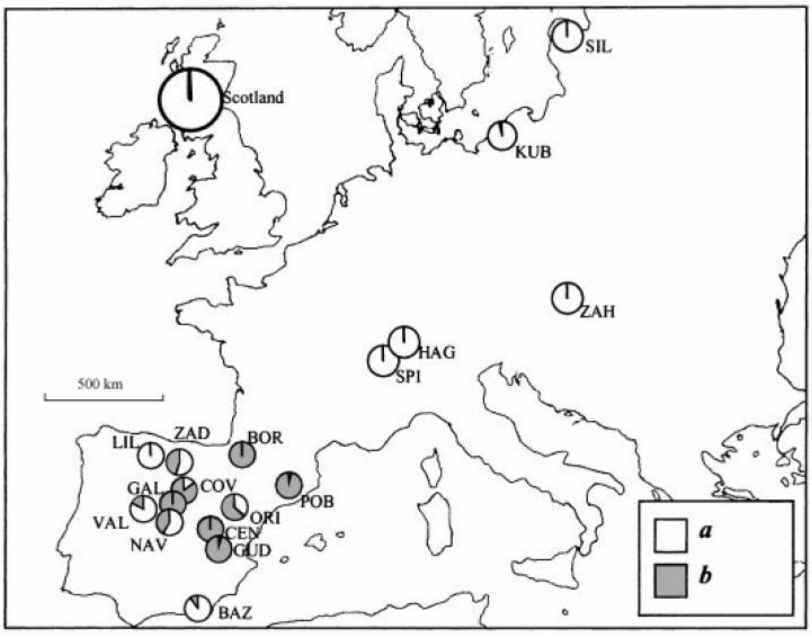


Figure 9. The distribution of Haplotypes in *Pinus sylvestris* populations in Europe (the site keys can be seen in table 3) (Soranzo et al, 2000).

The data produced suggests two origins for scots pine in Scotland, although much debate continues over the existence of a western refugium. The suggestion that the population in the west was endemic is extremely improbable when the glacial history of Scotland is taken into account (Kinloch et al, 1986). It is more conceivable that the populations in Scotland originated from southwest Ireland or western France considering the extent of the ice sheet covering Britain and the likelihood of pine surviving glaciation is very low (Ballantyne and Harris, 1994; Bennett, 1995).

2.2.3 The Mesolithic

Hardy and Wickham-Jones (2003) identified a number of sites within the Inner Sound that were occupied by very early Scottish settlers, some of Mesolithic age with one midden being dated to the eighth millennium BP. The excavation of a midden at Sand on the Applecross Peninsula has allowed the reconstruction of many aspects of everyday life of Mesolithic communities. A narrow blade microlith as well as various pointed and bevel ended bone and stone tools were recovered. The variety of artefacts found and the deposition of non-midden material, such as microliths, suggest that there was a variety of activities going on at Sand other than the activities involving the midden (Hardy and Wickham-Jones, 2003). However, it is evident from the findings that the sea and the coastline were essential to the everyday lives of the communities present at Sand during the early Holocene. The proportion of shellfish found suggest that they were a central food source, although there were fish and animal bones also present in the midden suggesting that there was ample resource rich bases to exploit. Although, there is the possibility that shellfish were used as bait as well as a food source (Hardy and Wickham-Jones, 2003). The Inner Sound itself provides an array of resource bases, with easy access to the sea and land. The finds at Sand also provide evidence for the exploitation of nearby woodlands; hazel nut shells that have been subject to heat not only suggest that the woodlands were used as another food source but that there was some sort of fire regime occurring at the time. Fire cracked stones, pot boilers and charcoal fragments were found. However, the Sand midden seems to have accumulated over a very short period of time, perhaps only a couple of seasons (Hardy and Wickham-Jones, 2003). If the occupation of a midden was typically only for a short period of time then it would be assumed that Mesolithic middens would be abundant throughout the peninsula and the west coast; however this is not the case. Hardy and Wickham-Jones (2003) have put forward the argument that the midden at Sand was abandoned due to a change of conditions. A period of famine has been suggested as the adverse conditions which caused these communities to leave. Edwards (1990) has suggested that this site could well date to the climatic downturn at 8200 cal BP; indeed, a climatic downturn could well lead to a period of famine (Hardy and Wickham-Jones, 2003). In support of this, the reconstruction of paleoshorelines has dated a rise in sea levels to around this time, with water rising to just beneath the rockshelter in which in the midden is situated. The climatic event at 8200 cal BP was a result of changes in the circulation of the North Atlantic, this would have had a detrimental effect on the resource bases used by such communities. Fish migration, coastline vegetation and land based resources would have all been affected (Cushing, 1982). Gonzalez-Samperiz et al (2009) suggest that migration from these affected areas is more likely than population collapse. However, Wicks and Mithen (2014) argue that this climatic event could have directly influenced the birth and death rates within the already low Mesolithic population initiating their collapse in western Scotland. They support their argument by suggesting that the Mesolithic population would have been unable to adapt their tools, technology and overall lifestyle to the differing environmental conditions, which ultimately led to these lower reproduction and higher mortality rates.

The possibility that there are many more Mesolithic sites that remain undiscovered is very high. In a number of cases, Hardy and Wickham-Jones (2003) found that rockfall had rendered a large proportion of the rockshelters unidentifiable as a complete examination of the site is difficult. There is also the problem that many of the sites particularly, along coastal areas, have been destroyed through farming practices and building.

2.2.3.1 The Mesolithic to Neolithic Transition

The Mesolithic to Neolithic transition in Britain occurred roughly between 6400 and 6000 cal BP (Woodbridge et al, 2012). Throughout the British Isles this period was characterised by a wide range of central economic and environmental changes

(Cayless and Tipping, 2002). A substantial growth in population led to the intensification of forest and vegetation interference, expansion of both pastoral and arable farming and variations in the fire regime. The emergence of farming and the cultivation of cereal crops was one of the first aspects of Neolithic culture to arrive in Northwest Europe around 6000 cal BP (Woodbridge et al, 2012). There is a continuing debate as to how long the Neolithic culture took to spread through the British Isles. Whittington et al (2011) argue that the Mesolithic to Neolithic transition was a very time consuming event due to its regionalised nature, first emerging in the southeast of England before taking over two-hundred years to develop and expand into other regions. On the other hand, Bocquet-Appel et al (2012) proposed that in comparison to the rest of Europe the arrival and expansion of farming activities in Britain was very rapid, so rapid in fact that Collard et al (2010) suggest that sea routes as well as overland spread must have played a major role. The view that the transition was a fairly discrete and short-lived event comes from Bonsall et al (2002), who argued that the transition occurred around 6100-5800 cal BP, over a half a millennium later than the prolonged cultural and economic changes suggested by some authors.

Between 6100 – 5800 cal BP the implementation of farming throughout the British Isles could well have been due to a change to an increasingly continental like climate, with elevated summer temperatures and decreased winter precipitation, heavily influencing the rapid and widespread advancement of farming throughout the British Isles from more southern continental areas (Bonsall et al, 2002). This climatic improvement may have been a significant reason for the transition from Mesolithic to Neolithic in Northwest Europe as these more favourable conditions would have improved the once unsuitable land for cereal cultivation (Bonsall et al, 2002). On several sites throughout the British Isles, there are very early incidences of cereal-type pollen; however, evidence on the landscape for human activity or agriculture prior to c.5800 cal BP is non-existent with the first evidence that suggests land clearance for agriculture occurring at this time (Bonsall et al, 2002).

There have been two conflicting theories that have been put forward to explain the expansion of agriculture and Neolithic lifestyle throughout the British Isles and subsequently into Northwest Scotland and Wester Ross. Case (1969) and Bradley (1984) argued that the immigration of farmers from Europe and their colonization of the British Isles resulted in the introduction of the Neolithic period. The second theory is that the transition occurred through 'Neolithization', which is the adoption of new technologies, ideologies and resources by Mesolithic populations native to the British Isles from Neolithic farming populations in mainland Europe (Bonsall et al, 2002; Dennell, 1983; Kinnes, 1985; Williams, 1989; Thorpe, 1996; Whittle, 1999).

2.2.3.1.1 The Immigration Model

The immigration model is based on three central theories: the first is the seemingly sudden withdrawal of Mesolithic lifestyle and culture, and the quick appearance of new more continental artefacts, monuments and burial rituals; secondly, the coincident shifts in material and economic cultures; and thirdly, the absence of evidence for any long-term or permanent settlements throughout the transition (Case, 1969; Bradley, 1984).

Bonsall et al (2002) provide strong arguments which refute this model. The difficultly in establishing the precise region in which these supposed immigrants arrived into the British mainland has proved a sticking point with the immigration model. Furthermore, the sudden change in culture and the movement of population settlements cannot be assumed as an indication of immigrants moving into an area (Bonsall et al, 2002). On the other hand, the relocation of settlements can be seen as a result of the economic change that signifies the transition into the Neolithic period. The invention of new technology would have been necessary as Mesolithic tools, such a 'T' shaped axes produced from the antlers of red deer, which are a common feature of late Mesolithic Scotland, would have been inadequate for agricultural use (Bonsall et al, 2002). Along the west of Scotland and the rest of the British Isles the adoption of agriculture as a dominant food source seems to be immediate at the commencement of the Neolithic; this would mean that investing in the advancement and development of new technologies would need to have been rapid. Indeed, the replacement of these 'T' shaped axes, used by hunter-gatherers, by stone axes which are lot more suitable to agricultural activities such as woodland clearance and the creation of fences around plotted areas, occurred in the early Neolithic (Bonsall et al, 2002).

Another theory put forward to support the immigration model is also contested by Bonsall et al (2002). In only a few rare instances, evidence for both Mesolithic and Neolithic settlements has been collected from the same site. Wickham-Jones (1990) excavated a site in western Scotland at Kinloch which showed that the site had been occupied for a relatively extensive period covering both the late Mesolithic and the early Neolithic. The site was in a sheltered location with access to both fresh water and cultivable well-drained land. During the Mesolithic period, typically communities settled on sheltered sites along the coast as they relied heavily on the sea as a food resource (Johnson and Bonsall, 1999). The transition into the Neolithic period saw the abandonment of these sites as the reliance on the sea for resources moved to the land, where there was ample land for cultivation and grazing livestock (Bonsall et al, 2002).

2.2.3.1.2 The Neolithization Model

The Neolithization model has also came under intense criticism, although Bonsall et al (2002) conclude it is more plausible than the immigration model. Kinnes (1985) argues that the rapidity of the expansion of farming throughout the British Isles cannot be explained by immigration itself, furthermore there is no evident area from which certain aspects that characterise Neolithic life in Britain, in particular technology, originated on mainland Europe; suggesting that some Neolithic components, such as the leaf-shaped arrowhead, were developed domestically. On the other hand, there has been the argument that the Early Neolithic was to somewhat a continuity of the late Mesolithic with the population still relying heavily on hunting, gathering and fishing, rather than domesticated subsistence, such as cultivation and the rearing of livestock (Thomas, 1991: Armitt and Finlayson, 1992; Whittle, 1999). However, Bonsall et al (2002) argue that although wild resources may well still have been used there is no clear evidence to support the assumption that is was the central food base within the economy of the early Neolithic period in Britain. There is also a lack of sufficient evidence to suggest that there was a continuity of settlements during the transition. The large dwellings that would be associated with a prolonged period of settlement are scarce and the shell middens present on Scotland's west coast that have been suggested to show the continuity of settlement from the Mesolithic to the Neolithic period (Armitt and Finlayson, 1992; Thorpe, 1996) have been called into question. The middens, which were created through the deposition of unwanted food material, were likely to have been located a fair distance from the land the communities inhabited. Shellfish remains, such as limpets, are a central component in such middens that are found in such close proximity to the coast. Bonsall et al (2002) argue that if the shellfish populations were exploited over a sustained period in one particular location on the coast then they would quickly diminish; therefore, each settlement would most probably have a number of different middens at different points along the coastline where, once collected, the unwanted shells and fish bones were deposited and only the important meat carried back to the settlement (Bonsall et al, 2002). It is likely that a single midden would have been used on a regular basis, perhaps annually by a number of different communities. This strategy for the collection and processing of shellfish is a common aspect throughout the west coast of Scotland during the late Mesolithic period and perhaps into the Neolithic and Bronze Age periods, therefore, Bonsall et al (2002) argue that these middens cannot be assumed as evidence for the continuation of a particular settlement from the Mesolithic to the Neolithic period. On the other hand, shell middens are useful for the comparison of technology from one period to another. Articles that were used for collecting shellfish, such as bevel ended stone, antler and bone tools are often found in many of the shell middens. The dating of such tools from Scotland show that their use continued from the Mesolithic up until the Bronze Age (Bonsall and Smith, 1990; Bonsall et al, 1995). Bevel ended tools were very widely distributed throughout the British coastline and perhaps within Europe, and there is evidence that very similar varieties of the tool were present in North America; consequently the assumption that this tool indicates that there was settlement continuity at a particular site should be considered with caution (Bonsall et al, 2002).

2.3 Mid-Late Holocene in Scotland (6000 cal BP – Present)

2.3.1 Vegetation and Climate

At 6000 cal BP, *Pinus sylvestris* was still the dominant forest type throughout the majority of Scotland (figure 10). However, around 4000 cal BP *Pinus* collapsed to near extinct levels forming the basis of the modern day vegetation composition and distribution that we see in Scotland today. This widespread and rapid decline of Scots pine can be seen in numerous palynological records and by sub-fossil pinewood that has been preserved in peat throughout Northern Scotland (Birks, 1972, 1975; Bennett, 1984, 1995; Bridge et al, 1990; Gear and Huntley, 1991). The common hypothesis for this 'Pine Decline' is that a deterioration in climate increased waterlogging and soil paludification, pushed pine past its survival limits (H.H.Birks, 1972; Kerslake, 1982; H.J.B.Birks, 1977, 1989, 1990; Gear and Huntley, 1991).

The Loch Maree regional record shows a major decline in *Pinus sylvestris* dated to 4780 cal BP (Birks, 1972) and 4043 cal BP (Kerslake, 1982). At Glen Torridon and Glen Carron this sudden demise is dated at 4410 cal BP and 3830 cal BP (Anderson, 1998). The decline at Glen Torridon is coincident with declines recorded at Cross Lochs in eastern Sutherland and the widespread decline to the south and southwest (Birks, 1975; Pennington et al, 1972). A decline is also evident in the pollen diagram from Badentarbet on the Coigach Peninsula in Northwest Scotland at 4320 cal BP (figure 11). This date correlates with the dates for the pine decline from elsewhere in the Northwest (Bunting and Tipping, 2004; Pennington et al, 1972).

At Eilean Subhainn a two phase pine decline can be seen. The initial decline recorded was relatively early just before to a shift to drier bog conditions at c.4310 cal BP. During this phase, the pine pollen concentration fell from high (c. 60 x 10^3 grains per cm^3) to moderate (c. 20 x 10^3 grains per cm^3) (Anderson, 1998; Kerslake, 1982). The pine pollen concentrations fell to low during second phase of decline which was recorded about 220 years earlier than the wet shift at 3460 cal BP. This two phase decline can be interpreted in a number of records in Wester Ross.

Pennington et al (1972) witnessed a two phase decline at Loch Sionascaig, as did Kerslake (1982) at Lochan Dubh. A climatic deterioration around 5000 cal BP along with enhanced paludification of soil and the spread of blanket peat could well have signified the beginning of the end for *Pinus sylvestris* as it was pushed beyond its survival limits, particularly as it was already growing in marginal areas. Due to the longevity of *Pinus sylvestris* and its high tolerance of acidic and nutrient poor soils, a time lag between the shift in climatic conditions and a decline in pine pollen is expected. The variations and wide age range between the dates given for the declines in different catchments is understandable, as local environmental conditions, and the degree to which they deteriorate enough to push the vegetation past its survival limits, can vary (Anderson, 1998).

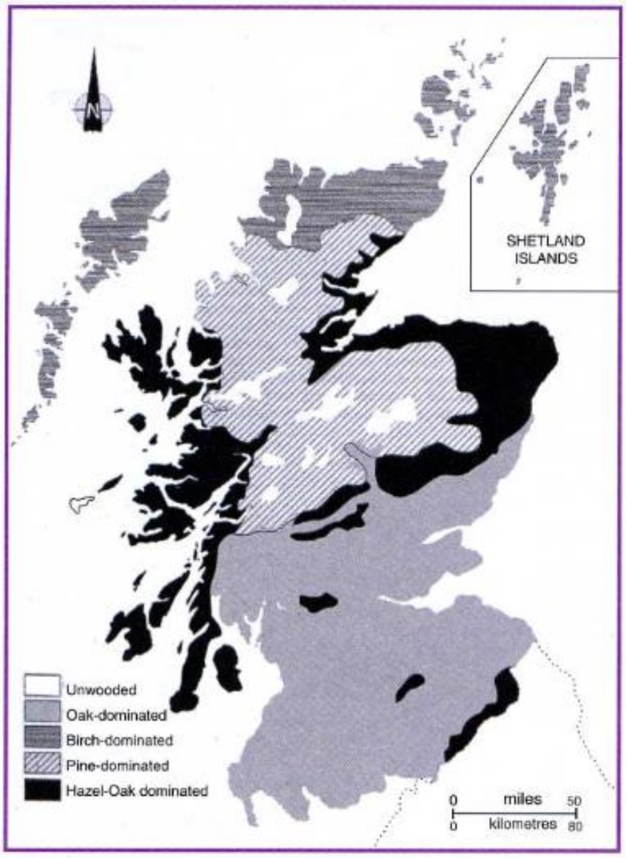


Figure 10. 6000 years ago pine was the most dominant forest type (Tipping, 1997).

A local shift to warmer and drier bog conditions after 4500 cal BP may well have halted or slowed the initial pine decline, particularly at Glen Carron and Eilean Subhainn (Anderson, 1998). The beginning of this dry phase is not characterised by a substantial rise in pine pollen however, suggesting that the dry shift was not the result of any re-expansion of pine woodland close to the bogs, which would have led to an increase in evapotranspiration rates and in turn a decrease in water table depth (Anderson, 1998).

Rather it suggests that the drier conditions were induced by climatic improvement. In the Glen Carron sequence there is a pine pollen rise during this dry period in contrast to the Eilean Subhainn and Glen Torridon sequences (Anderson, 1998). It is possible that by the time the climatic improvement occurred, blanket peat had already taken hold of the Eilean Subhainn and Glen Torridon catchments, making it impossible for pine to show any substantial regrowth and regeneration (Anderson, 1998).

The common hypothesis for the second pattern of decline is also of a climatic downturn similar to the one that occurred around 5000 cal BP (Anderson, 1998; H.H.Birks, 1972; Kerslake, 1982; H.J.B.Birks, 1977, 1989, 1990; Gear and Huntley, 1991). Both the Glen Carron and Eilean Subhainn sequences infer a shift to wetter conditions between 4070 and 3620 cal BP (Figure 12, 13 and table 5), this shift signifies the transition from relatively dry conditions during the early to mid-Holocene (Anderson, 1998). The growth and germination of pine decreases with increased waterlogging, therefore precipitation levels and climate can be inferred if a reconstruction of water table levels from palynological evidence is achieved (Anderson, 1998).

Seeds from pine trees on drier soils can be transported by wind, water and by animals; however, if the surface is water logged the seeds will not germinate. Therefore, pine germination over a broad area suggests a lowering of the water table prior to germination (McVean, 1963). The Glen Carron sequence in particular supports the hypothesis that scots pine growth is determined by climatic conditions as the abrupt final decline at this site is coincident with a shift to wetter conditions

(Figure 12). Deuterium isotope work on pine macrofossils in the Cairngorms indicate that this time was characterized by increased wetness (Dubois and Ferguson, 1985). The rapid and widespread nature of the decline and the lack of cultural pollen indicators further support the theory of a climatic shift rather than anthropogenic interference was the reason behind the pine decline (Anderson, 1998).

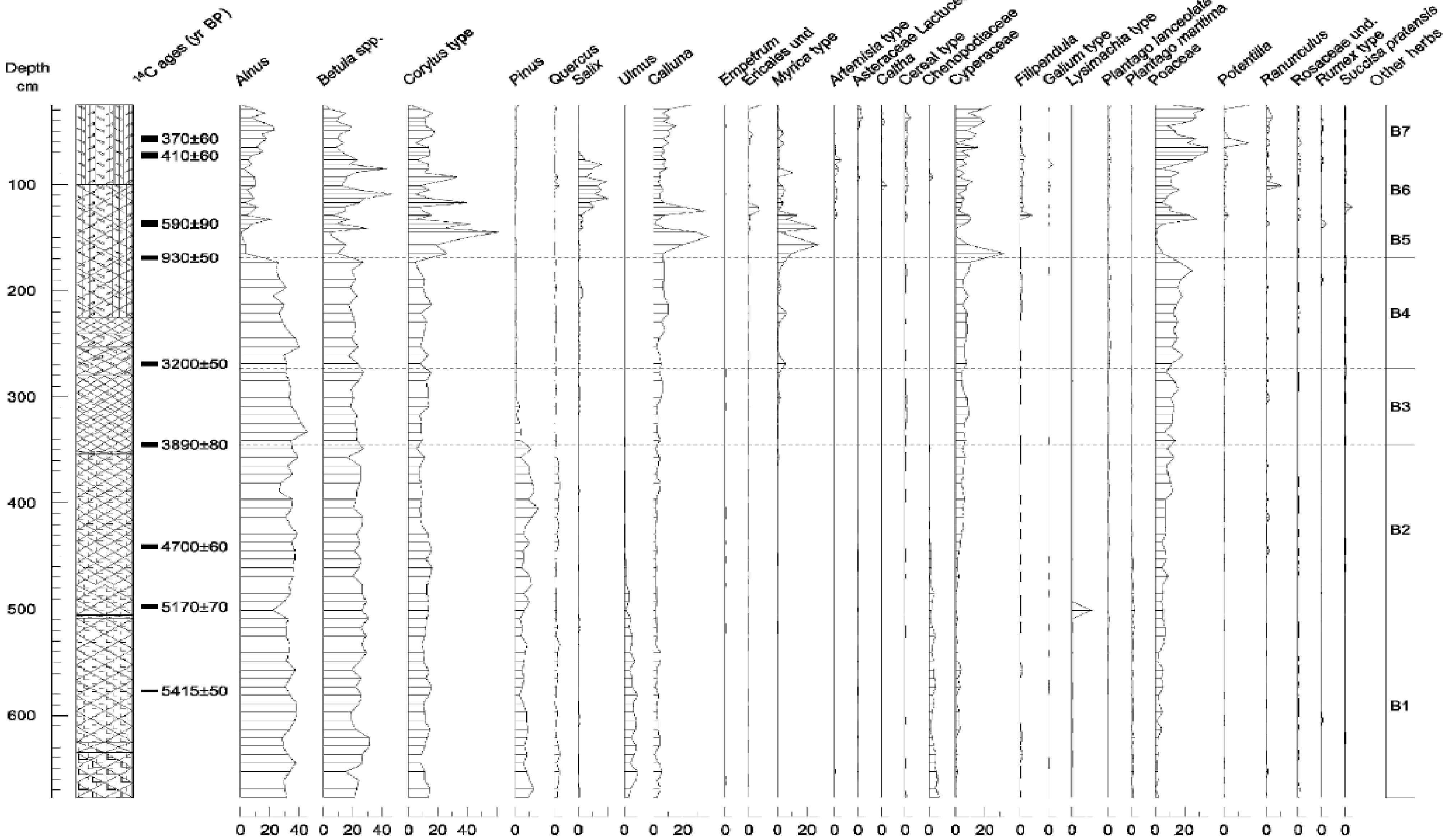


Figure 11. Pollen percentage diagram from Badentarbet on the Coigach Peninsula. A pine decline is evident at 3890±80BP (4320 cal BP) (Bunting and Tipping, 2004).

At 3500 cal BP a shift to wetter and cooler conditions is inferred from the majority of peat profiles throughout Scotland and elsewhere in the British Isles (Barber et al, 1994a, b). Both the Glen Torridon and Glen Carron cores indicate human disturbance from this times onwards with an increase in *Plantago lanceolata* and charred particles. There were further climatic shifts in the late Holocene with a dry shift between c.1550 and 1100 cal BP known as the 'Little Optimum' (Anderson, 1998). Ice core data from Greenland shows that this period was characterised by temperatures that are 1-2°C higher than they currently are (Dansgaard et al, 1989; Lamb, 1964, 1982). The high percentages of *Calluna* recorded throughout northern Scotland probably reflect the drier peatland (Charman, 1990). The 'Little Ice Age' between c.940 and 760 cal BP is the final significant climatic shift to affect the Holocene before present day. This transition to cooler and wetter conditions can be seen by a reduction in the percentages of *Plantago lanceolata* and charred particles in the Glen Torridon and Glen Carron cores; perhaps reflecting the movement of human activity out of the region as conditions became unfavourable (Anderson, 1998).

Eilean Subhainn-palaeohydrological shifts (cal BP)	Estimated 2σ age range (cal BP)	Glen Torridon palaeohydrological shifts (cal BP)	Estimated 2σ age range (cal BP)	Glen Carron palaeohydrological shifts (cal BP)	Estimated 2σ age range (cal BP)
(1) dry shift – 1410 in 10 cm (*c*. 145 yrs)	(direct date) 1550–1340	(1) wet shift 900 in 4 cm (*c*. 120 yrs)	940–800	(1) wet shift – 870 in 2 cm (*c*. 45 yrs)	940–760
(2) wet shift – 2690 in 11 cm (*c*. 200 yrs)	2850–2540	(2) dry shift – 1210 in 3 cm (*c*. 90 yrs)	1260–1100	(2) dry shift – 1390 in 2 cm (*c*. 45 yrs)	1480–1230
(3) wet shift – 3460 in 5 cm (*c*. 95 yrs)	(direct date) 3630–3270	(3) wet shift – 2400 in 2 cm (*c*. 60 yrs)	2490–2220	(3) wet shift – 3830 in 4 cm (*c*. 230 yrs)	(direct date) 4070–3630
(4) dry shift – 4310 in 5 cm (*c*. 100 yrs)	4510–4120	(4) wet shift – 3170 in 2 cm (*c*. 60 yrs)	(direct date) 3340–3060	(4) wet shift – 4400 in 4 cm (*c*. 110 yrs)	4620–4180
(5) wet shift – 5260 in 10 cm (*c*. 200 yrs)	5500–5070	(5) wet shift 3850 in 2 cm (*c*. 60 yrs)	4020–3620	(5) dry shift – 4770 in 4 cm (*c*. 110 yrs)	4970–4540
(6) wet shift – 6220 in 10 cm (*c*. 200 yrs)	6500–6040	(6) dry shift 4170 in 6 cm (*c*. 180 yrs)	4330–3890	(6) wet shift – 5130 in 4 cm (*c*. 110 yrs)	(direct date) 5230–4880
(7) wet shift – 7500 in 5 cm (*c*. 100 yrs)	7830–7320	(7) wet shift – 4950 in 8 cm (*c*. 160 yrs)	5120–4700	(7) dry shift – 5870 in 4 cm (*c*. 140 yrs)	6040–5660

Table 5. Summary of local palaeoehydrological shifts at Eilean Subhainn, Glen Torridon and Glen Carron (Anderson, 1998).

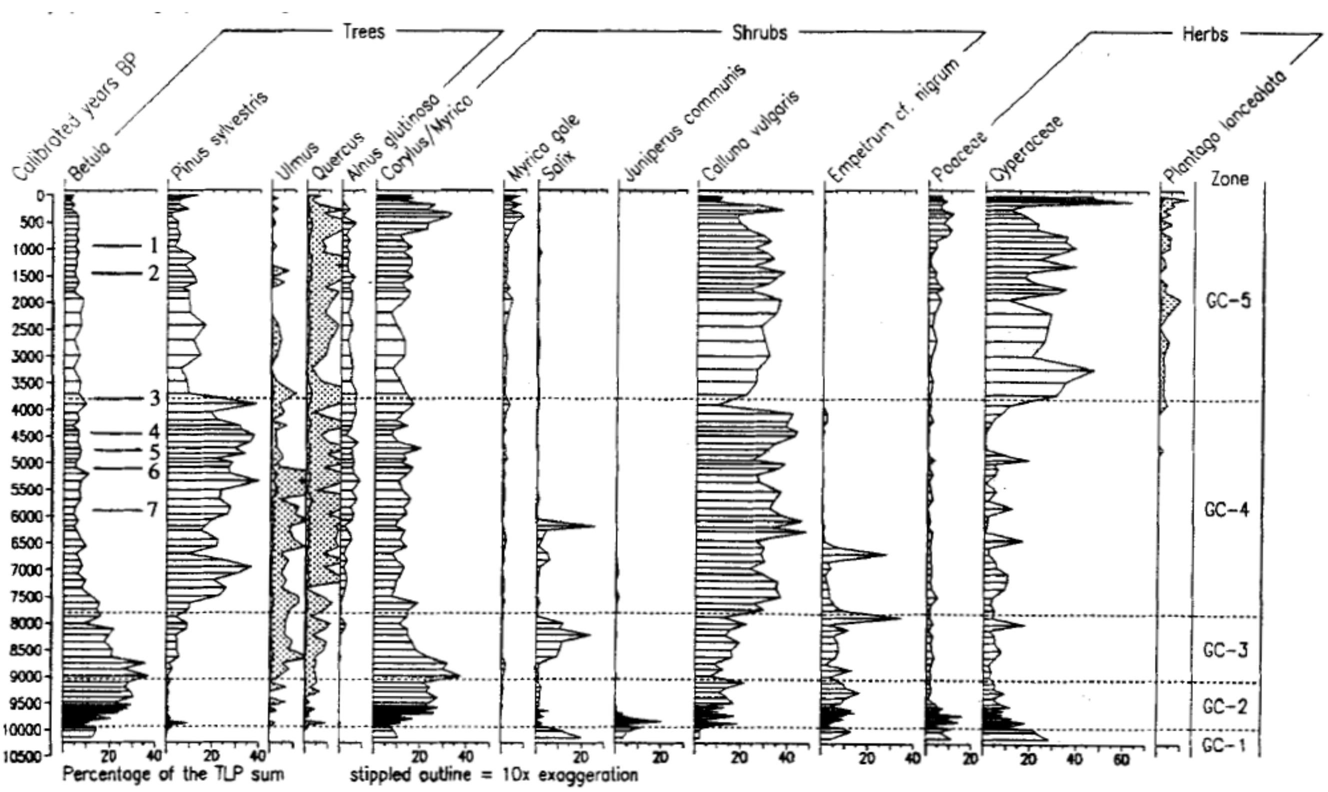


Figure 12. Summary percentage pollen diagram from Glen Carron. Local wet and dry shifts are marked by the solid lines and numbers and can be correlated with table 5 (Anderson, 1998).

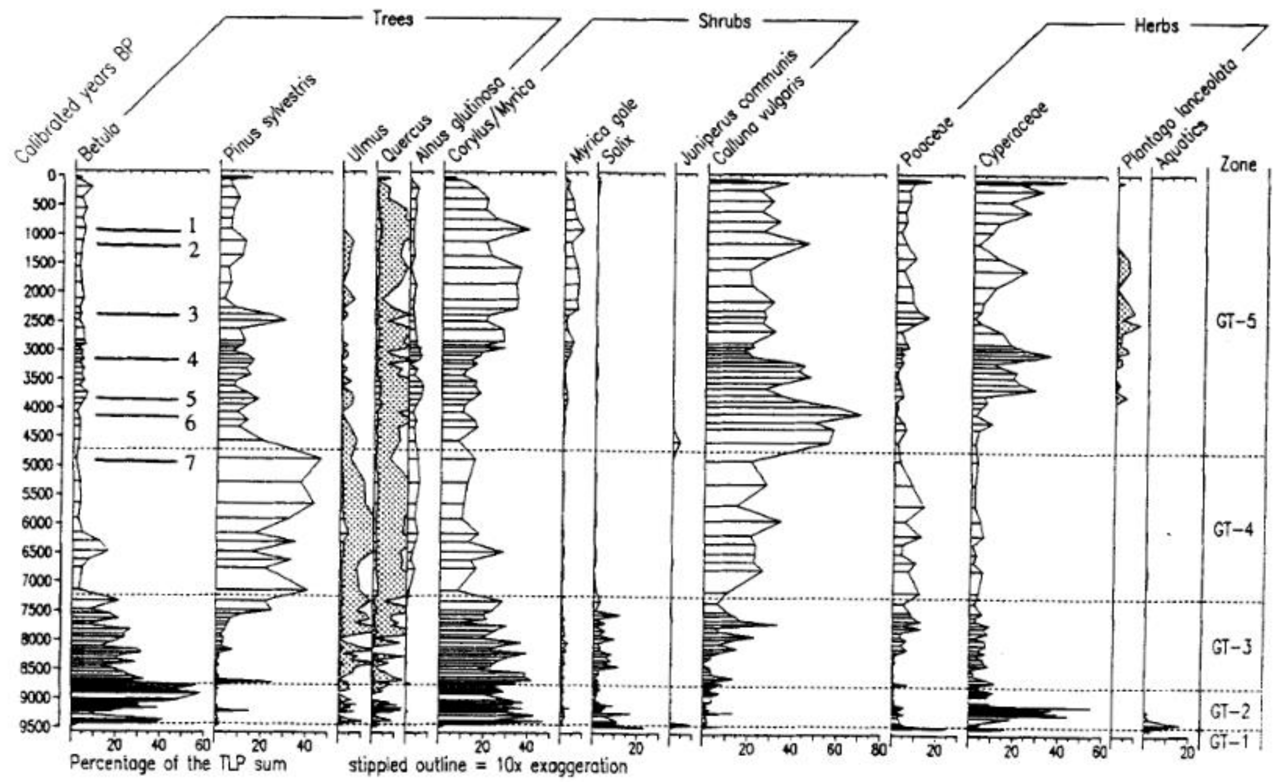


Figure 13. Summary percentage pollen diagram from Glen Torridon. Local wet and dry shifts are marked by the solid lines and numbers and can be correlated with table 5 (Anderson, 1998).

2.3.2 Late Bronze Age Climatic Deterioration and Settlement Stability

There has been much debate about the impact the climatic deterioration at 2800 cal BP had on late Bronze Age communities in northern Scotland. The Bronze Age is dated from 4500BP to 2800 cal BP. Tipping et al (2008) provides evidence to support the abandonment of marginal areas as conditions worsened at Loch Farlary at 2700 cal BP. Smith (1998) further supports this argument with evidence from Achany Glen with abandonment dated to 2900 cal BP; while Tipping (2002), Davies (2003) and Davies et al (2004) all propose that communities in west Glen Affric left the area at 2860 cal BP. The de-intensification or cessation of crop production at these sites suggests that communities left these more exposed upland locations for more sheltered lowland areas (Tipping et al, 2008). However, the reduction or termination of crop production at these sites does not necessarily mean that the settlements were abandoned (Tipping et al, 2008). In fact, there is evidence to suggest that the grazing of livestock continued in these areas after climatic deterioration. The evidence put forward by Tipping et al (2008) to support this retreat by communities to lower altitudes and perhaps even different regions due to climatic severity is more suggestive rather than conclusive. The low temporal resolution of the evidence does not provide a decisive argument for abandonment.

There is evidence from lowland Reidchalmai to suggest that the clearance of woodland intensified after 2930 cal BP to allow for more extensive land use. This period of land use lasted much longer than the brief periods of land use during the Bronze Age indicating that there was a more consistent reliance on agriculture with cereal production increasing at 2200 cal BP (Tipping, 2003). Tipping et al (2008) suggest that this increase in land use at lower altitude areas such as Reidchalmai is most probably due to local increases in population as a result of the prosperous agricultural economy. However, there is the possibility that the population increase is due to the arrival of communities who have fled the exposed upland settlements such as Loch Farlary for more sheltered lowland locations (Tipping et al, 2008). It is more likely that these uplands localities were not abandoned during this period of increased climatic stress but in fact the communities adapted new strategies to help them cope in these conditions. Crop production may well have lessened or ceased but the rearing and therefore grazing of livestock may have become a more integral aspect of subsistence life (Tipping, 2002, 2005; Davies et al, 2004).

2.4 The Complex History of *Pinus sylvestris*

A reconstruction of past vegetation is vital in order to understand past climate change and in turn predict and prepare for any climatic alterations (Birks, 1970, 1975; Charman, 1990, 1994). For centuries, peatland and dry land pines have been used as valuable climatic indicators (Linderholm et al, 2002; Birks, 1970, 1975; Charman, 1990, 1994; Anderson, 1996, 1998) with climate-growth relationships evident throughout Scotland and the northern Hemisphere. The growth and contraction of *Pinus sylvestris* can be used as an imprecise target for climate change - Linderholm et al (2002) correlated both peatland and dry land pine growth fluctuations to changes in temperature and precipitation in Sweden. After a substantial expansion of *Pinus sylvestris* between 9000 and 8000 thousand years ago, native pine woodlands became the most dominant forest type, replacing the birch-hazel forest that spread during the post-glacial warming of the early Holocene. This can be seen by the preservation of sub-fossil pine stumps throughout large parts of the Scotland Highlands (figure 14) (Birks, 1975; Bennett, 1984, 2005b; Gear and Huntley, 1991; Daniell, 1997).

Pollen sequences throughout Scotland support evidence for a rapid and synchronous decline in *Pinus sylvestris* around 4000 cal BP, a major event in Scottish vegetational history which is known as the ‘pine decline’ (Birks, 1972, 1975, 1994; Bennett, 1984, 1995; Blackford et al, 1992). Only isolated scattered pockets of native Scots pine woodland exist (Steven end Carlisle, 1959; Forestry Commission, 1999). There has been much debate as to what forcing mechanism pushed pine past its survival limits causing its collapse; however, the overall view is that a reduction in temperature and a rise in effective precipitation were behind this shift (Tipping, 2008; Birks, 1970, 1975; Charman, 1990; 1994; Anderson, 1996, 1998).

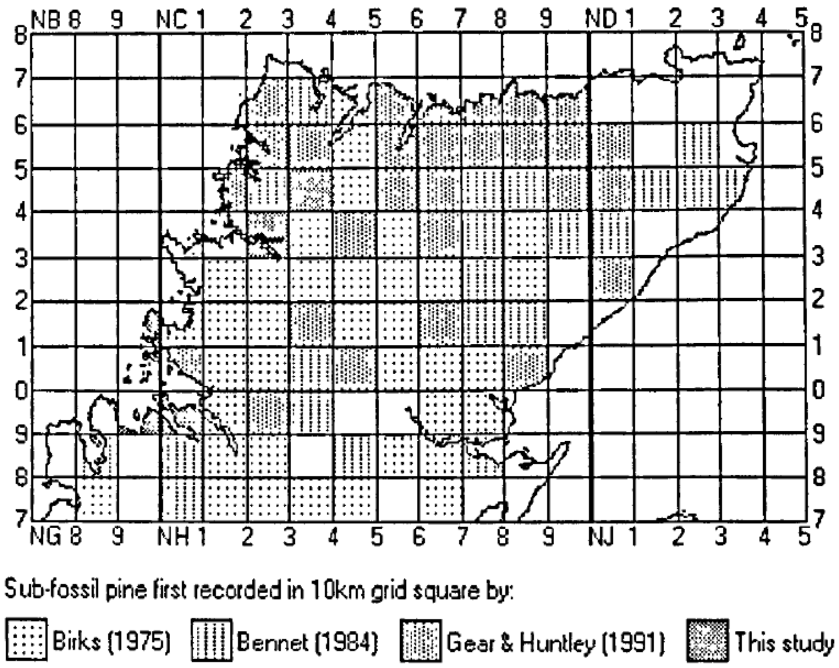


Figure 14. Distribution of sub-fossil *Pinus sylvestris* found by Birks (1975), Bennett (1984), Gear and Huntley (1991) and Daniell (1997). 'This Study' refers to Daniell (1997) (Daniell, 1997).

There are also continuing debates as to what caused this climate change, whether it is solar variability (Blauuw et al, 2004), changes in the North Atlantic Ocean and ice rafted debris events (Bond et al, 1997; Anderson, 1998) or volcanic induced climatic variation (Blackford et al, 1992; Dugmore, 1989; Bennett et al, 1992; Pilcher and Hall, 1992). Some authors argue that the pine decline was independent from any climate shift and that other factors were at play, such as anthropogenic activity (Birks, 1975; Charman, 1994; Bennett, 1995; Tipping, 2003) or pathogenic and insect attack (Matias and Jump, 2012)

The history of pine in northern Scotland is particularly interesting as native pine forests are one of Scotland's most striking forest types (Tansley, 1949) and *Pinus* woodland has an exceptionally complex history throughout northern and Northwestern Scotland (Tipping et al, 2008).

2.5 The Effect of Climate Change on the Germination and Growth of *Pinus sylvestris*

The Holocene has been a period of intense climatic fluctuations (Bell and Walker, 2005; Mayeski et al, 2004), however understanding the nature and consequence of these fluctuations has still yet to be achieved. An improved understanding of why these past changes in the climate occurred and the impacts they had would allow us to prepare for future changes, in particular the impact of global warming on the planet (Moir et al, 2010).

Atmospheric precipitation feeds ombrotrophic bogs with nutrients and water, making them very sensitive to the smallest of climatic changes (Moir et al, 2010). Danish Japetus Steenstrup (1841) was one of the first to acknowledge how bogs can preserve plant remains which in turn could be used to reconstruct past environments. Similarly to Scotland, the history of pine in Denmark is also very interesting, Danish peat bogs were once home to pine trees up until their sudden extinction. Streenstrup argued that pine had grown under cold conditions within Denmark; however Christina Vaupell disagreed saying that pine had in fact grown under dry and warm conditions (Birks, 2007).

Pine tree sequences were collected from ombrotrophic bogs in northern Scotland to ensure their increased sensitivity to changes in the climate. These sequences were then used to reconstruct climate change during the middle Holocene (Moir et al, 2010). During the Holocene, the synchronicity of variations in the range and altitude of scots pine over a large distance across northern Scotland has been attributed to climate change (Anderson et al, 1998; Bennett, 1984; Birks, 1975; Bridge et al, 1990; Dubois and Ferguson, 1985; Gear and Huntley, 1991; Huntley et al, 1997). Pine from nine sites suggests a widespread, short lived and rapid expansion in *Pinus Sylvestris* (Moir et al, 2010). Simultaneous germination at the three earliest sites probably occurred around 5200 cal BP, while, pine germinated at another four sites between 5165 cal BP and 5145 cal BP. Initial germination at all nine sites occurred within a short 100 year time interval between 5200 and 5100 cal BP indicating that this increase in both germination and growth of pine was widespread supporting the theory that climate change was responsible. Conditions favouring pine growth lasted from about 5200 to 5000 cal BP. This can be seen by the continuation of germination and two phases' of high decadal growth (7mm/yr) between 5110-5090 cal BP and 5040-5020 cal BP (Moir et al, 2010).

As pine growth and germination decreases with increased waterlogging, the reconstruction of water table levels from palynological evidence, can be used to infer precipitation levels and climate from that period. A waterlogged surface can even prevent any seeds travelled from pine trees on the drier soils around the edges of the bog, from germinating (McVean, 1963). Therefore, pine germination over a broad area suggests a lowering in the water table before the occurrence of germination.

Low water tables can be inferred from high decadal growth (Moir et al, 2010). The growth and survival of pine on bogs, is reliant on the level of the water table (Boggi, 1972; Laine et al, 1995; Legg et al, 2003); with the first germination of *Pinus sylvestris* at both the most northerly and highest sites sampled coinciding with a reduced water table (Moir et al, 2010). When compared, the pollen/spore records of pine from both White Moss in the Northwest of England (Lageard et al, 1999), and Campemoor in Germany (Leuscharer et al, 2007), support Moir et al's (2010) theory, suggesting pine declined and eventually died off due to continuous episodes of heightened water table. From 5000 to 4900 cal BP, the cessation of germination, a 73% decline in trees and decreases of decadal growth (0.6mm/yr) suggests that conditions were not suitable for pine growth, therefore pine only survived for one short generation at all nine sites (Moir et al, 2010). Periods of dry climatic conditions may have seen the pine become denser, while periods of wetter climate conditions saw the thinning of pine woodland (Godwin, 1956). This theory does not agree with the hypothesis that pine died by a brief cataclysm in northern Scotland. The decline of pine established from palynological evidence is an imprecise target for the timing of climate change. This evidence supports Gear and Huntley's (1991) hypothesis that pine in the far north of Scotland initially expanded and eventually declined due to regional climate. There are many contrasting theories as to what forcing's were responsible for this climatic downturn that resulted in the collapse of *Pinus sylvestris* around 4000 years ago, these are discussed in the next section.

Studies such as these are however prone to circular reasoning. Climate change is considered to affect pine growth and then pine growth (or lack of it) is used to infer climate change. One way to avoid the danger of circular reasoning is to use climate reconstructions that are not produced using biological proxies, such as vegetation. Preferably meteorological data would be used as it is independent from vegetation. Although, climate data has been reconstructed using the vegetation profile from Meall Bad á Chrotha, wherever possible meteorological data or independent climate reconstructions, collected from varying authors, have been used to support or conflict with the climate data inferred from the vegetation in this study – in an attempt to minimise the impact of a circular argument.

2.6 Forcing Mechanisms behind Climate Change

2.6.1 Ice Rafted Debris Events and Changes in the North Atlantic

Bond et al (1997) propose that several cycles occurred during the Holocene when cold, ice packed waters migrated south from northern Iceland and Greenland via the Labrador and East Greenland Currents, travelling down as far as northern Britain. These cycles referred to as 'Bond Events' are dated to 1400, 2800, 4200, 5900, 8100, 9400, 10,300 and 11,100 cal BP. These ice rafting debris events reduced sea surface temperatures, which greatly affected the climate in regions around the North Atlantic, in particular Northwest Scotland.

Bond event number 3 at 4200 cal BP coincides with the widespread climatic deterioration and shift to wetter conditions inferred from numerous proxies throughout northern Scotland at the time of the pine decline (Anderson, 1998). Indeed, the wet to dry then back to wet oscillations which resulted in the two phase pine decline identified by Anderson (1998) at Eilean Subhainn, Glen Carron and Glen Torridon can be linked to the two periods of cooler sea surface temperatures associated with the ice rafted debris events and the separating warmer phase identified by Bond et al (1997) in the North Atlantic.

2.6.2 Solar Variability

Variations in solar activity could well have been an imperative factor influencing the mid-Holocene climate change which signalled the decline of *Pinus sylvestris.* The $\Delta^{14}C$ proxy can be used to reconstruct any Holocene solar variability changes (Blaauw et al, 2004; Stuiver and Braziunas, 1993, 1998; Stuvier et al, 1998; Chambers et al, 1999; Beer, 2000; Goslar, 2002). Climate changes during the Holocene and fluctuations in $\Delta^{14}C$ have been linked in numerous studies (Blackford and Chambers, 1995; Karlen and Kuylenstierna, 1996; Chambers et al, 1999; Björck et al, 2001; Bond et al, 2001; Hodell et al, 2001; Neff et al, 2001; Magny, 2004). Chronologies of high precision can be obtained from peat using ^{14}C wiggle-match dating. The transition from the Sub-Boreal period (5-2.5ka), which was characterized by dry and warm conditions, to the Sub-Atlantic (2.5ka-present) period, which is the coldest and wettest period of the Holocene, corresponded with a sudden influx in $\Delta^{14}C$ (Van Geel et al, 1996; Speranza et al, 2000, 2002; Mauquoy et al, 2002a; 2002b). ^{14}C wiggle-match dating was used to identify this high precision link between the increase in ^{14}C and the wet shifts recorded in peat deposits throughout the Northwest and Europe.

Blaauw et al (2004) used the data collected from two peat cores from raised bogs in the Netherlands to examine the link between mid-Holocene climate changes and solar variability. The two cores undergone local vegetation reconstruction and high resolution ^{14}C wiggle match dating (figure 15 and 16), in order to produce a record from c.4500 to c.340 cal BC (Blaauw et al, 2004). The chronologies obtained from ^{14}C wiggle-match dating of cores MSB-2K and Eng-XV can be seen in figure 17. Raised bogs are ideal archives of climate change as they depend solely on precipitation for water and nutrients. Any fluctuations in precipitation or moisture levels can be easily inferred from the plant taxa which occupy the bogs at any given time. Water table depth influences which type of plant taxa grow, as each taxon requires different moisture levels (Malmer, 1986; Hammond et al, 1990; Økland, 1990; Van der Molen, 1992; Wheeler and Proctor, 2000; Økland et al, 2001). Figures 15 and 16 along with table 6 show that the majority of significant increases in residual $\Delta^{14}C$ correspond with shifts to wetter conditions (MSB – 1, 3, 4, 5, 6, 7; Eng – 5, 8, 10).

However, there were two major $\Delta^{14}C$ rises with no correspondent wet shift (core MSB – 2k during the $\Delta^{14}C$ rise of c.4265 – 4215 cal.BC and core Eng – XV during the $\Delta^{14}C$ rise of c.1535 – 1485 cal.BC) (Blaauw et al, 2004).

Possible links between climate change and solar variability have long been suggested and studied (Wijmstra et al, 1984; Blackford and Chambers, 1995; Bond et al, 2001; Chambers and Blackford, 2001; Magny, 2004); however these studies were largely imprecise due to the chronologies used. The introduction of ^{14}C wiggle match dating has dramatically improved the precision of peat chronologies allowing the comparison of short-term climatic events, such as decadal to centennial, and $\Delta^{14}C$ fluctuations (Blaauw et al, 2004).

Past fluctuations in $\Delta^{14}C$ (atmospheric ^{14}C content) have been due to changes in ^{14}C production (regulated by solar variability, geomagnetic field strength and/or galactic cosmic ray flux) and/or carbon cycle changes (particularly ocean ventilation). Galactic cosmic rays that enter the earth's atmosphere produce cosmogenic isotopes such as radiocarbon and ^{10}Be. A reduction in solar wind as a result of decreased solar activity can lead to an increase in the level of galactic cosmic rays entering the earth's atmosphere. Ordinarily, the earth's magnetic field and the solar wind, which is a low density proton electron gas emitted from the sun, would create a barrier stopping cosmic rays entering the earth's atmosphere (Hoyt and Schatten, 1997); Beer, 2000).

2.6.2.1 $\Delta^{14}C$ and Solar Activity

There are a number of abrupt and significant $\Delta^{14}C$ rises recorded during the Holocene (2850 cal BP and the "Little Ice Age") that are the result of a reduction in solar activity (Blaauw et al, 2004; Stuiver and Braziunas, 1993; 1998; Chambers et al, 1999; Beer, 2000; Beer et al, 2002; Goslar, 2002). Changes in the climate and sunspot indices have gone hand in hand with the levels of radiocarbon and ^{10}Be (Beer, 2000). In recent years however nuclear bombs and fossil fuel burning has interfered with ^{14}C.

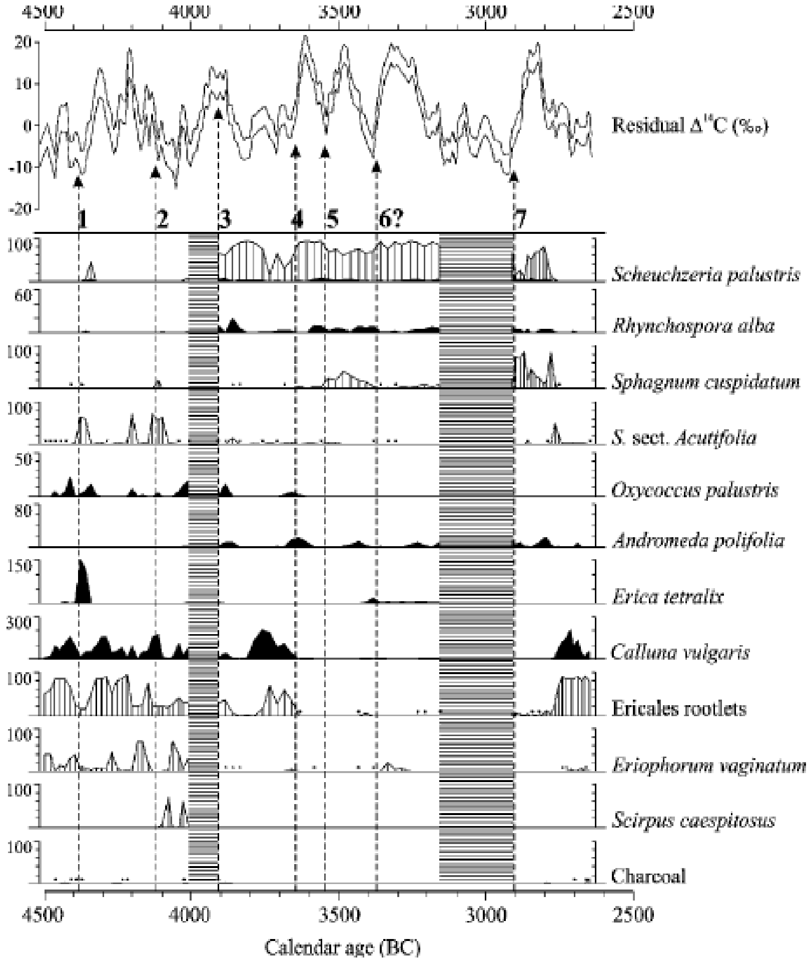


Figure 15. Local vegetation changes and residual $\Delta^{14}C$ from the MSB-2K core. Chronology is from figure 17**.** Hiatuses are shown by the stripped vertical boxes. The vertical numbered lines show wet shifts. Macrofossils in the estimated volume percentage are indicated by the vertically hatched areas. The black areas indicate where macrosfossils were counted as numbers and the black dots show where macrofossils were counted at low amounts (Blaauw et al, 2004).

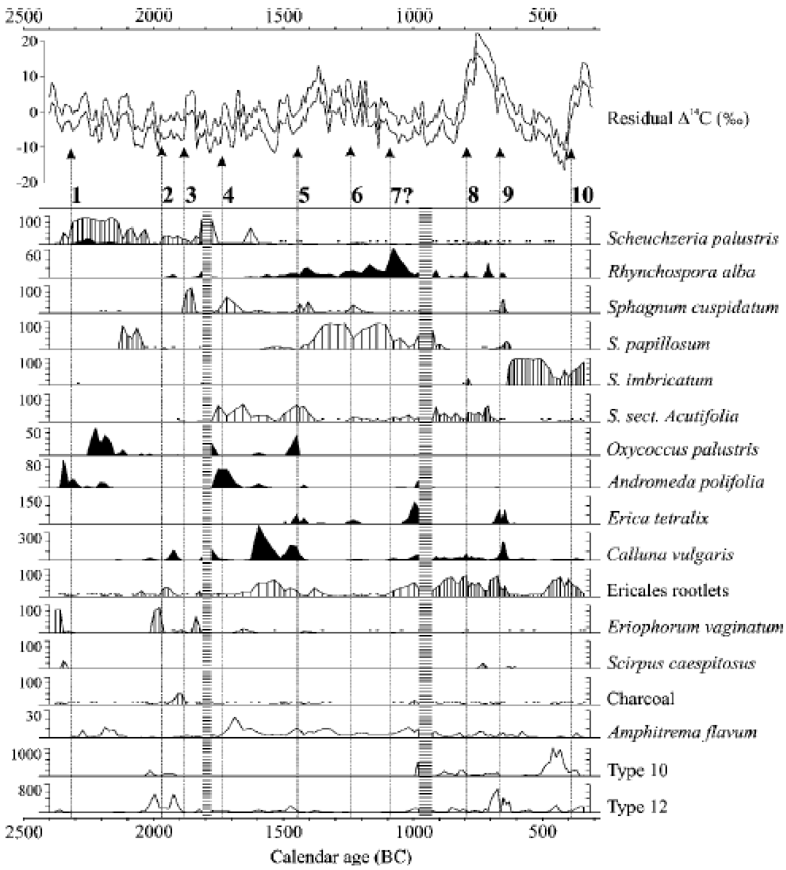


Figure 16. Local changes in vegetation and residual $\Delta^{14}C$ from Eng-XV core with the chronology from figure 17. See figure 15 for detail (Blaauw et al, 2004).

Major $\Delta^{14}C$ rise (start–end; cal. BC)	Amplitude $\Delta^{14}C$ rise (in ‰)	Corresponding wet-shift (cal. BC)
c. 4375–4315	18.7	MSB-1 (*c*. 4390)
c. 4265–4215	18.6	(no wet-shift)
c. 4005–3935	18.2	MSB-3 (*c*. 3910)
c. 3665–3615	20.4	MSB-4 (*c*. 3635)
c. 3545–3485	17.1	MSB-5 (*c*. 3535)
c. 3385–3325	22.9	MSB-6 (*c*. 3360)?
c. 3105–3075	10.6	(hiatus)
c. 2925–2825	26.9	MSB-7 (*c*. 2895)
c. 2505–2455	14.8	(no record)
c. 1535–1485	11.2	(no wet-shift)
c. 1465–1365	13.4	Eng-5 (*c*. 1435)
c. 845–755	26.0	Eng-8 (*c*. 785)
c. 415–345	25.5	Eng-10 (*c*. 385)

Table 6. Mid-Holocene major $\Delta^{14}C$ rises, their duration and the corresponding wet shifts found in the cores (figure 15 and 16) (Blaauw et al, 2004).

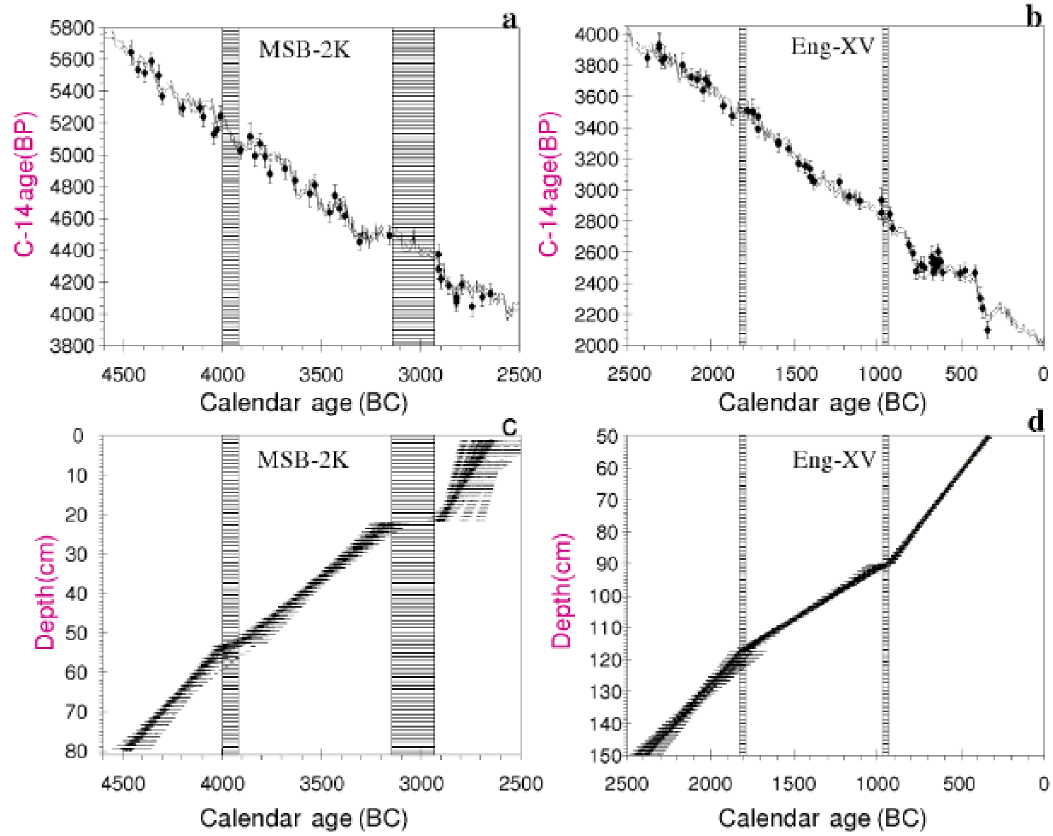


Figure 17. Chronologies from cores MSB-2K and Eng-XV obtained from ^{14}C wiggle-match dating. ^{14}C sequences were divided up into three subsets based on stratigraphy changes and the best position on the calibration curve for the ^{14}C dates (a and b). Every depth is plotted with the most likely calendar year age (c and d). 1 σ confidence intervals have been given for ^{14}C calendar ages. Hiatuses (break or interruption n in the sequence i.e. wet shifts) are shown by the vertical boxes with horizontal lines (Blaauw et al, 2004).

2.6.2.2 $\Delta^{14}C$ and Cosmic Ray Intensity

The levels of ^{14}C and other cosmogenic isotopes in the atmosphere can be influenced by intensification in the galactic cosmic ray flux. Intensification in the cosmic ray flux can be attributed to supernovas for example (Blaauw et al, 2004). However, increases in cosmogenic isotopes due to a rise in galactic cosmic ray intensity have not been linked to any climate events during the Holocene (Shaviv, 2002). There have also been numerous debates suggesting that the cosmic ray flux is actually relatively stable (Blaauw et al, 2004).

2.6.2.3 $\Delta^{14}C$ and Geomagnetic Field

The level of ^{14}C in the atmosphere could also be linked to changes in the geomagnetic field. The assumption among the majority of studies is that any geomagnetic fluctuations will only influence cosmogenic isotope changes over an extended time period (>3ka) (Merrill et al, 1996; Beer, 2000; Beer et al, 2002). Records collected from several regions indicate geomagnetic variations during the Holocene (Valet et al, 1998; Ali et al, 1999; Gogorza et al, 2000; Yang et al, 2000; Laj et al, 2002; Ojala and Saarinen, 2002; Snowball and Sandgren, 2002). Over extended time scales, geomagnetic activity is fairly continuous and comparable between records, however over the shorter term there are differences; with some records showing short-term declines, inclines and fluctuations in intensity. Geomagnetic activity is highly complex and we have very little knowledge to help us explain any alterations in activity. It is not known whether the changes recorded were global or the result of chronological problems, sedimentation changes and/or local, non-dipolar variations in the geomagnetic field (Merrill et al, 1996). However, local variations in the geomagnetic field would not trigger a global rise in cosmogenic isotopes (Blaauw et al, 2004). Furthermore, the $\Delta^{14}C$ variations recorded and the geomagnetic changes did not correspond. "The current lack of high resolution reconstruction of the geomagnetic field intensity, it cannot be assumed that short-term (<10^3 year) variations in solar activity are solely responsible for similar duration anomalies in the production rates of cosmogenic nuclides, as the internal dynamics of the earth's geodynamo may promote similar features" (Snowball and Sandgren, 2002). Merrill et al (1996) emphasise that even if significant $\Delta^{14}C$ fluctuations were the result of geomagnetic field changes, it is possible that solar activity changes were responsible for these geomagnetic activity changes.

2.6.2.4 $\Delta^{14}C$ and the Ocean

The shift to cold conditions which signified the Younger Dryas are reported to have been partly due to a major interruption in the North Atlantic Thermohaline Circulation (Hughen et al, 1998, 2000; Muscheler et al, 2000; Marchal et al, 2001) which in turn caused a substantial rise in $\Delta^{14}C$ levels. It has been suggested that the level of $^{14}CO_2$ in the atmosphere changed due to an interruption in the exchange of CO_2 between the atmosphere and the ocean (Blaauw et al, 2004). Atmospheric ^{14}C oxidises to form $^{14}CO_2$ and becomes part of the global carbon cycle. It is possible that the rise in $\Delta^{14}C$ levels occurred when the atmosphere increased its level of CO_2 uptake from the ocean. Also, a reduction in ocean ventilation along with continued $^{14}CO_2$ production would create an increase of $^{14}CO_2$ in the atmosphere (Blaauw et al, 2004). However, a reduction in solar activity as supposed to a change in ocean dynamics has been proposed as an alternative hypothesis for the $\Delta^{14}C$ rise which led to the Younger Dryas cooling event (Goslar et al, 2000; Renssen et al, 2000). In order for the ocean to have caused such major and abrupt increases in the levels of ^{14}C in the atmosphere, a significant change in ocean circulation would have been required. However, no ocean circulation changes which would be significant enough to cause such an event have been reported during the Holocene (Chapman and Shackleton, 2000; Keigwin and Boyle, 2000; Bond et al, 2001). Consequently, Stuiver et al (1991) concluded that ocean circulation could not have been responsible for the major rises in $\Delta^{14}C$ during the Holocene.

2.6.2.5 Solar forcing of Climate Change

The main supporting evidence for a change in the production of ^{14}C, as supposed to ocean circulation, was the fact that ^{10}Be also showed similar changes alongside $\Delta^{14}C$ (Bard et al, 1996). Within the last c.350 years, variations in the sunspot cycle and solar irradiance coincided with changes in ^{14}C and ^{10}Be levels (Hoyt and Schatten, 1997).

Nine out of eleven significant rises in atmospheric ^{14}C recorded by Blaauw et al (2004) during the mid-Holocene coincided with wet shifts which indicate a shift to cooler conditions. These $\Delta^{14}C$ rises were more than likely due to reductions in solar activity. The paper by Blaauw et al (2004) emphasises the importance of not only the role that solar variability played in Holocene climate change but also how important studies looking at solar activity and climate change are in helping predict any future climatic events such as Global Warming.

2.7 Volcanic Activity and the Final Pine Decline

Pollen frequencies show an abrupt decline and southward shift in *Pinus sylvestris* which coincide with a layer of volcanic ash found in northern Scottish peat (Blackford et al, 1992) and other Holocene peat sites throughout the northern British Isles (Dugmore, 1989: Bennett et al, 1992; Pilcher and Hall, 1992). Icelandic eruption clouds typically travel north, east and Southeast of Iceland, carried by the westerly winds (Thorarinsson, 1980). The effect of acid deposition, such as plant damage, has been reported in Norway and Sweden (Thorarinsson, 1981). Many tephra falls have been reported in both Ireland and Scotland and it is likely that acid deposition has been associated with such events (Hunt, 1993).

Isochronous marker horizons provided by tephra layers are valuable if they can be traced over broad areas and can be linked to a particular eruption (Einarsson, 1986). Studies carried out in northern Scotland suggest a link between the tephra fallout from the Hekla-4 Icelandic eruption and the near extinction of pine (Hall et al, 1993). About 9km^3 of volcanic material was ejected when Hekla-4 tephra was formed during one of the largest Icelandic post-glacial eruptions (Larsen and Thorarinsson, 1977). Tephra layers in northern Scotland are located in an ideal area for research into the history of pine during the Holocene. An abrupt and extensive decline in the level of *Pinus sylvestris* pollen at 4000 cal BP (Pearson et al, 1986) has been identified by many pollen and macrofossil studies throughout the northern British Isles (Bennett, 1984). Blackford et al (1992) carried out a tephra linked pollen study in Altnabreac, in The Flow Country, northern Scotland. A decline in *Pinus* and the deposition of volcanic ash from Hekla-4 is distinct in the Altnabreac pollen, charcoal and tephra diagram (figure 18). Between 950 and 974mm tephra appears, peaking at $>1 \times 10^6$ shards cm^{-3} at 960mm. Just above the tephra maximum the abundance of *Pinus* pollen decreases to 1% Total Land Pollen (TLP) from 17% TLP (Blackford et al, 1992). The levels of *Sphagnum,* Cyperaceae and Ericaceae increase at the base of the tephra layer, while the level of *Calluna* declines before increasing again at 954mm. At Altnabreac the pollen records show changes in mire taxa which are coincident with ash fall indicating the fallout caused by a volcanic eruption may have played a direct role in the *Pinus* decline (Blackford et al, 1992). Volcanic dust veils can cause evaporation and insolation, reduced temperature and an increase in precipitation; the increases in Cyperaceae at Altnabreac indicate that there was a shift to wetter conditions around the time of the Hekla-4 eruption and the *Pinus* decline (Blackford et al, 1992). A decrease in *Pinus* frequencies has been linked with an increase in bog surface wetness suggesting that the final pine decline could have been caused by a sudden climatic deterioration induced by volcanic activity (Blackford et al, 1992). Overall, the Altanbreac study suggests two possible causes for the sudden decline in pine, both a result of the Hekla-4 eruption; a climatic deterioration as a result of volcanic dust veils or acid pollution produced by volcanic chemicals.

Numerous dendroclimatological studies suggest that large scale volcanic activity can be followed by a climatic downturn and reductions in pollen productivity (Hall et al, 1993). La Marche and Hirschboeck (1984) and Ballie and Munro (1988) conclude a close relationship between dust veils produced by significant volcanic eruptions and frost damaged or exceptionally narrow tree ring bands in modern and subfossil wood from softwoods and hardwoods sampled from extensive areas of the British Isles and the USA.

Tephra linked pollen studies have also been carried out in Northern Ireland at Sluggan Bog and Garry Bog (Hall et al, 1993). In both profiles, below the tephra layer, the pine pollen percentages are low <1% at Sluggan Bog and 2 – 3% at Garry Bog. Pollen concentration values for pine at Sluggan Bog are well below those at Garry Bog. Even though, there are slight falls in pollen percentages which coincide with the Hekla-4 eruption, the falls were so slight that they are within the error margin of the concentration values. Therefore, there is no evidence to support the theory that the reduction in pine pollen productivity was the result of volcanic activity (Hall et al, 1993).

Caseline et al (1998) studied three pollen profiles in west Corlea, Country Longford, Northern Ireland. It is clear that the level of bog surface wetness increased in conjunction with the tephra layer. There is substantial evidence to support a shift in humification and increased wetness approximately c.4300 cal BP from these three adjacent sites and extensive areas of Northwest Europe (Barber et al, 1994). Some studies have suggested earlier dates for the humification and wetness shift (Nilssen and Varren, 1991) however, it is probable that the increased wetness initially began prior to the Hekla-4 eruption, and therefore the volcanic event just contributed to the increased wetness trend and humification shift (Dugmore et al, 1995). Although, a correlation between distressed Irish oak trees and acid peaks produced by volcanic activity has been found in Greenland ice cores (Ballie and Munro, 1988), while the disaster evident in archaeological records in Britain has been attributed to the Hekla-3 eruption during the late second millennium BC (Burgess, 1989).

Dugmore et al (1995) argue that temperature records and models indicate that climate change produced by Icelandic volcanic eruptions and the impact such change had on the environment of the British Isles was minimal; however, Grattan and Charman (1994) suggest that acid volatiles and halogens transported by such eruptions are the more likely cause for any palaeoenvironmental change. The eruption of Lake Fissure in AD1783 is said to have caused substantial damage to trees and crops in Britain and Western Europe due to acid deposition (Grattan and Charman, 1994). Table 7 shows detailed recordings of the climatic conditions and acid damage caused after the Lake Fissure eruption. The recordings demonstrate how a wide area can be affected by an eruption and the consequences the eruption can have. Climatic fluctuations have long been suggested as the primary cause for any change in the ecosystems of the British Isles, however, many now argue that the emittance and deposition of acids and halogens from eruptions in Iceland might have a more significant impact on ecosystems than climate change (Grattan and Charman, 1994). Gases generated from volcanoes can fill the lower atmosphere creating a haze; there is evidence to show that this haze was present during the Lake Fissure eruption in Western Europe.

Source	Location	Date	Climatic conditions	Damage reported
Cowper (1783), in King and Ryskamp (1981)			Reduced visibility, smoky haze	
Ipswich Journal, 12 July 1783	East Anglia (Ipswich)	22–28 June	'Uncommon gloom', profuse dews, sun partially obscured	Cereal crops: yellowed Wheat: least damage (awns of bearded wheat) Barley: withered awns Oats: calyces withered, grain unaffected
Cambridge Chronicle and Journal 5 July, 1783	Cambridge	23 June	Frost	Dried pasture Beans and 'stag of the wheat': whitened, dying
Cullum (1783)	Bury St Edmunds	23 June	Severe frost	Barley and oats: brown and withered Rye: mildewed Wheat: unaffected Larch, Weymouth pine, Scotch fir: leaf tips withered Ash, walnut, cherry, peach, filbert, hazel: leaf loss Barberry, *Hypericum perforatum*, *H. hirsutum*, blackthorn, sweet violet: 'pinched' Mulberry, fig, vine: in sheltered places undamaged
Sherbourne Mercury, 14 July 1783	East counties of England	23–24 June	Severe frost	Barley and oats: yellowed Walnut: leaf loss Larch and firs: 'suffered severely'
White (1789)	UK and Europe	23 June–20 July	Haze, smoky fog	
Caledonian Mercury, 5 July 1783	Leith, Scotland	2 July	Storms, heavy rain	Various fish killed (attributed to lightning)
Aberdeen Journal, 18 August 1783	Provence, France	11 July	Odorous, dry fog	
Ipswich Journal, 9 August 1783	Embden, Germany	12 July	Thick, dry fog, 'infectious smell'	Withered leaves Leaf loss in trees

Table 7. Detailed recordings of climatic conditions and the acid damage reported after the Lake Fissure eruption, AD1783, in Britain and Western Europe (Gratton and Charman, 1994).

The correlation between these events and volcanic activity has been explained by a simple volcanically induced climate shift to wetter and cooler conditions (Ballie and Munro, 1988; Ballie, 1989), however Hunt (1977) and Nicholls (1988, 1990) have questioned the significance of the relationship between volcanic forcing and atmospheric circulation. Dugmore et al's (1995) argument that the relationship between surface temperature and volcanic forcing is minimal is supported by a number of authors (Newell, 1981; Self et al, 1981; Kelly and Sear, 1984; Bradley, 1988; Mass and Portman, 1989).

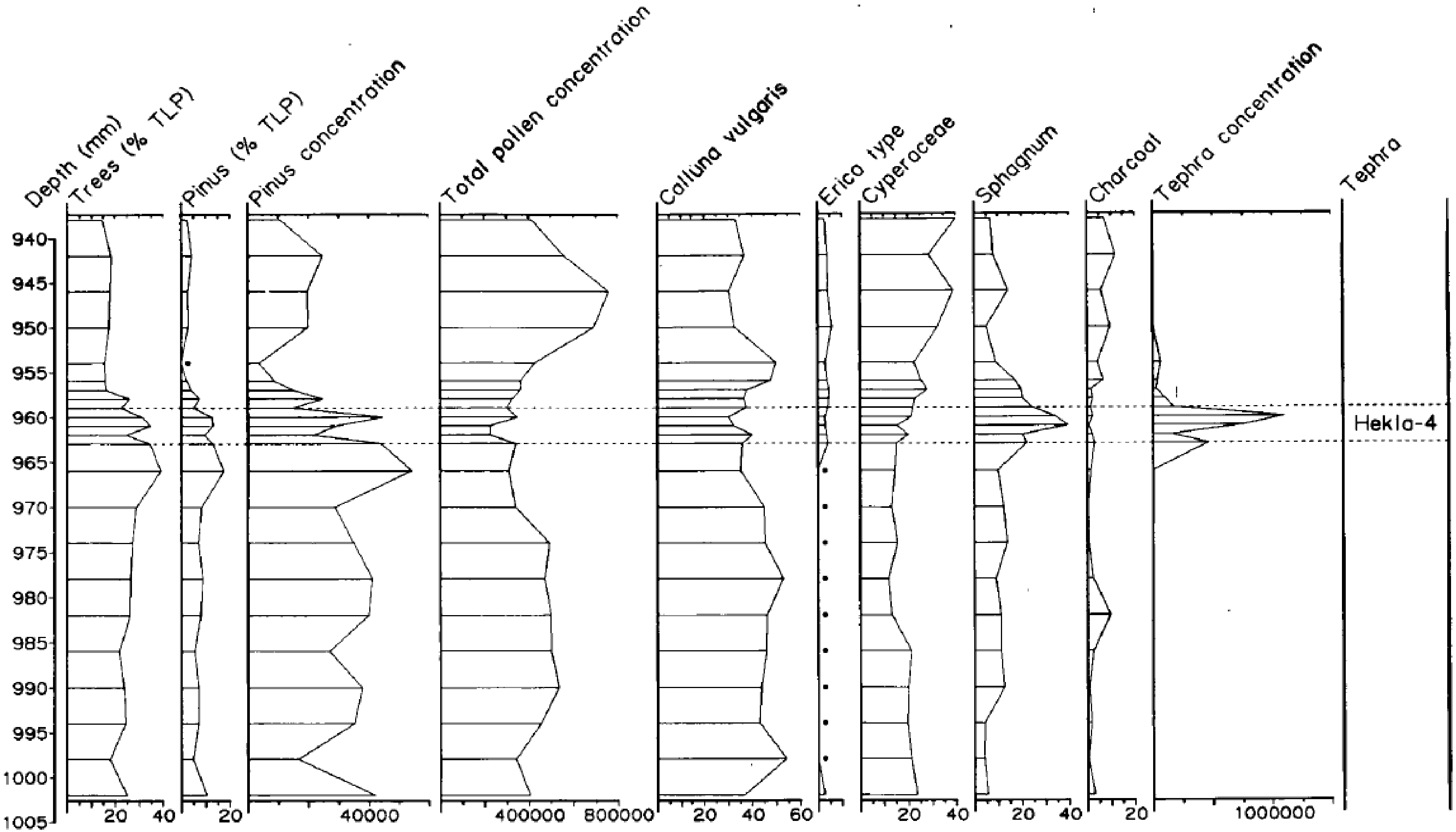


Figure 18. Tephra, pollen and charcoal diagram from Altnabreac. Percentages are based on total land pollen, concentrations are in pollen grains or shards cm -3, and charcoal concentrations are in terms of fragments >10µm per pollen grain. The circles show a percentage of less than 2% TLP and horizontal broken lines show the maximum tephra concentration zone (Blackford et al, 1992).

Blackford et al (1992) suggests a substitute explanation. Soluble volatiles and other pollutants are absorbed onto the tephra particles in the eruption cloud (Rose, 1977; Oskarsson, 1980). These tephra particles are then transported over great distances, with halogens and acid gases. They are then deposited as dry particles (Fowler, 1984), acid rain and snow (Koerner and Fisher, 1982; Davies et al, 1984) or acid fog, mist and dew (Wisniewski, 1982). *Pinus* is especially vulnerable to acidic precipitation (Caput et al, 1978) as are the principal taxa of the mire habitat in which they grow (Lee et al, 1986). Acid pollution was produced when Lake Fissure erupted in AD1783, with various areas being affected, although the nature of the eruption and the accompanying atmospheric conditions encouraged the transport of the acid pollution, therefore a wide area was affected (Gratton and Charman, 1994).

The subject of the Hekla-4 eruption and the final pine decline in northern and Northwestern Scotland has caused significant debate over the link between volcanism and vegetation (Bennett, 1984; Baille and Munro, 1988; Blackford et al, 1992; Birks, 1994; Hall et al, 1994). Nevertheless, when carrying out palaeoenvironmental research, volcanic activity must be strongly considered (Blackford et al, 1992).

2.8 Anthropogenic Activity and the Decline of *Pinus sylvestris*

Cores collected from Glen Carron and Glen Torridon indicate a shift to increased bog wetness between 4070 and 3620 cal BP (Table 5) (Anderson, 1998). This shift coincides with the final pine decline and the appearance of *Plantago lanceolata* and an increase in charred particles. This indication of human activity complicates the interpretation for the regional wet shift as land clearance by burning can influence catchment hydrology due to the reduction of soil permeability and increased waterlogging, if conducted on a substantial scale (Anderson, 1998). In many areas of upland Britain, this prehistoric human activity is thought to have led to the establishment and expansion of blanket peat (Moore, 1972, 1986a, 1986b).

Although, human activity can be an important factor in catchment hydrology, the bogs studied within Wester Ross, Glen Carron and Glen Torridon are isolated so it is unlikely that such activity had a profound effect on bog hydrology (Anderson, 1998). The synchronicity and widespread nature of the final pine decline throughout Scotland suggests that the decline was not related to human activity as at this time the level of human disturbance was still minor and incapable of causing such a significant decline event; although, human interference can be attributed to some land clearance in scattered areas further south around this time. The levels of *Plantago lanceolata* found at Glen Carron and Glen Torridon are very low suggesting that any land use in the form of agriculture was not in close proximity to the bogs, and if it was, it certainly was not intensive, therefore the wetter conditions are more likely due to climatic deterioration.

Tipping (1994) has disregarded the clearance of pine woodlands by humans due to the synchronicity of the collapse over such a wide area. However, Birks (1975), Charman (1994), Tipping (2003), Tipping et al (2007) and Bennett (1995) have all argued Tipping's (1994) assumption. Bennett (1995) argues that the intensification of grazing by domesticated animals pushed trees, which were already growing in marginal conditions in the north and west, past their survival limits. As a result, woodland began to decline which allowed for the expansion of blanket bog which in turn sealed off the mineral soils required for any regeneration of tree species (Bennett, 1995). The extent and rate of woodland collapse varied over time and space due to the differing degrees of clearance through the use of fire and tools, or the continuing clearance through sustained grazing (Bennett, 1995). Charman (1994) claims that it would be unreasonable to ignore the role human impact played in the collapse of *Pinus sylvestris*. Certainly in The Flow Country, there is extensive evidence for the presence of people since the Neolithic period. At Cross Lochs, the level of charcoal fragments rise substantially, further supporting the hypothesis that human activity did indeed play a central role in shaping the treeless landscape that we see today (Charman, 1994). Hedges (1984) supports Charman's

(1994) claim by stating that there were progressive communities present on the Orkney Islands at the time of the pine decline and evidence at Loch Farlary suggests human interference did contribute to woodland dynamics (Tipping, 2003).

An early decline in pine and then a regeneration of the species can be witnessed at Loch Farlary in Sutherland, however, in contrast to the two phase decline within the Loch Maree catchment, the reason behind pine dynamics at Farlary is harder to determine. *Alnus* became a more dominant tree species after the decline in *Pinus sylvestris* around 6400 cal BP due to a wet shift, until farming communities disturbed woodland during the late Neolithic for 150 – 200 years from 5100 cal BP for pastural purposes and the cultivation of cereal (Tipping, 2003). This anthropogenic activity and small scale but continuous grazing resulted in the collapse and retreat of *Alnus, Corylus* and *Betula* and the prompt spread of heathland at Loch Farlary. Woodland regeneration and pine re-colonisation began at 5000 cal BP when there was a reduction in grazing stress (Tipping, 2003).

The majority of the pine stumps identified at Loch Farlary which showed axe cuts grew between 5200 and 4350 cal BP. Metal axes were used on these pines after the trees had died; these axe marks were dated to the period after 4200 cal BP. Progressively from 4000 cal BP to 3250 cal BP, *Pinus sylvestris* declined until it disappeared from Farlary and the surrounding area. Tipping et al (2007) suggests that these pine stumps were most likely already dead and covered in a layer of peat when they were cut during the middle to late Bronze Age, therefore disregards the theory that the pine decline was a result of tree felling. Farming communities most likely exploited the land through peat cutting, which is still a common occurrence today. By this time the landscape would have been nearly entirely treeless, therefore the preserved wood in the peat would have been used as a valuable resource (Carter, 1988; Tipping et al, 2007). There is no evidence to support the common assumption that a transition to wetter conditions pushed pine past its growing threshold at Loch Farlary (Dubois and Ferguson, 1985; Bridge et al, 1990; Gear and Huntley, 1997). In fact, data suggests that one of the driest periods recorded occurred between 4100 – 3800 cal BP. However, Tipping et al (2007) does present evidence to support the theory that the grazing of livestock may have been the contributing factor in the final pine decline. Both *Pteridium* and *Plantago lanceolata* increase in the record after 3500 cal BP, with also a rise in charcoal concentrations after 3040 cal BP. This rise in charcoal concentrations may well have been due to the controlled burning of *Calluna vulgaris* as a means of improving the quality of grazing. Therefore, the conclusion is that the final decline at Loch Farlary was most likely encouraged by low intensity but sustained grazing by domestic livestock as opposed to a climatic downturn and hydrological changes (Tipping et al, 2007).

Bradshaw (1993) suggested that the decline in *Pinus sylvestris* may be due to a reduction in fire regularity, which is necessary for the regeneration of pine seedlings and for the exclusion of pine competitors, therefore allowing pine to be the dominant taxon. However, Froyd (2006) presents data from four sites within Scotland to argue that there is no link between the fire regime and the abundance of scots pine. A relationship between mire and heath communities and fire regularity is more pronounced (figure 19). Although, it is unknown whether it is the expansion of such communities that encourage the rise in fire activity or the rise in fire activity which promote the spread of these communities (Froyd, 2006). The discussion as to whether the fire regime in Scotland during the Holocene was largely controlled by wildfires or anthropogenic, domestic scale burning is still ongoing.

2.9 Insects and Pathogenic Attack

Matias and Jump (2012) have suggested that during periods of increasingly warm conditions and in particular warm winters, the possibility of *Pinus sylvestris* becoming subject to attack from insects and pathogens rises markedly. *Thaumetopoea pityocampa* is a caterpillar that is responsible for a decrease in the growth and regeneration rates of pine in southern Spain by up to 50% (Hódar et al, 2003). Rising temperatures have allowed *T. pityocampa* to spread to higher altitudes, attacking pines that are already at their growing limits. Watts (1961) provides evidence to suggest that the decline of elm witnessed throughout northern Europe during the mid-Holocene was due to pathogenic attack; this decline can be seen clearly in all pollen diagrams within the taxon range. The decline of *Pinus sylvestris* is not as uniform and clear; indeed there is a distinct lack of evidence for the attack of pine in Scotland during the Holocene by insects and pathogens (Bennett, 1995). However, within the coming decades, the rising temperatures could encourage the expansion of *Thaumetopoea pityocampa* in higher altitudes and latitudes, therefore threatening the remaining and already vulnerable *Pinus sylvestris* populations further north (Dobbertin etal, 2007; Sikström et al, 2011; Stöcklin and Körner, 1999).

2.10 Mineral Deficiency and *Pinus sylvestris*

The role of mineral deficiency within a mire must also be considered when evaluating the growth and regeneration of *Pinus sylvestris.* This hypothesis remained mainly untested until Mighall et al (2004) investigated pine remains discovered at Cadogan's Bog in Southwest Ireland. The pine decline at Cadogen's Bog occurred in two phases, the final phase happening around 4160 ± 50 years BP. Mighall et al (2004) present evidence to suggest that the final fall in pine pollen percentages, to less than 3% TLP, coincide with a decrease in mire mineral content, with noticeable drops in potassium, phosphorus, calcium, magnesium, iron, zinc and sodium. However, there is also evidence to suggest that there was a coincident increase in mire wetness, shown by the rising level of water tolerant taxa. The change in hydrology hypothesis is the widely accepted consensus when it comes to the pine decline (Birks, 1975; Bennett, 1984; Bridge et al, 1990; Anderson, 1995). Mighall et al (2004) conclude that the most likely cause of the collapse in pine populations in this area of Southwest Ireland was due to a shift to a period of increased wetness (caused by a climatic downturn), combined with mineral deficiency. Indeed, McVean (1963) also suggests that low mineral levels can cause the reduced growth and collapse of pine trees. Although, the data collected by Mighall et al (2004) can be misleading. Some of the data suggests that the pine pollen percentage decreases while certain minerals (Ca, Cu, Mg, K and P) increase. This is also coincident with a sharp rise in *Sphagnum*. Certain elements in particular have a close relationship with *Sphagnum* mosses as their porous structure and high levels of palygalacturonic acids and cation exchange capacity attract the elements. As a consequence the peak in mineral content may be due to the high *Sphagnum* levels immediately local to the sampling site and therefore may not be an accurate interpretation of mineral bog status during that period (Mighall et al, 2004). Further studies into mineral content and pine growth are required to gain a more accurate understanding of this relationship.

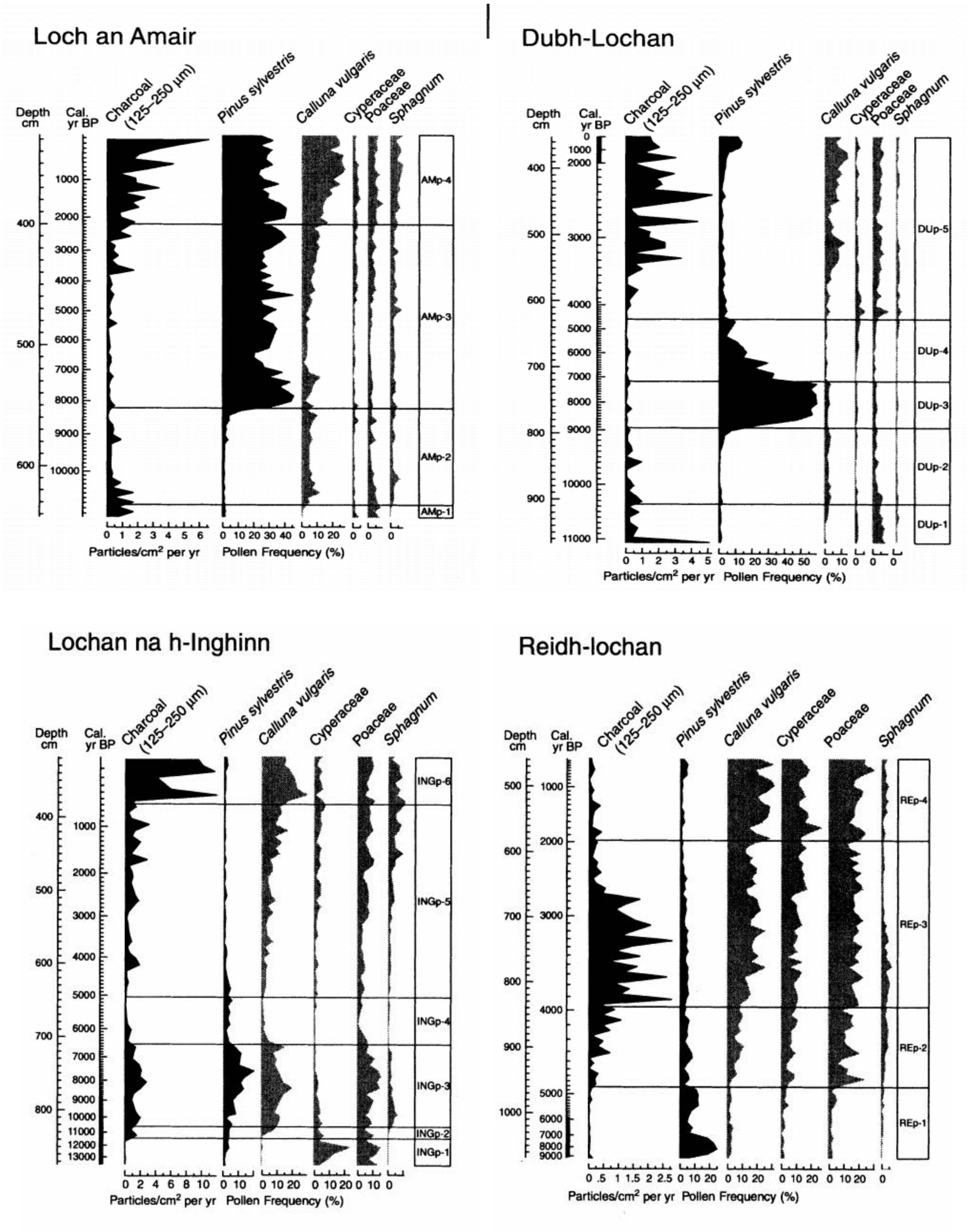


Figure 19. Diagrams showing the relationship between *Pinus sylvestris*, blanket peat vegetation and macroscopic charcoal at four sites in Scotland (Froyd, 2006).

Chapter 3: Research Area

3.1 Study Site - Meall Bad á Chrotha, Wester Ross, Northwest Scotland

The bog (NG 77251 72518) selected for investigation is on Meall Bad á Chrotha on the South Erradale Peninsula, Wester Ross, in the Northwest Scottish highlands (figure 20). 'Meall' refers to a 'large hill' or 'small mountain' in Gaelic. Meall Bad á Chrotha stands at 358ft (109m) above sea level and rises above the shores of Loch Cláir. The South Erradale peninsula lies between Loch Gairloch, to the North, and Loch Torridon, to the South. The peninsula is relatively open with vast areas of blanket bog and lochans dotted around, while the coastline is broken up by small fishing communities and isolated pockets of woodland to the North. Bordering the peninsula to the west is the Torridon mountain range that surrounds Loch Maree, dominated by Beinn Eighe sitting at over 1010 metres high (figure 20). Both the South Erradale and adjacent Applecross peninsula form part of the Inner Sound, a narrow stretch of sea dividing the peninsula's on the mainland from the Inner Hebridean Islands which include south Rona, Skye and Raasay.

Figure 20. Location of the South Erradale peninsula within Wester Ross and the Inner Sound. Also shown is the coring site, Meall Bad á Chrotha and the Applecross Peninsula (created using EDINA Digimap, 2014).

3.2 Current Vegetation

The dominant vegetation around the core location is *Sphagnum* moss with abundant *Calluna vulgaris*, *Eriophorum* and grasses present in the drier areas (figure 21). The South Erradale Peninsula itself is predominantly treeless with vast expanses of blanket peat and heather grassland covering much of the area (figure 22); however, immediately to the north and west there are pockets of woodland that remain (figure 23). Closest to the coring site around Loch Bad á Chrotha there are pockets of native upland birch; present in the wetter isolated river valleys where the conditions are more suitable. Just north of the peninsula a mix of woodland persists with oak woodland the most common around the coast, where as birch and native pine woodland are present in the upland areas. Pine woodland becomes more prevalent further inland towards Loch Maree, where the populations are internationally important due to their abundance and distinctiveness (Ratcliffe, 1977). The South Erradale Peninsula and the surrounding area are within the pine zone – the area of Scotland in which *Pinus sylvestris* is a native species. The majority of woodland in the area is 100% native; although, there are pockets of non-native species, the result of conifer plantations, which can be seen in figure 24 at only 35% nativeness.

3.3 Scotland's Present Forested Regions

In Scotland today, woodland fragments provide evidence for the four major forested regions (McVean and Ratcliffe, 1962). In the south and west, oak forest with birch and hazel is present, while central and eastern Scotland, pine forest with some birch and oak is characteristic. Birch forest is locally frequent in the in the north and west (Bonsall et al, 2002). In the exposed coastal areas in the far north and west, including the smaller Hebridean islands and Orkney and Shetland, forest is absent due to the extreme climatic conditions which make it unsuitable for tree growth (Shotton, 1977; Bonsall et al, 2002).

Figure 21. A photo taken at Meall Bad à Chrotha, the coring location, facing south.

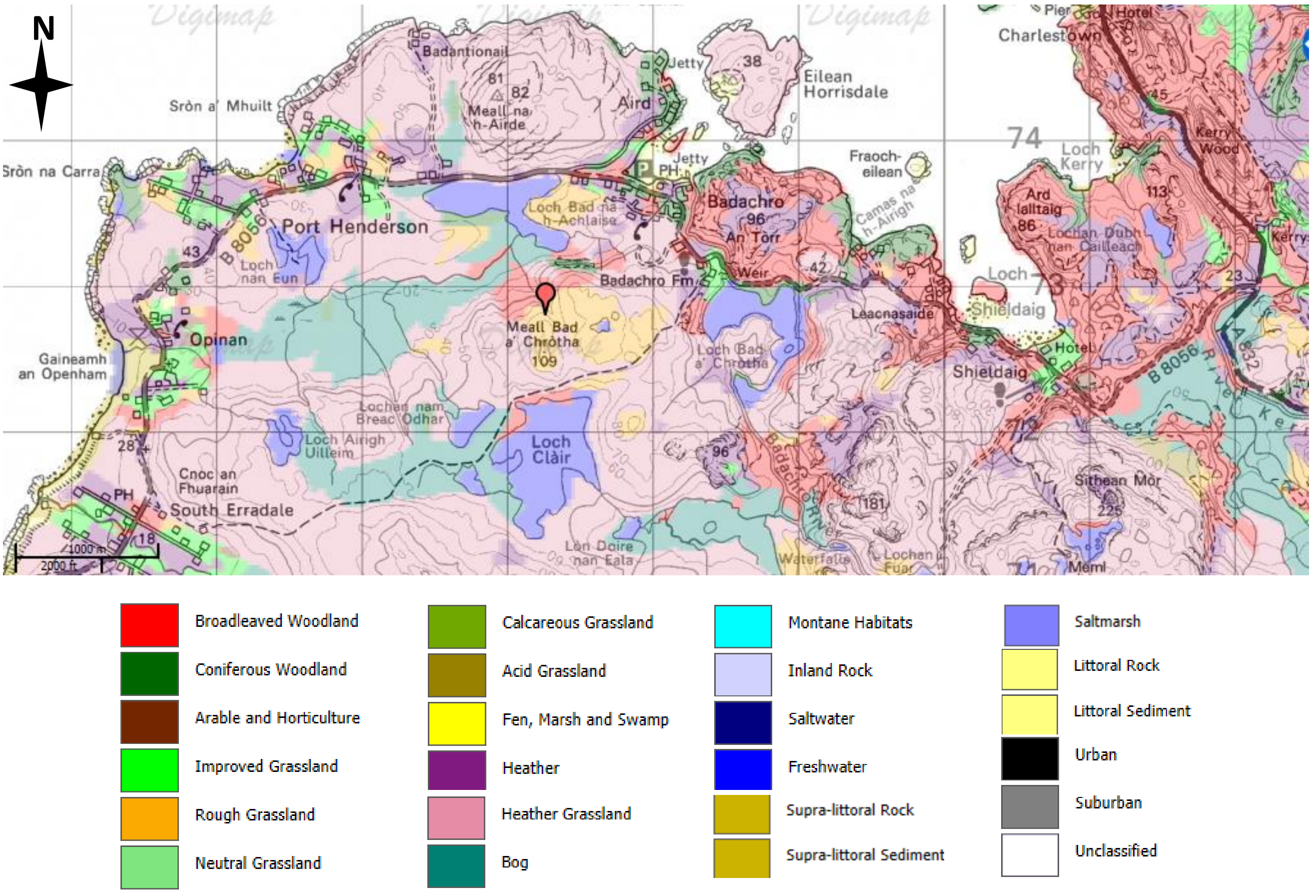


Figure 22. The different land uses, vegetation types, sediment and rock formations surrounding Meall Bad á Chrotha and the South Erradale Peninsula (created using EDINA Digimap, 2014).

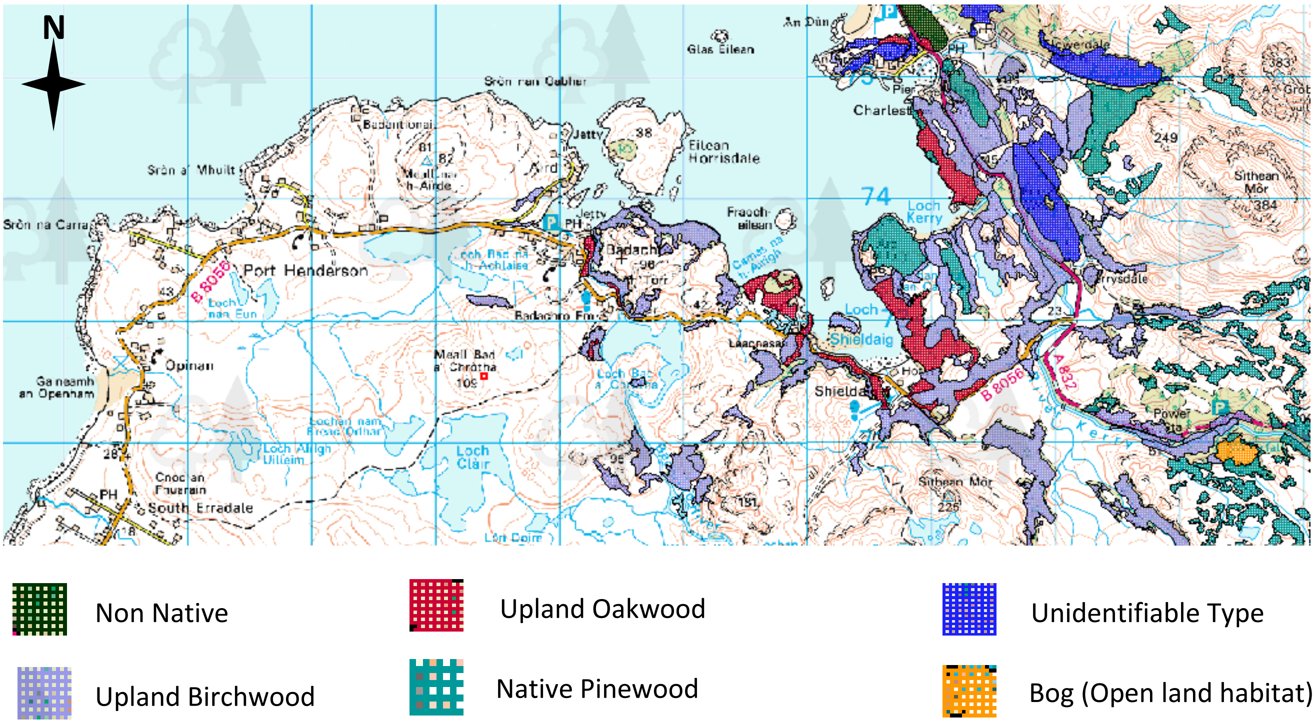


Figure 23. Habitat and woodland type around Meall Bad á Chrotha created using Native Woodland Survey of Scotland by the Forestry Commission (2014). All of the area shown is part of the 'Pine Zone' – the area where scots pine is native to the area.

Figure 24. The nativeness of woodlands around Meall Bad á Chrotha. The differing percentages indicate the nativeness of a particular woodland pocket. For example, 100% means that all trees in that pocket are native to the area (created using the Forestry Commission Scotland map viewer, 2014).

3.4 Geology and Relief

The underlying geological structure of Wester Ross has a pronounced effect on the relief of the region. The eastern part of the Wester Ross is predominantly an undulating peat-covered plateau about 300m above sea level; in comparison to the west which is characterised by numerous peaks reaching over 700m and a much more mountainous terrain (Daniell, 1997). This contrasting relief from the west to the east is due to the geological split that separates the region. Figure 25 clearly shows the separation of the west from the east by the Moine Thrust which stretches from Sleat on the Isle of Skye to Loch Eriboll on the northern coast. The west is predominantly Torridonian sandstones and grits, with Lewisian gneisses to the north and south. The Lewisian gneisses are responsible for the 'knob and lochan' features which characterise many landscapes within northern Scotland (Daniell, 1997). These 'knob and lochan' features reflect past glacial activity acting on and eroding specific geological formations. There are also small sections of quartzite and limestone further north which follow the Moine Thrust division. Moine Schists and granulites lie to the east (Daniell, 1997).

3.5 Climate

Northwest Scotland is particularly interesting in terms of climate due to its proximity to the North Atlantic Ocean. The location of the Oceanic Polar Front and the Gulf Stream are closely related to the climate oscillations in western Scotland (Ruddiman, 1977). The Oceanic Polar Front position is recognised by the joining of the cold, fresh currents from the Arctic Ocean and the warm, saline currents of the Gulf Stream, therefore how far north the Gulf Stream extends at any given time determines the position of the front (McCormack, 2011). Warm air masses move west over the North Atlantic, therefore Scotland is supplied with abundant precipitation, which between November and April tends to fall as snow (Benn, 1997). Northwest Scotland is currently characterized by a mild, maritime climate due to the Gulf Stream, with maximum sea surface temperatures reaching $4^{o}C$ in

January. The milder climate is maintained by the onshore Southwest winds and the oceans ability to retain heat (McCormack, 2011).

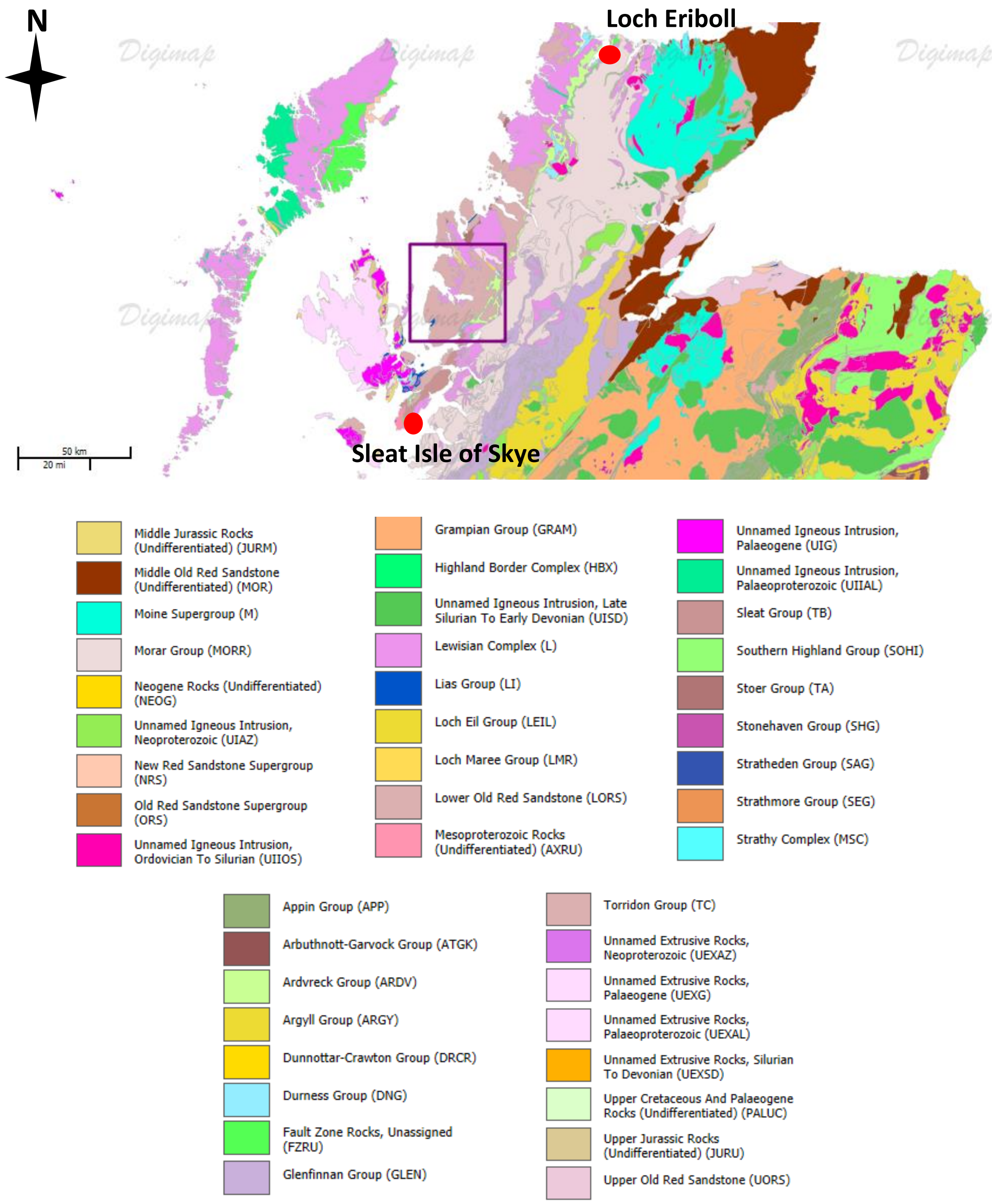


Figure 25. The geology and rock formations of Northern Scotland and Wester Ross. The South Erradale and Applecross peninsulas are highlighted by the purple grid. The Moine Thrust Belt is a 120 mile geological feature that trails from the north coast at Loch Eriboll to the Sleat Peninsula on the Isle of Skye (shown by the red dots) (created using EDINA Digimap, 2014).

Prevailing westerly airstreams create a pattern of precipitation from west to east. The coastal areas in the west, particularly Wester Ross, are affected the most as the air moisture level is at its greatest here resulting in high rainfall. Moisture levels in the air decrease as the distance the air travels increases; therefore, the east is markedly drier than the west (Lindsey et al, 1988). The mountain range that runs along the west coast has a further impact on precipitation levels, increasing them up to more than 2500mm; in comparison with the east with sees rainfall levels between 650 and 1000mm per year (Lindsey et al, 1988).

The climate of northern and Northwestern Scotland is suitable for the widespread development of blanket bog and deep peat. The growth of ombrotrophic bog across the region is encouraged by the favourable conditions such as high and regular precipitation, high atmospheric humidity, relatively cool mean temperature (the Northwest has a daily average temperature of 8.5^{o}C (figure 26)) and a small annual temperature range (Lindsey et al, 1988). The development of ombrotrophic bogs not only requires high levels of precipitation (figure 27) but regular precipitation as they receive all their nutrients and water from rainfall. Northwest Scotland has between 180 and 200 wet days per year (figure 28); the formation of blanket bog requires a threshold of more than 160 wet days per year (Lindsey et al, 1988). The decrease in precipitation levels moving eastwards across Scotland is evident in the distribution of peatland as the soils are predominantly mineral in type (figure 30). The development of blanket bog is further encouraged by increased levels of water surplus (the excess of rainfall over evapo-transpiration during the 6 months April to September). The Northwest has particularly high levels of water surplus (more than 20cm) in comparison to the east (less than 10cm) (figure 29). This combination of regular days with high levels of precipitation, water surplus and cool average temperature demonstrates the suitability of Northwestern Scotland and Wester Ross, to the development of blanket bog.

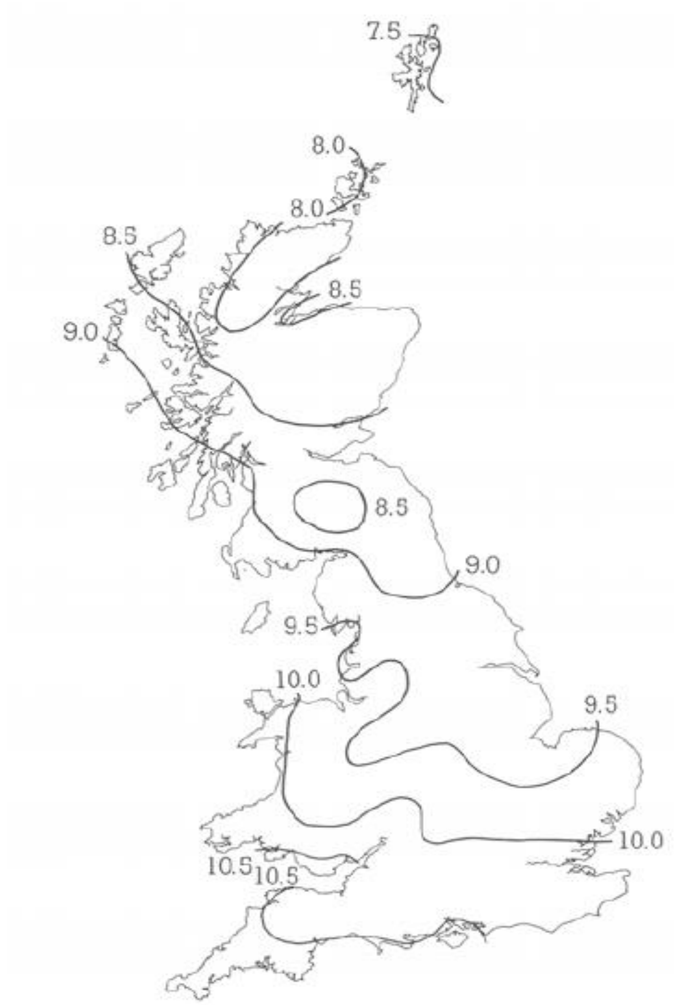


Figure 26. Average daily mean temperature in the British Isles (Metorological Office, 1952; Lindsey et al, 1988).

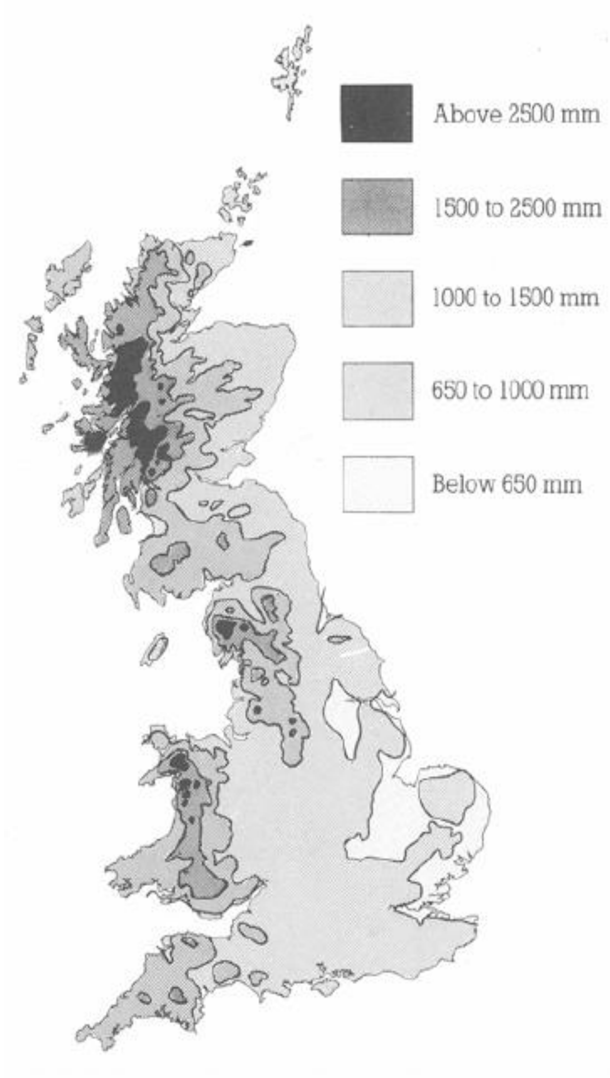


Figure 27. Average annual rainfall, after Climatological Atlas of the British Isles (Meteorological Office, 1952; Lindsey et al, 1988).

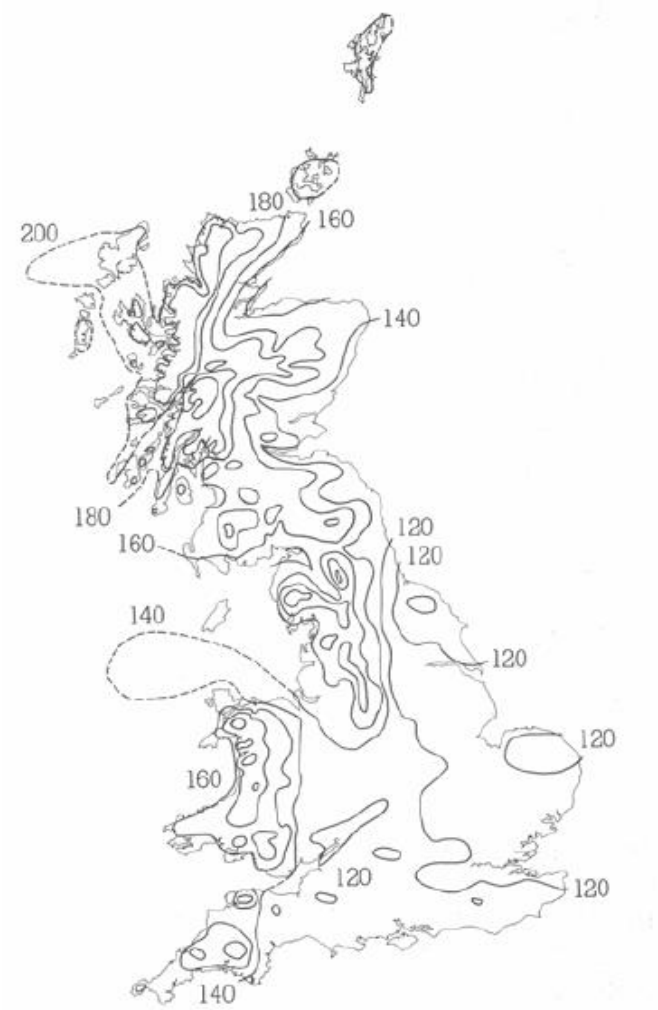


Figure 28. The distribution of wet days in the British Isles. If 1mm of precipitation is recorded with a period of 24 hours then it is classed as a wet day (Lindsey et al, 1988).

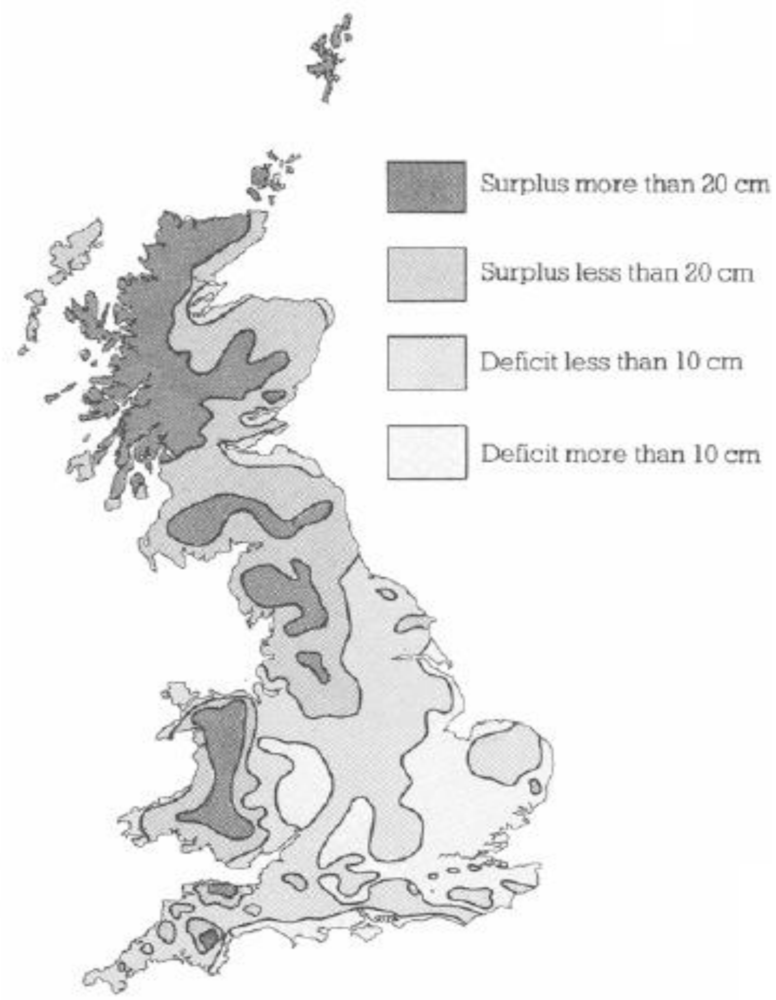


Figure 29. Levels of oceanicity across the British Isles from the west to the east. Difference between precipitation and evapo-transpiration for 6 months from April to September (Lindsey et al, 1988).

3.6 Peatland Distribution

The vast uninterrupted and undisturbed expanses of peatland in Scotland make the area perfect for palaeoecological research. 60% of the United Kingdom's peatland and 4% of Europe's peat carbon store is located in Scotland (UKCCC, 2011). The regions of Caithness and Sutherland are home to the Flow Country - the largest uninterrupted expanse of blanket bog in the British Isles. Blanket bog alone covers more than 20% of Scotland's land area, approximately 1.8 million hectares[4] (Marsden and Ebmeier, 2012; JNCC, 2011). In comparison to England, Wales, Ireland and the majority of Europe, Scotland is in the most suitable location for the formation of peaty or peat-topped soils.

The precise location of coring at Meall Bad á Chrotha was in a small valley bog. Small areas of blanket peat and valley bogs are characteristic of the west coast due to the Torridonian sandstones and grits and mountainous terrain. The 'knob and lochan' terrain is characterised by outcrops of bare rock, peats and thin soils (Daniell, 1997). In areas of poor drainage, in small valleys and hollows, deeper peats can form. The greatest extent of blanket peat lies to the east of the Moine Thrust; the cool climate, high precipitation, relatively low altitude along with the low mineral content and acidity of the underlying rocks create model conditions for the formation and growth of peat. In the Northwest, peaty soils are dominant with small isolated areas of peat between 50-100cm in depth (figure 30 and 31); however, in valley bogs such as Meall Bad á Chrotha and where peat cover is more extensive peat depth can be more than 100cm. Finding these areas with deeper peat depth in the Northwest in order to carry out palaeoecological investigations, in which peat depth of a few metres is needed, is very difficult; in comparison, the abundant blanket bogs of the north and Northeast are generally deeper than 100cm.

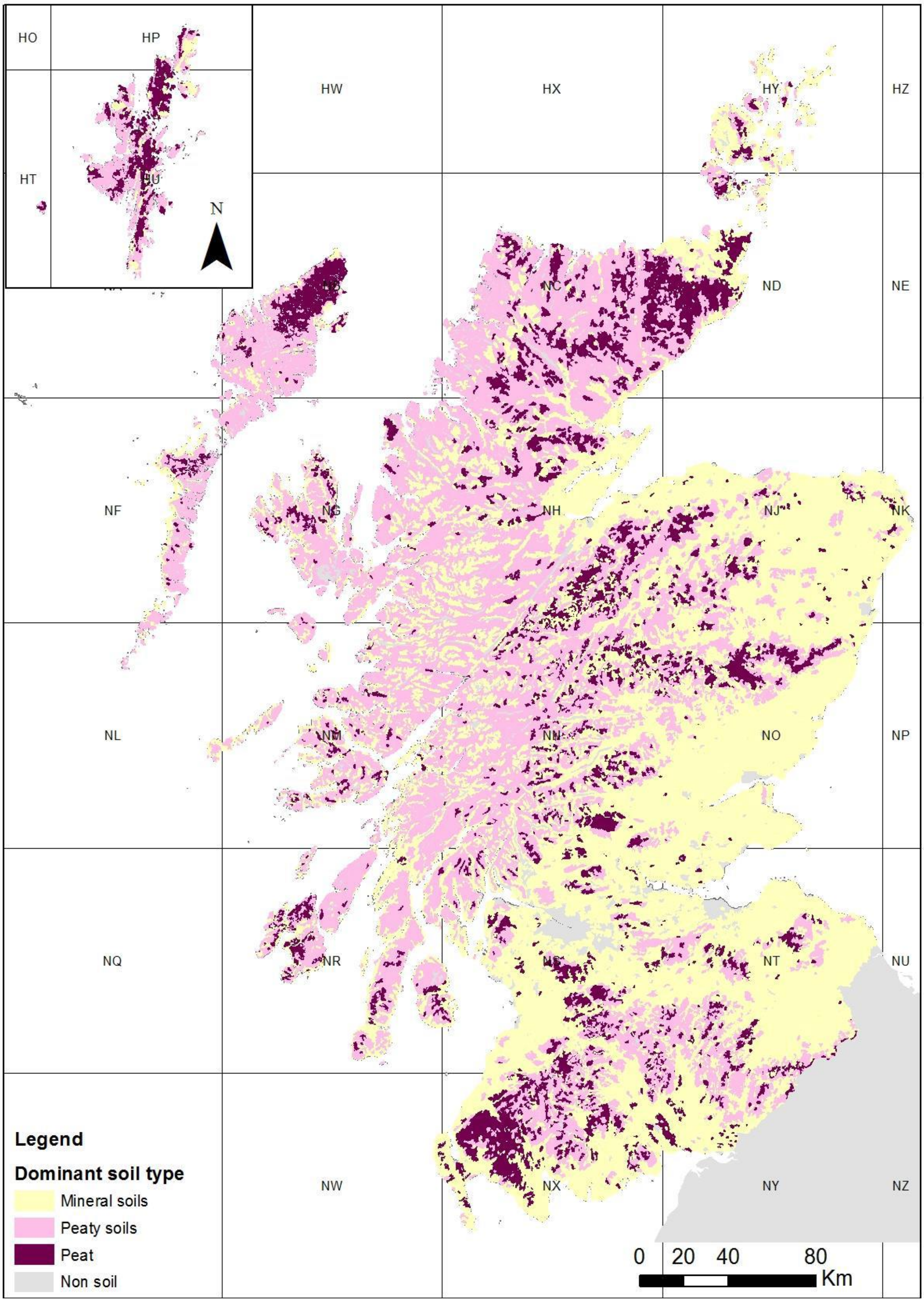


Figure 30. Dominant soil types in Scotland. Scottish Soil Map was used to identify the location of mineral soils, peaty soils, peat and non-soil (Bruneau and Johnson, 2014).

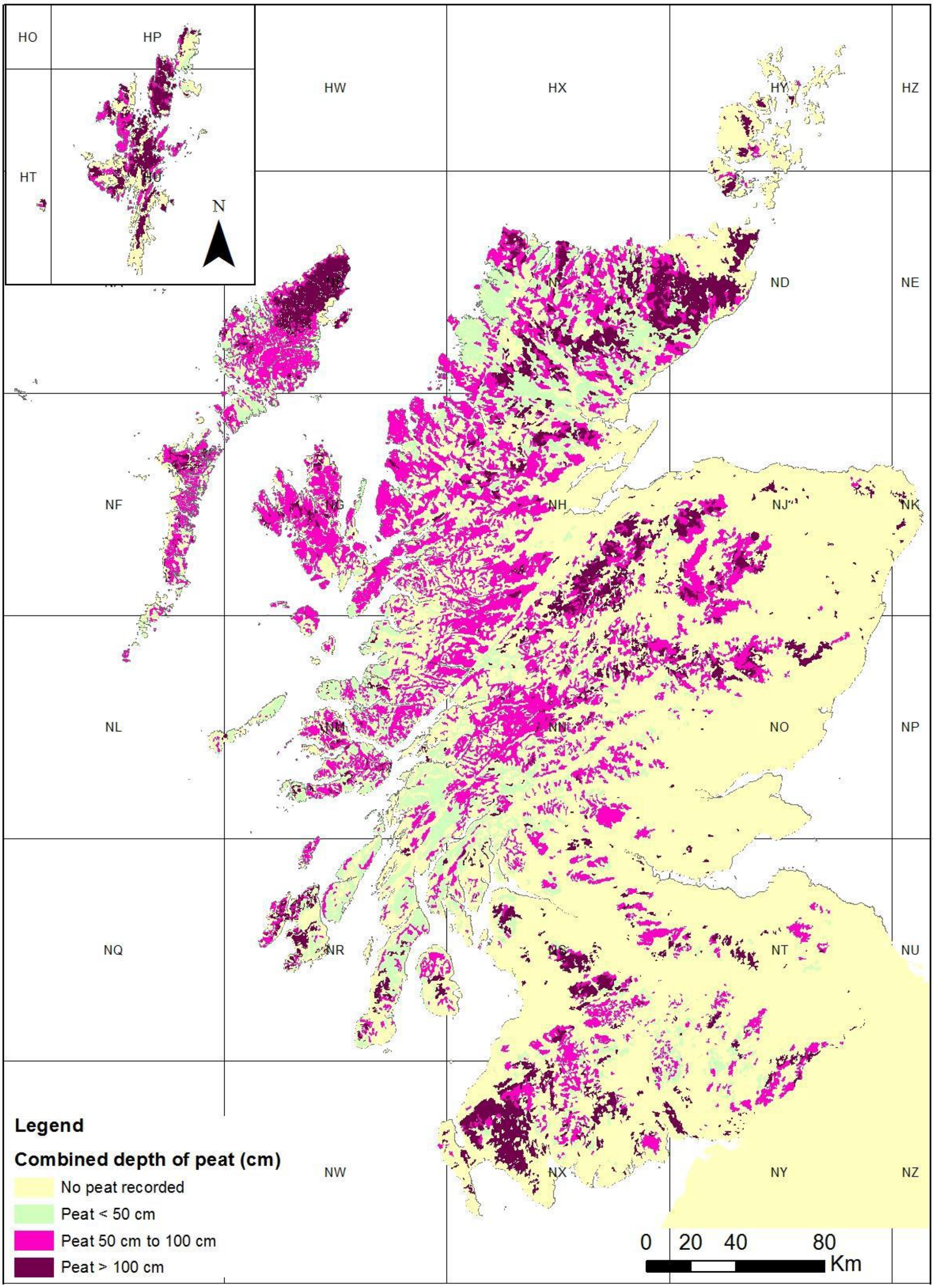


Figure 31. The depth of peat in Scotland. The Hutton peat depth database was used to identify the peat depths (Bruneau and Johnson, 2014).

3.7 Peatland Damage

Various land management strategies are responsible for the damage and loss of large areas of peatland throughout Great Britain and globally. In Scotland, substantial areas of peat have been extracted or cut for commercial and domestic use, predominantly being used for the production of compost. This removal of peat can lead to drying out, loss of vegetation and decomposition, therefore contributing to the peat loss even further. An estimated 5.5% of Scottish peatlands have been cut at some point in time whether it's by crofters for personal use or for commercial use by whisky distilleries who use peat during 'kilning' (Marsden and Ebmeier, 2012; Artz et al, 2012). For the last ten years, peat extraction in Scotland for commercial use has varied around 440,000m^3 (UK National Statistics). At Meall Bad á Chrotha there was no evidence to suggest peat cutting had taken place, most probably due to the inaccessibility of the site. Peat cutting usually takes place on lowland raised bogs as they are easier to access and the peat is more suitable to use for compost.

Afforestation is also a significant cause of peatland loss throughout the Great Britain and globally. Approximately 200,000 hectares of peatland in Great Britain has been lost due to afforestation (Anderson, 2010). The Forestry Commission was created after the First World War and was given incentives to plant forests; in particular, non-native conifers plantations. 1,516,000ha which makes up a total of 7% of land area has been planted with conifers in Britain; 993,000ha of this is in Scotland (Dumfies and Galloway Council). In order for plantations to be developed, the peatland has to be drained and aerated to allow the rapid growth and stability of the trees; thus destroying the peat and ceasing the absorption of CO_2 from the atmosphere (Hargreaves et al, 2003). Scottish peatlands alone hold an estimated 1,620 Mt of Carbon; therefore preserving these areas are vital. The large majority of woodlands in that area of Wester Ross are made up of native birch, oak and pine, however there are small pockets of non-native plantations as can be seen by the yellow section further north in figure 24 which is comprised of only 35% native species.

Grazing and trampling is also an issue with Scottish peatlands. When done intensively and continuously, they can have an impact on bog taxa and eventually cause the eroding of peat. Drainage for grazing and agricultural purposes results in the lowering of the water table and, therefore loss of taxa such as *Sphagnum* which contribute to peat formation (Marsden and Ebmeier, 2012). However, since the realisation of how vital peatlands are, in particular their importance as a Carbon store, restoration and preservation programmes have been put in place. The UK Biodiversity Action Plan aims to restore 600,000 hectares of Scotland's blanket bog; this could prevent 2.7Mt CO_2 from being released into the atmosphere each year (RSPB Scotland, 2011).

3.8 Human Activity in the Region of Meall Bad á Chrotha

Immediately around the study site there is little evidence prehistoric human activity. However, there have been archaeological sites identified from past work done along the Badachro River. Twenty-two sites in particular have been identified (figure 32). At Loch Bad á Chrotha there are remains of a roundhouse indicating prehistoric activity (Wentworth, 1989). Thirteen of the sites were dated to the Bronze or Iron Age and five were dated as Early Modern. The remainder of the sites were unable to be dated. All however would have been abandoned during the early 19th century (Dagg, 2010).

Modern day human activity on the South Erradale Peninsula is concentrated in the existing townships such as Opinan and Badachro or on the crofting and farming land surrounding these small, scattered settlements. Meall Bad á Chrotha overlooks Loch Cláir which lies on the path between Badachro and Red Point. Badachro, which is about half a mile away and the closest settlement to the coring location, is a very small collection of no more than a dozen houses or holiday lets and a public house. Red Point is another small scattered settlement on the opposite side of the Peninsula to Badachro. Three further scattered crofting communities can be found on the Peninsula, the fishing communities of Port Hendersen, Opinan and South Erradale.

The harsh climate keeps the population density low but the habitats on the South Erradale Peninsula and in Northern Scotland are under increasing pressure due to population growth and the growing demand for tourism. However, the majority of Northern Scotland remains inaccessible; coastal low lying areas and valleys are the only areas suitable for roads. Human activity and interference is evident immediately surrounding the townships and crofting epicentres. Sheep farming is an integral aspect of the economy in these rural areas as the rough terrain and inaccessibility of the area has little agricultural value, only suitable for rough grazing. Sheep farming does occur on the Peninsula but there is no evidence to suggest that sheep farming took place as high up as the study site. Between 1770 and 1870, sheep farming became the dominant land use just as the population in the highlands of Scotland peaked just before 1770AD, when there were many small settlements scattered across the region (Lindsey et al, 1988). Today, the main industries in the Northwest of Scotland are sheep farming, deer hunting, fishing and tourism (Gear, 1989). In recent years, conifer plantations have also became a dominant sector in the economic community and impacted heavily on large areas of the landscape.

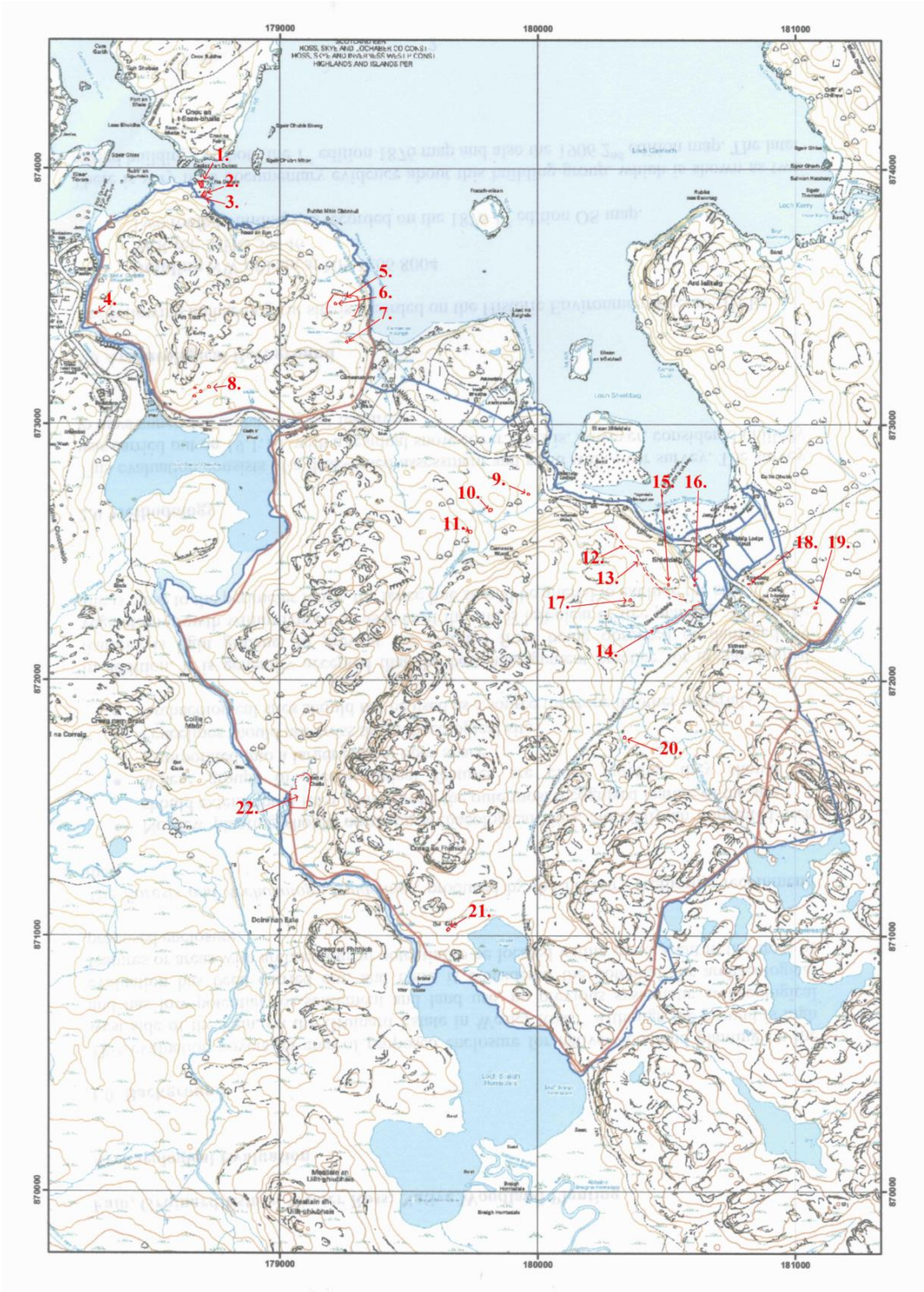


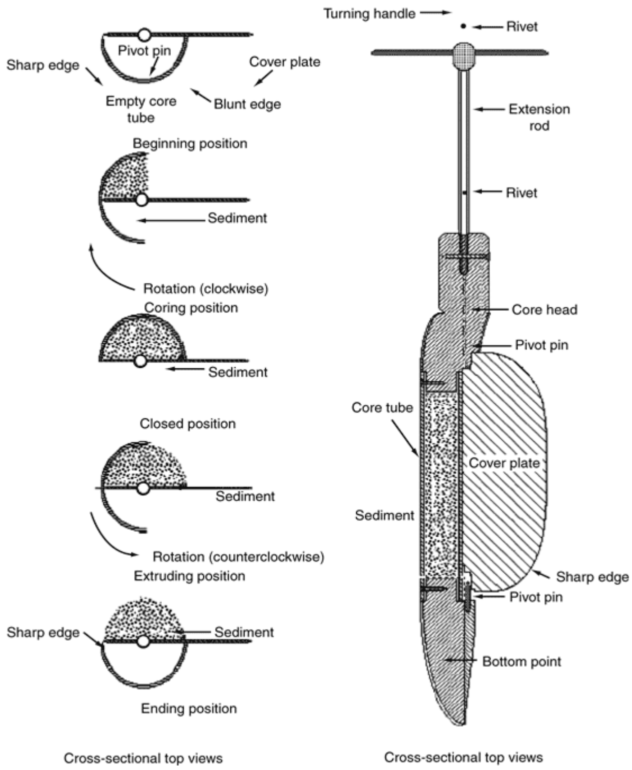
Figure 32. Twenty-two hut circles, some dated back to the Bronze Age, found along the Badachro River on the Northern side of the South Erradale peninsula. Site number 8 situated at Loch Bad á Chrotha is closest to Meall Bad á Chrotha (Dagg, 2010).

Chapter 4: Research Methods

4.1 Study Site Selection and Fieldwork

This study is focused at Meall Bad á Chrotha on the South Erradale Peninsula. The site chosen to collect the sample was partly dictated by the site itself and the need for an undisturbed area of bog which had suitable peat depth and was relatively accessible. After selecting a suitable site, successful coring is necessary to ensure a sediment or peat sequence as compete and undisturbed as possible (Berglund, 2003). A Russian corer was used to take the core as they are mainly used for soft sediments, such as peat. The corer was pushed down into the peat until the cylinder was covered before being turned clockwise to cut and collect the undisturbed sediment into the half cylinder (Fægri and Iversen, 1989). The cylinder is 50cm long, therefore this process had to be repeated 10 times to collect the 5 metre core. Each section of the core was then placed into plastic tubing and wrapped in plastic sheeting as it is important to prevent drying out of the sample as shrinkage can cause deformation of the pollen grains (Berglund, 2003). Samples were kept sealed to prevent water loss and the development of aerobic conditions, which increase the rate of decomposition. The core was then stored wet at <4°C before being sub-sampled.

Figure 33. Russian corer diagram (USEPA, 1999)



4.2 Troels-Smith Classification Scheme (1955)

The Troels-Smith Classification Scheme was used to identify three specific elements of each contrasting layer of the core:

1) Composition/Components
2) Degree of Humification
3) Physical properties of the sediment layer

This classification scheme was introduced by Troels-Smith in 1955 as a means of describing and analysing organic-rich wetland and lake deposits within the Northern Hemisphere. This was carried out within the laboratory in five stages using specific classification guidelines set out by Long et al (1991) (table 13).

Stage 1:

After identification of a section for analysis, the core was cut vertically and cleaned from top to bottom with a spatula to avoid contamination.

Stage 2:

The distinct stratigraphic layers were then identified and marked on the face of the core.

Stage 3:

The depth and boundary nature (gradual or sharp) was recorded (table 8).

Stage 4:

This stage involved the analysis of each sediment layer and the description of physical components. Each individual component within one layer is recorded on a scale of 1 (25%) to 4 (100%). For example, if 25% of the layer was *Turfa bryophytica* then it would be abbreviated as Tb^1 and if the remainder of the layer was *Detritus lignosus* then its abbreviation would be Dl^3 and so forth.

Humification analysis was carried out also using a five point scale. Liquid colour released through squeezing and the strength of the sediment were the two properties examined in order to determine the degree of humification (table 9). If the water squeezed from the sediment was clear then it would be given a score of 0, whereas the more turbid the liquid the higher the value given. By recording the degree of humification this allows the interpretation of how oxidized the sediment is, therefore, is an approximate indicator of water table depth.

Stage 5:

The physical properties of the layers were recorded during this stage. The darkness (*nigror)* (table 10), dryness (*siccitas*) (table 11), stratification (*stratification*) (table 12) and elasticity (*elasticity*) were all recorded using scales from 0 to 4. A colour Munsell chart was used to record the colour of the sediment and various other properties were also noted such as the overall structure of each layer e.g. fibrous, granular etc. Table 13 summarises the Troels-Smith (1955) classification scheme for the description and analysis of the physical components of sedimentary deposits.

Boundary nature and width (Lim)	
Lim 0	>1cm
Lim 1	<1cm - > 2mm
Lim 2	<2mm – >1mm
Lim 3	<1mm – >0.5mm
Lim 4	<0.5mm
Lim sup	Upper boundary of the deposit

Table 8. Boundary width and nature (Sharp or diffuse) within a deposit based on the Troels-Smith (1955) classification scheme.

Humification	
0	Fresh plant structure – releases clear water when squeezed
1	Well preserved – darker water released when squeezed loosing 25% of sediment
2	Partially decayed – 50% of sediment lost through squeezing
3	Decayed – 75% of sediment lost through squeezing
	Very unnoticeable or no plant structure – 100% off sediment lost through squeezing

Table 9. Degree of humification within a deposit based on the Troels-Smith (1955) classification scheme.

Darkness (Nig)	
Nig 0	Lightest shade that can occur within a sediment, e.g. marl, clear quartz
Nig 1	Light shades e.g. marl gyttja, calcium rich clay
Nig 2	Medium shades e.g. detritus gyttja, newly formed peat, weathered moraine clay
Nig 3	Dark shades e.g. coarse detritus gyttja, decaying peat
Nig 4	Darkest shade e.g. entirely decayed *Sphagnum* peat

Table 10. The degree of darkness within a deposit based on the Troels-Smith (1955) classification scheme.

Dryness (Sicc)	
Sicc 0	Water
Sicc 1	Entirely saturated
Sicc 2	Saturated
Sicc 3	No saturation
Sicc 4	Warm air dry

Table 11. Degree of dryness within a deposit based on the Troels-Smith (1955) classification scheme.

Stratification (Strf)	
Strf 0	Highest degree of stratification – complete homogeneity/ breaks with the same force in each direction
Strf 1	High
Strf 2	Medium
Strf 3	Low
Strf 4	No Stratification – separates very easily in thin horizontal layers

Table 12. Degree of Stratification within a deposit based on the Troels-Smith (1955) classificification scheme.

Troels-Smith Classification Scheme			
Component	**Name**	**Code**	**Sediment Nature**
Substantia	*Substantia humosa*	Sh	Humified organics beyond identification
Turfa	*Turfa herbaceae*	Th	Roots, stems, rhizomes of herbaceous plants
	Turfa lignose	Tl	Stumps, roots, trunks, twigs of ligneous plants
	Turfa bryophytica	Tb	Stems, leaves, rhizoids, protonema of mosses
Detritus	*Detritus herbosus*	Dh	Leaves, stem fragments from herbaceous plants more than 2mm
	Detritus lignosus	Dl	Bark, wood fragments from ligneous plants more than 2mm
	Detritus granosus	Dg	The humified remains from ligneous and herbaceous plants less than 2mm but more than 0.1mm
Limus	*Limus calcareus*	Lc	Calcuim carbonate or marl, unlike calcareous turfa is it relatively soft. Particles less than 0.1mm
	Limus detritousus	Ld	Detritus material from plants and animals less than 0.1mm
	Limus ferrugineus	Lf	Iron oxide less than 0.1mm
Grana	*Grana minora*	Gmin	Fine, medium or course particles between 0.06 to 2.0mm
	Grana majora	Gmaj	Fine, medium or course particles between 2 to 60mm
Argilla	*Argilla steatodes*	As	Clay particles less than 0.002mm
Argilla	*Argilla granosa*	Ag	Silt particles 0.06 to 0.002mm
Anthrax	Charcoal	Anth	Particles of charcoal
Stirpes	Tree trunks	Stirp	Tree trunks
Stratum confusum	Disturbed stratum	Sc	Disturbed sediment

Table 13. Troels-Smith (1955) classification for the description and analysis of the physical components of sedimentary deposits.

4.3 Sub-Sampling

Pollen analysis requires a sample size of 1cm^3. A sample was taken every 10cm making a total of 50 samples. Taking a sample every 10cm would guarantee the recording of any shorter term, as well as longer term, vegetation and charcoal changes. The face of each section of core was scrapped before the sample was taken from the centre to reduce the risk of contamination. Each sub sample was covered in distilled water to prevent drying out, while waiting for the laboratory procedures to begin.

4.4 Laboratory Techniques

The laboratory techniques and pollen preparation followed Berglund (2003).

1) Evacuation of Alkali-Soluble Organic Compounds

Potassium hydroxide (KOH) was added to each sample before being heated in a water bath for 30 minutes. The samples were then decanted through a 180 micron sieve and the residue was washed. The residue was then centrifuged. This process of decanting and washing was repeated until the supernatant liquid was unstained.

2) Evacuation of unaltered Lignin and Cellulose

Glacial acetic acid (CH_3COOH) was then added to each sample before the samples were stirred, centrifuged and decanted. The acetylation mixture (1:9 conc. Sulphuric acid (H_2SO_4) – acetic anhydride ($(CH_3CO_2)_2O$)) was then added to each sample before being stirred. The removal of cellulose was carried out with extra care and safety as the acetoylsis mixture reacts violently with water (Allen, 2007); therefore, it was essential that all water was removed before the mixture was added. Once the mixture was added the samples were placed in a boiling water bath for 1 minute and topped up with glacial acetic acid. The samples were then centrifuged and decanted, before more glacial acetic acid was added and the samples went through the centrifuge and decanting process again. Distilled water was then used to top up each sample before being put through the centrifuge and then decanted; this stage was repeated twice.

3) Staining and Mounting

The samples were then washed with ethanol (C_2H_6O) to make sure all the water had been removed. The samples then went in the centrifuge before being decanted; this was repeated twice. 2mls of tertiary butyl alcohol ($C_4H_{10}O$) and 2 drops of safranin ($C_{20}H_{19}ClN_4$) were then added before the samples were transferred into small vials, centrifuged and decanted. Silicon fluid was then added; the amount of silicon fluid added equalled the volume of sample; this preserves the pollen grains. Each sample was then stirred and plugged with cotton wool until they were ready for mounting onto slides in preparation for analysis.

4.5 Analysis and Identification of Samples

The slides were then analysed using a magnification of 10x for the overview and 40x for counting and identifying the pollen to the smallest possible taxon. A number of reference books such as Fægri and Iversen (1989) were used to identify the different pollen types as well as reference material held at the Department of Geography, Durham University. A minimum of 300 pollen grains were counted at each depth to ensure the counts were statistically representative. If there was not 300 grains present on the slide, then further slides would be made up and counted until the minimum requirement of grains was reached.

4.6 Charcoal Analyses

The point count method (Clark, 1982) was used as this is the quickest and easiest approach as the charcoal is counted on the pollen slides without additional preparation. The number or area of charcoal particles were calculated along a series of transverses or on a grid. On a microscope, an appropriate grid may be obtained by using an eyepiece reticle with an array of points and moving the field of view on transects across the slide, recording the points falling on charcoal (Clark, 1982).

4.7 Pollen Percentage Diagram

The most reliable and accurate way to express the pollen data is in percentages (Berglund, 2003). These percentages were then plotted against depth in a pollen percentage diagram. The number of pollen grains recorded for each plant type were converted into a percentage of the total number of tree, shrub and herb pollen grains recorded in a single slide. These plant groups made up the “Total Land Pollen”. The diagram was created using TILIA software which allows the analysis and visualisation of ecological and palaenvironmental data (Juggins, 2007). Stratigraphically constrained cluster analysis in CONISS was used to numerically determine the different pollen zones (Grimm, 1991). The absolute abundance of each pollen type is usually calculated with reference to the number of exotic spores (*Lycopodium*) encountered in the routine pollen count (Moore et al, 1991), however due to the unavailability of *Lycopodium* tablets this could not be done. *Sphagnum* and *Myriophyllum alterniflorum* were not included in with the Total Land Pollen sum as their large abundances can depress the percentages of the other taxa. Taxa which contributed less than 1% to the total land pollen percentage were excluded from the pollen diagram, unless they were a rare taxon. This prevents the diagram from becoming overly ‘busy’ or ‘cluttered’ which would take focus away from the more dominant taxa. However, all pollen found, including taxa which contributed less than 1% to the total land pollen percentage, were included in the total count.

4.8 Non-Pollen Palynomorphs

Non-pollen palynomorphs, such as fungi, algae and plant animal fragments, have become increasingly popular in reconstructing past climate, as they can add an extra dimension to any pollen data and they do not need any additional preparation as they are present on the pollen slides. Microfossils can prove to be vital indicators of any climate or hydrological change; in particular microfossils can help determine any wet or dry shifts for example a shift to wet mire conditions can be seen at Glen Torridon by a sharp increase in the level of Copepoda (water flea) spermatophores (Anderson, 1996). However, identifying non-palynomorphs is problematic making it difficult to apply the indicator species approach (Montoya et al, 2010; Birks and

Birks, 1990). Although, every time a non-pollen palynomorph is identified it is added to a classification system and environmental interpretation is made using information from past studies (Montoya et al, 2010). Initially, fungi and algae were going to be identified and counted, however once the slides were prepared and analysis began the level of both fungi and algae were very low, therefore it was decided to focus solely on identifying and counting the pollen.

4.9 Pollen – Climate Transfer Functions

The C2 package (Juggins, 2007) was used to produce palaeoclimatic reconstructions. All taxa have a specific optimal preference within each environmental variable. If any environmental variable was to fluctuate outside the optimal preference then the taxa could not survive (Shennan, 2012). Through the use of transfer functions, past environmental variables, such as temperature, can be reconstructed using fossil pollen assemblages. A database containing 2696 Eurasia surface pollen samples was provided for use by the geography department at Durham University. Before the reconstruction could be carried out the training set required trimming. Samples were excluded from the training set using both representational and visual exclusion. Representational exclusion involves the exclusion of samples which do not represent past climatic conditions within a specified distance from the research area. Sites with altitudes above 1000 metres were also excluded. A number of models were run and tested using the training set and linear regressions were produced for the estimated values and residuals against the observations (figures 34, 35 and 36 show a few of the models produced). This testing revealed numerous samples that were 'outliers' therefore, were excluded through visual representation (figure 35). Repeating this process of running and testing models and removing outliers improved the statistical significance of r^2 and RMSEP eventually creating the model required for reconstruction. Edwards et al (2004) argue that having a broader dataset encompassing many contrasting environments will increase the predictive ability of the modern training set; however, it will reduce the precision of the reconstruction due to the introduction of noise. Once a suitable model with an improved statistical performance (figure 36) was attained (a suitable model has a good linear regression, high bootstrapped r^2 and low bootstrapped RMSEP (table 14) it was then run against the fossil assemblage from Meall Bad à Chrotha to allow the reconstruction of the mean temperature of the warmest month.

However, there are disadvantages when it comes to using fossil pollen assemblages to reconstruct past climate as pollen-climate transfer functions can prove to be very problematic and do not always work. The indicator-species, assemblage and the multivariate transfer function approaches are the three key approaches in any quantitative reconstruction of past climate (Birks, 1981, 2003). These approaches are based on one basic assumption – Methodological Uniformitarianism. This is the assumption that modern-day relationships and observations can be used as a model for past conditions and that organism-environment relationships, such as ecological tolerances, niches and climatic preferences of certain taxa, have not varied over time, at least, specifically, during the Late Quaternary (Birks, 2007). Any reconstructions and models that are based on assumption come with the risk that the assumptions are incorrect, therefore increasing their unreliability. There is the possibility that human interference has reduced the reliability of climatic inference from the pollen assemblages within the Eurasia surface pollen dataset (Birks et al, 2010). There is also the possibility of spatial autocorrelation error and multimodality within the model due to the close proximity of some sites in the dataset meaning they will exhibit many similar variables (Huntley, 2012). In order to minimise this problem bootstrapping was used to derivate sample-specific error estimates. However, despite there being many problems involved with pollen-based climate transfer functions, when they are compared to other independent reconstructions of climate they largely agree and show consistency, on millennial time scales at least (Birks and Seppa, 2004).

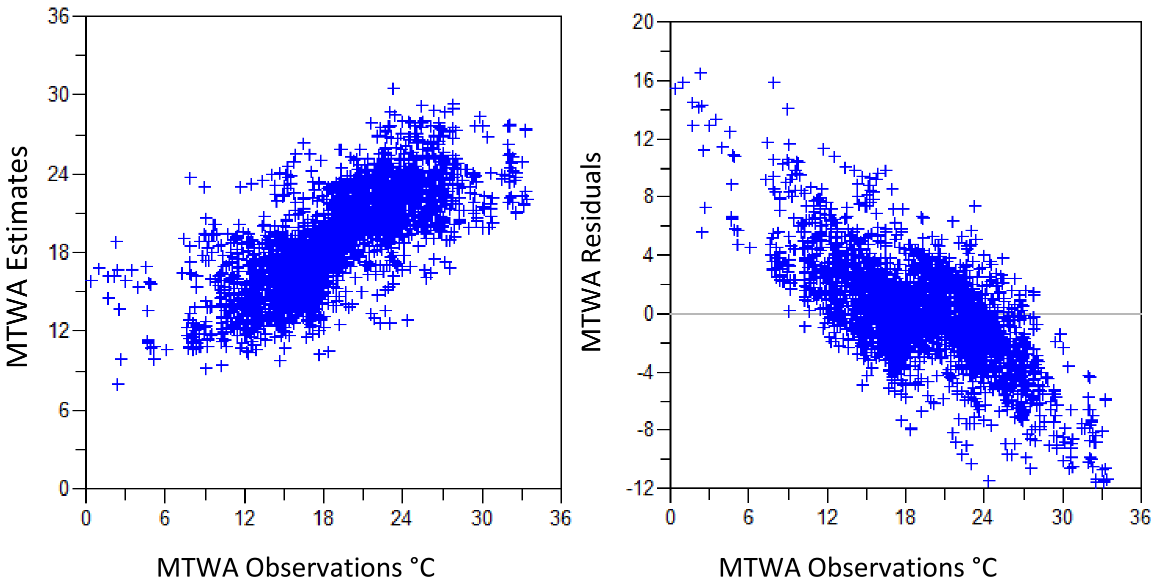


Figure 34. Estimates and residuals models before sample exclusion. This model has all 2969 samples included from the modern pollen-climate training set. Using ‘representational exclusion’ samples were excluded on the basis that they did not represent the palaeoclimate of Meall Bad á Chrotha. All sites with an altitude above 1000 metres and a latitude and longitude outside a suitable range were excluded.

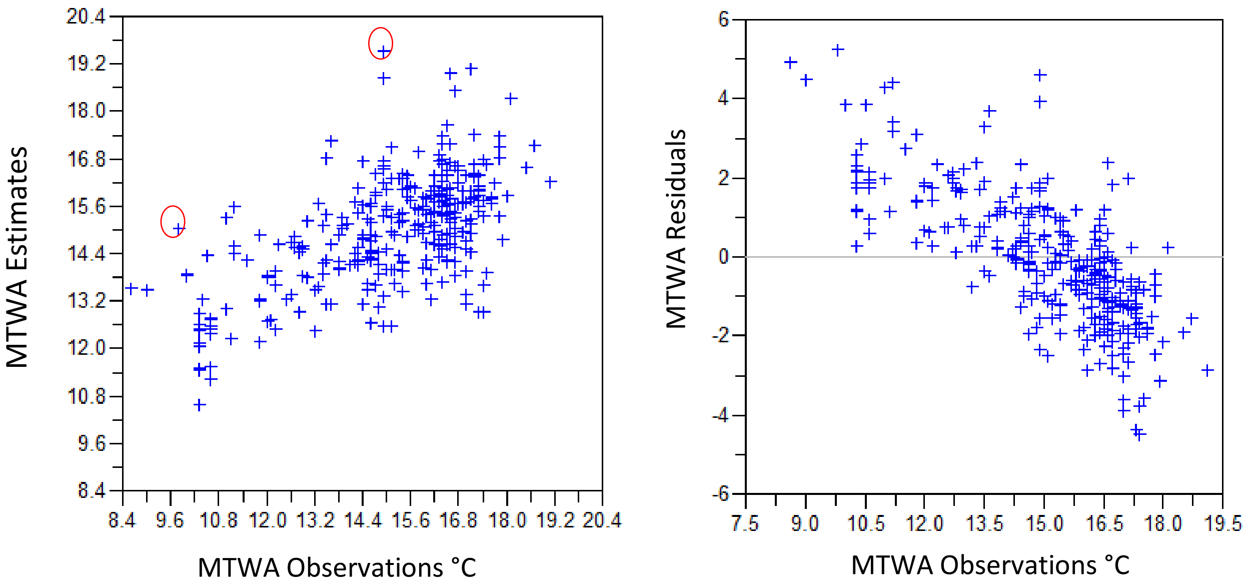


Figure 35. Estimates and residual models after sample exclusion. 316 samples were included in this model. Samples were now excluded through visual means in order to improve the RMSEP and r^2 performance statistics. The samples circled in red are examples of ‘outliers’ which were excluded before running the final model.

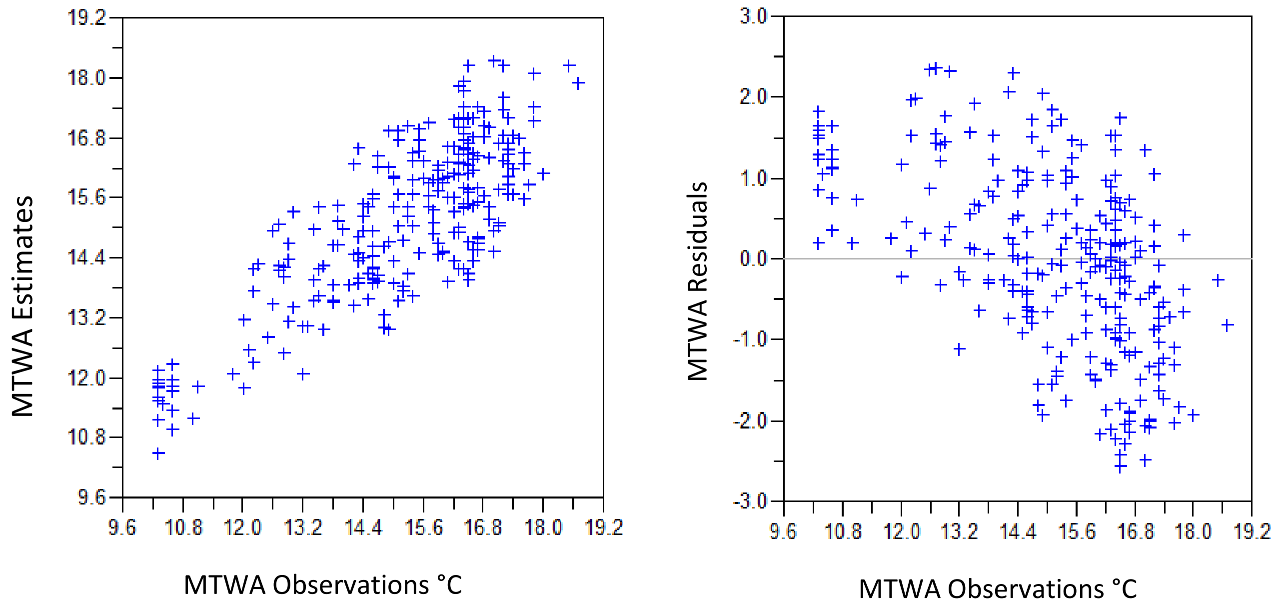


Figure 36. The final transfer function model which was used for the reconstruction. 257 samples were included after both representational and visual exclusion of sites. There is a marked improvement in the linear distribution of observed values plotted against predicted values. This model also had the lowest bootstrapped RMSEP and highest bootstrapped r^2.

Number of Samples	Exclusion Method	RMSEP	r^2
2969	Representational	3.3023	0.56933
316	Visual	1.6159	0.42041
257	Visual	1.1028	0.68866

Table 14. Performance statistics before and after sample exclusion from the modern pollen-climate training set. Samples were excluded on the basis that they did not represent the palaeoclimate of the research area and were out of the suitable ecological and altitude range. Samples were also excluded through visually identifying outliers in the previous models. Weighted-averaging partial least squares (WA-PLS) regression was used to run and test the models. An ideal model has a low bootstrapped root mean square error of prediction (RMSEP) and a high bootstrapped coefficient of determination (r^2). Statistical parameters produced during regression and calibration includes r^2 that measures the strength of a relationship between the observed and inferred values (Birks, 1995). RMSEP measures the predictive abilities of the training set and is calculated using bootstrapping (Birks, 1995). The table shows how the removal of outliers identified through the running and testing of previous models has improved model performance by lowering the RMSEP value and increasing the r^2 value. These performance statistics are based on component 3 which was used for the reconstruction as it showed the best RMSEP, r^2 value and linear regression.

4.10 Chronology and Dating

The core from Meall Bad á Chrotha has not been radiocarbon dated due to a number of restraints; however, through correlation with other well dated records from numerous authors, a chronology has been developed. Although, when any core has not been radiocarbon dated caution must be taken when correlating similar events from one profile to another (explained in Disadvantages section).

To make the correlation of events between contrasting investigations possible some dates were converted into 'calendar' or 'calibrated' years BP using the Fairbank0107 calibration curve which can be found in the appendices (Fairbanks et al, 2005). However, in some cases there was not an error provided with the radiocarbon date meaning the dates could not be converted.

4.11 Advantages and Disadvantages of Methodology

4.11.1 Advantages

There are many reasons why pollen analysis has become one of the most widely applied techniques in palaeoenvironmental reconstructions. Pollen grains are extremely resilient and can be found in deposits in which other types of fossils have been diagenetically destroyed. Pollen evidence from the late glacial 14,000-10,000 years ago survives in generally good condition (Roe et al, 1991). As well as being produced and retrieved in enormous numbers, pollen grains are also more widely and evenly distributed than larger (sub) fossils and are therefore less dependent on their mother plants having been members of the community forming the deposit (Fægri and Iversen, 1989). By defining the tolerances and climatic preferences of the taxa a reconstruction of past climates and any hydrological changes can be inferred from the pollen-analytic information (Fægri and Iversen, 1989).

Pollen analysis also allows for the reconstruction of past anthropogenic activity. The identification of certain pollen types, such as *Plantago lanceolata,* which is a weed commonly found on cultivated land, can indicate the production of crops and subsequent land use. The introduction or intensification of grazing animals can also be identified through specific vegetation changes. There is also no extra preparation needed for the analysis of charcoal once the pollen preparation has been complete. Charcoal is counted on the same slide as the pollen and is another indicator of anthropogenic activity.

One of the principal aims of this research is to reconstruct the history of *Pinus sylvestris*, in particular the pine decline. One advantage of pine pollen is that it is very distinct from other types of pollen therefore its identification is very straight forward.

4.11.2 Disadvantages

It has been argued that pollen analysis as a single proxy is unreliable as the vegetation lags behind in its response to climate change (Committee on Surface Temperature Reconstruction for the Last 2,000 years, 2006), therefore a multi-proxy approach is preferable, to allow a comparison between proxies and to increase reliability and accuracy. Although, in practice, a multi-proxy approach is rarely executed due to the time, funding and organisation required (Blundell et al, 2005).

A multi-proxy approach is also preferred due to the risk of contamination, which is the primary cause for any uncertainty in any palaeoecological studies (De Vleeschouwer et al, 2010). During collection and preparation of samples cleaning is crucial and the outer section of the core must be disregarded in order to minimise the risk of contamination.

A continuing debate is the use of palynology in reconstructing anthropogenic activity. It has been suggested that this methodology is not sensitive enough to identify any small scale human disturbances and land use such as isolated clearings, which may well have been a common occurrence in many landscapes during the Holocene (Davies and Tipping, 1997; Halstead, 2000; Sugita et al, 1997; Willis and Bennett, 1994). It was typical of farming communities during the Holocene to use well drained soil to produce crops or allow the grazing of domestic animals. This type of land is often some distance from the wetland deposits needed for pollen preservation and palynological investigations (Tipping et al, 2009).

Chronologies that have not been subject to radiocarbon dating produce a further problem. The assumption that dates from similar chronologies from within the research area can be transferred into an undated assemblage must be used with caution. There are many factors that can influence local catchments such as hydrology, human interference, topography and soil type. These factors can influence vegetation growth and contraction, such as the pine expansion and decline, at contrasting rates therefore, even sites local to each other can have varying dates.

In comparison to pine pollen which is very easy to identify, *Corylus* and *Myrica* are very difficult to distinguish between. One of the main challenges faced when pollen counting is indeed trying to distinguish between near identical pollen grains which are in fact different taxa (Bradley, 1957). Folding and deterioration of pollen grains also contribute to the difficulty involved with pollen identification. *Corylus* and *Myrica* show very minor structural differences which are not always obvious when using light microscopy, making is sometimes unfeasible to distinguish between the two taxa (Edwards, 1981). It is quite common in palaececological studies that *Corylus* and *Myrica* are combined to make a *Corylus/Myrica* curve; wherever the curve shows a significant exaggeration then it is most likely due to the inclusion of *Myrica* at that particular point in the pollen diagram. A combined curved has been used in this study. *Empetrum* and *Calluna* also have a combined curve (Ericaceae) in this study even though they are easily identifiable. This is due to human error and there not being enough time to correct it once the error had been identified. As a result, different profiles for varying authors within Northwest Scotland have been used in identifying which taxon would have more likely been growing at a particular time at Meall Bad á Chrotha.

Pollen is easily transported over long distances by various means, mainly wind and water therefore some of the pollen recorded in an investigation may be not indicative of the local vegetation (Bennett, 1984). Furthermore, different taxa produce different amounts of pollen. For example, an indicator for local growth of *Pinus sylvestris* has been established due to its high pollen production. Once pine is over 20% TLP then it is highly probable that its source was immediately local, whereas anything under 20% TLP can be assumed that the pollen travelled from somewhere else within the catchment (Bennett, 1984).

There is also the possibility that pollen has been downwashed through the deposit by falling water which could lead to misinterpretation of the vegetational record. However, the theory of pollen downwash has been strongly disregarded by Erdtman (1922) who investigated pollen content at different depths. The investigation showed that pollen content was similar throughout the profile therefore conflicting with the suggestion of pollen downwash.

Chapter 5: Results

5.1 Troels-Smith Classification

The Troels-Smith Classification Scheme was used to identify three specific elements of each contrasting layer of the core:

1) Composition/Components
2) Degree of Humification
3) Physical properties of the sediment layer

Depth (cm)	Unit	Description
0-73	9	Tb^2, Th^2, nig 3, strf 0, elas 0, sicc 2, lim 0, humification 1
73-142	8	Sh^1, Tb^3, nig 3+, strf 0, elas 0, sicc 2, lim 0, humification 1
142-325	7	Th^2, Sh^2, nig 4, strf 0, elas 0, sicc 2, lim 0, humification 1
325-370	6	Sh^3, Th^1, Dl+, nig 4, strf 0, elas 0, sicc 2, lim 0, humification 1
370-448	5	Tb^1, Sh^3, nig 4, strf 0, elas 0, sicc 2, lim 0, humification 1
448-475	4	Th^1, Tb^2, Sh^1, nig 4, strf 0, elas 0, sicc 2, lim 1, humification 1
475-485	3	Ld^3, Tb^1, Dl+, nig 4, strf 0, elas 1, sicc 2, lim 1, humification 1
485-495	2	Ld^4, Th+, nig 3, strf 0, elas 1, sicc 2, lim 1, humification 1
495-500	1	Ld^4, Th+, nig 4, strf 0, elas 1, sicc 2, lim 1, humification 1

Table 15. Troels-Smith classification from Meall Bad à Chrotha core. See tables 8-13 for classification description meanings.

Unit 1: 495 – 500cm

Fine detritus organic mud (particles <0.1mm) is the dominant sediment type at the base of the core with *Limus detritousus* making up nearly 100%, further suggesting that a lake was present at Meall Bad à Chrotha at this time. There are a few roots and stems of herbaceous plants present (*Turfa herbacea*+).

Unit 2: 485 – 495cm

Limus detritousus continues to make up nearly 100% of the sediment similarly to unit 1. Herbaceous plants (*Turfa herbacea*+) would have been present around the lake.

Unit 3: 475 – 485cm

Lake mud (*Limus detritousus*) makes up 75% of the sediment type suggesting sediment accumulation and vegetation were beginning to infill the water body. An increase in *Sphagnum* mosses is evident as *Turfa bryophytica* increases to 25%. There is also wood and bark fragments present (*Detritus lignosus*+).

Unit 4: 448 – 475cm

The percentage of humified organics beyond identification (*Substantia humosa*) rises to 25%; while another 25% is made up of the roots, stems and rhizomes of herbaceous plants as there is a transition to drier conditions and terrestrialisation continues. Half of the sediment is *Turfa bryophytica*, of predominantly *Sphagnum* nature indicating that there was peat-formation taking place. Wood and bark fragments can be seen.

Unit 5: 370 – 448cm

Substantia humosa is the dominant sediment type at 75%, with *Turfa bryophytica* declining down to 25% with fewer *Sphagnum moss* roots and stems present suggesting a reduction is peat-formation conditions perhaps due to a decline in bog surface wetness.

Unit 6: 325 – 370cm

The level of humified organics beyond identification (*Substantia humosa*) make up 75% of the sediment type. There is a transition from a more *Sphagnum* composition to a more herbaceous composition with *Turfa herbacea* making up the remaining 25%. There are detrital fragments of wood and bark (>2mm) (*Detritus lignosus*+) present within the sediment.

Unit 7: 142 – 325cm

There is a decline in the level of humified organics beyond identification as *Substantia humosa* falls to make up only half of the sediment type. *Turfa herbacea* makes up the other half with roots and stems of herbaceous plants present. At 295cm a small fragment of wood can be seen.

Unit 8: 73 – 142cm

This unit is well humified, with a transition from a more herbaceous nature to *Turfa bryophytica* which makes up 75% of the sediment indicating peat-formation. The remaining 25% consists of humified organics beyond identification, *Substantia humosa.*

Unit 9: 0 – 73cm

The uppermost section of the core is fully humified fresh *Sphagnum* peat, *Turfa bryophytica* makes up 50% of the sediment with abundant stems, roots, protonema, leaves and rhizods of *Sphagnum moss. Turfa herbacea* makes up the remaining 50% of the sediment, with *Myriophyllum alterniflorum* stems and roots present.

5.2 Pollen and Charcoal

The following nine pollen zones have been recognised within the Meall Bad à Chrotha record. The zones were determined using stratigraphy constrained cluster analysis (CONISS) in Tilia (figure 37).

MBAC-1: 500 – 475cm

Open Lacustrine Environment (*Myriophyllum alterniflorum*)

The start of this zone at 500cm is the earliest part of the record from Meall Bad á Chrotha. *Myriophyllum alterniflorum* (water milfoil), an aquatic pollen, is the dominant taxon at the base of the record. This type of pollen was not added into the Total Land Pollen (TLP) sum as the vast amounts produced and recorded would suppress the percentages of the other pollen types. 1994 *Myriophyllum alterniflorum* pollen grains were recorded at the base of the record indicating a water body perhaps 1 - 2 metres in depth present at the site. There were also large quantities of algae, most probably *Pediastrum*, and diatoms found further supporting the presence of a relatively clear water body at this point in time. There are also fairly high levels of the water-tolerant fern Polypodiaceae (fern), remaining around 25-35% TLP which was most likely present in the damper areas around the pond along with lower abundances of Cyperaceae (sedges). *Myriophyllum alterniflorum* shows a rapid, sharp decline and then disappears from the record at the end of the zone, not to appear again in the pollen sequence suggesting terrestrialisation and encroachment of vegetation and sediment into the water basin. Polypodiaceae shows a drop by over half from around 30% going into MBAC-2. The abundance of algae and diatoms also drops as the water body gets infilled.

Ericaceae (heather) is also a dominant contributor to the vegetation composition at Meall Bad á Chrotha, peaking to nearly 55% TLP after the decline in *Myriophyllum alterniflorum*. Ericaceous dwarf shrubs, most likely *Empetrum*, and Poaceae (grass) were probably present in the drier areas of the site. The lowering of the water table and the establishment of marshy conditions would allow the colonisation and expansion of less water-tolerant communities. *Artemisia* (daisy), after a peak at the base of the record, also declines, indicating a shift from more pioneer communities, which can colonise and grow on very poor, bare soils, to more nutrient-demanding communities.

The percentage of arboreal pollen is very low during this zone, suggesting a more open lacustrine habitat at Meall Bad á Chrotha. *Betula* (birch) is the most abundant tree pollen during MBAC-1, although was still in fairly low abundance. After a slight dip in the centre of the zone, it increases to around 35% TLP. This *Betula* pollen most likely represents *Betula nana*, or birch shrub, and was probably present in the damper areas around the water body. *Corylus/Myrica* (hazel/bog myrtle) appears at 490cm, slightly later than the appearance of *Betula*, rising to 15% TLP. It probably joined *Betula* in the damper areas as they can successfully grow together and both can tolerate damper conditions. The rising level of arboreal pollen towards the end of the zone coincides with the decline in wet indicators. It is likely that the drying out and terrestrialisation of the water body would provide more suitable space allowing for the expansion of such taxa, such as *Betula* and *Corylus/Myrica,* from their sheltered refugia in the damper areas within the catchment. The warming and drying indicated from the vegetation shift would provide a more adaptable ecological niche for the growth and spread of woodland.

After a minor peak in *Pinus sylvestris* (Scots pine) at the start of the zone to around 15% TLP, it begins to decline. A threshold of 20% TLP has been assigned as an indicator of local pine growth (Bennett, 1984). There could have been small pockets of pine present on the drier locations within the catchment that contributed to the pine pollen recorded at the site. *Salix* (willow) also peaks to 10%, at 500cm; highest percentage recorded for the taxa during the profile, although shows a steady decline during the rest of the zone. It was probably locally present, growing in the damper areas around the water body although not in large abundances, just before the appearance of *Corylus/Myrica* and the transition to drier conditions which contributed to its decline.

There is a very minor occurrence of charcoal at 490cm. The irregularity and small scale of this charcoal occurrence along with the lack of any other evidence of anthropogenic activity suggests that this occurrence is most likely due to natural wildfires. Indeed, it is possible that the transition to a drier landscape inferred from the vegetation could well have promoted occasional, single fires of natural origin somewhere within the area; the small level of charcoal found suggests that the fire was not immediately local to the site.

MBAC 2: 475 –395cm

Birch Woodland *(Betula,* Ericaceae*, Sphagnum)*

Entering into the zone at 480cm, *Betula* is the most dominant vegetation type peaking to 45% TLP at 460cm, indicating open birch woodland locally to Meall Bad á Chrotha. Also, at 460cm, Ericaceae (presumably *Calluna vulgaris*) reaches its lowest value during the zone, just 10% TLP. However, after a decline entering the zone, Ericaceae increases to become the most dominant taxon, peaking to 70% TLP, as *Betula* steadily declines, dropping to 25% at 400cm, indicating a thinning in woodland density. Both *Betula* and Ericaceae show a corresponding relationship, particularly in the second half of the zone. The decline of *Betula* would allow the expansion of light demanding Ericaceous dwarf shrubs. Although, there is the possibility that the local bog taxa, such as Ericaceae, is masking the more regional pollen, *Betula*, rather than it declining. The *Corylus/Myrica* curve remains between 5-15% TLP throughout the majority of the zone. *Corylus* is more likely to have contributed to the *Corylus/Myrica* curve during this zone as mixed birch/hazel woodlands are a common feature in pollen diagrams from across Scotland just before the rise in *Pinus sylvestris* which occurs in MBAC-4. There are no indications which suggest that *Myrica* would be present over *Corylus* during this zone. It would also be expected that if *Myrica* was present, the curve would be more inflated as both pollen types would be contributing to the curve. Both *Pinus sylvestris* and *Salix* are present throughout the zone, although neither reaching above 10% TLP. *Alnus* (alder), *Quercus* (oak), *Ulmus* (elm) and *Tilia* (lime) are infrequent aspects of the zone and none contribute more than 5% to the total land pollen. The increase in arboreal pollen indicates the establishment of a predominantly *Betula* woodland, mixed with low levels of *Corylus*. The very low percentages of other tree taxa recorded during MBAC-2 may be attributed to the transportation of pollen from other sites within the catchment where they were present in higher abundances. Although, very small sheltered pockets or individual trees may have been present immediately on or around Meall Bad á Chrotha in the more suitable areas, therefore producing these very low percentages seen during this zone.

Both Poaceae and Cyperaceae are dominant in the herb layer at beginning of the zone before declining and remaining below 10% after 460cm. Cyperaceae peaks for the first time in the record to 40% TLP at 470cm. It would have been present in the wetter, marshier areas left by the retreating waters of the former pond. However, this peak is short lived indicating a continued transition to drier conditions. As Poaceae and Cyperaceae decline to low values, *Pteridium* (bracken) appears at 450cm for the first time in the record; although it remains under 5% TLP. *Pteridium* can grow in a wide variety of conditions but most likely grew in acidic, boggy areas along with *Sphagnum* which acidifies its surroundings. Following the peak at the start of the record, *Artemisia* declines down to under 5% TLP and becomes an infrequent component during this zone, signifying an end to the pioneer period as soil was becoming increasingly rich in nutrients, allowing more nutrient-demanding taxa to colonize.

Sphagnum peaks to its highest value so far in the record, just above 200 at 460cm. It then shows a relatively sharp decline before rising again to around 100 at the end of the zone. *Sphagnum* is also not included in the Total Pollen Sum as its large abundances can suppress the percentages of the other taxa. This increase in *Sphagnum* could indicate that peat-formation was taking place at Meall Bad á Chrotha. The lowering of water body depth would have left a marshy area with small water bodies where mosses could thrive. The high level of algae, diatoms and water-tolerant vegetation found during MBAC-1 continue to decrease substantially going into MBAC-2 with Polypodiaceae failing to rise above 15% TLP for the majority of the zone. The transition to drier conditions witnessed during MBAC-1 continues throughout MBAC-2.

Similarly to MBAC-1, there is only one very minor occurrence of charcoal at 430cm. However, at 420cm, *Plantago lanceolata* (ribwort plantain) occurs for the first time in the record; although, only to 1-2% TLP. *Plantago lanceolata* is a common weed found on cultivated land, therefore usually indicates human activity. Although, the small percentage found, the lack of any further evidence of agricultural practices and land use and the early timing of its appearance suggests that this taxon most probably occurred in minor abundances where there had been some natural disturbance for a very brief period. The occurrence of charcoal could again be a result of natural fires somewhere within the area. The possibility of contamination being responsible for this occurrence of *Plantago lanceolata* cannot be ruled out, however, there were measures put in place to prevent any contamination.

MBAC-3: 395 - 345cm

Birch/Hazel Woodland (*Betula, Corylus/Myrica,* Ericaceae)

At the start of MBAC-3, *Betula* peaks to between 45-50%, accompanied by a slight drop in Ericaceae. *Betula* then shows a decline, although still remains the dominant arboreal pollen never falling below 25% TLP. The *Betula* woodland that began to develop during MBAC-1 becomes increasingly mixed with *Corylus/Myrica* as it shows a steady increase throughout the zone, rising from around 15% TLP at 390cm to around 35% TLP at 350cm. *Alnus* becomes a continuous component throughout this zone, although remaining low at around 5% TLP. Also, consistent but below 5% TLP is *Pinus sylvestris*. Similarly to MBAC-2, there are also infrequent low values of *Quercus, Ulmus*, *Fagus* (beech), *Carpinus* (hornbeam) and *Salix* recorded, although none contribute more than 1-4% to the total land pollen suggesting that they were present within the area but not immediately close to Meall Bad á Chrotha. The appearance of *Fagus* in MBAC 1-3 is surprising as it is very early on for this taxon to appear, although found in minor values. Birks (1989) suggested *Fagus* took advantage of abandoned agricultural land 3000 years ago which signified its entrance into Britain. Therefore, the appearance of this taxon during these three zones must be taken with caution as it is possible that human error or contamination has played a role.

Ericaceous dwarf shrubs, most likely *Calluna vulgaris*, begin to increase gradually peaking to 50% TLP at 360cm becoming a dominant taxon in the woodland under-story shrub layer. It then shows a rapid and sudden decline at the end of the zone at 350cm to just 5% TLP; the lowest percentage of Ericaceae recorded throughout the Meall Bad á Chrotha profile. Also, at 350cm there is a very sharp and rapid increase in the percentage of *Betula* rising from around 25% TLP to just over 60% TLP, the highest percentage of *Betula* recorded throughout the record. *Corylus/Myrica* also peaks, along with *Betula*, at 350cm indicating dense birch/hazel woodland up until 350cm.

The total percentage of herbs during MBAC-3 is the lowest recorded throughout the profile. Poaceae and Cyperaceae remain in low abundance with Cyperaceae disappearing from the record at 360cm; although, Poaceae does show a slight increase at the end of the zone. After a sharp decline in *Sphagnum* at the start of the zone, it falls to very low values and disappears from the record towards the end of MBAC-3. Polypodiaceae also makes a transition to very low values at 380cm. Their decline and the decline of Cyperaceae, another wet indicator, suggest that there was a further shift to drier conditions and a reduction in peat-formation.

The occurrence of charcoal increases during this zone, however the level of charcoal found does not indicate that any burning occurred directly next to the site as they are just 'background' levels. There is the possibility that there may well have been some anthropogenic activity which interfered with the fire regime further away from Meall Bad á Chrotha. However, there are no other indications of human activity such as land clearance or agricultural indicators; therefore, it is still likely that these charcoal concentrations were a result of natural fires.

MBAC-4: 345 – 295cm

Open Landscape (Ericaceae)

The opening of the zone shows the end of dense birch/hazel woodland. The *Corylus/Myrica* and *Betula* peaks at the end of MBAC-3 were short-lived soon falling back down to lower percentages at 340cm along with a sharp rise in Ericaceae; although, *Betula* remains a dominant taxa at 30% TLP. The decreasing percentage of arboreal pollen and increasingly high percentage of shrubs indicate the establishment of a progressively open landscape, although with low levels of birch still present, perhaps in the wetter areas within the Meall Bad á Chrotha site. Ericaceae shows a gradual rise throughout the zone, peaking to over 80% TLP at 300cm. *Betula* shows a correlation with Ericaceae in this zone also, similarly to the rest of the profile. As Ericaceae increases, *Betula* declines. *Betula* declines from around 30% at 340cm down to just above 10% but shows a steady decline throughout MBAC-4. Similarly to MBAC 2, there is also the possibility that the rise in local bog taxa, such as Ericaceae, is masking the more regional *Betula* pollen rather than it declining. This hypothesis should be considered. The *Corylus/Myrica* curve, after a sharp decline during MBAC-3 remains around 10% before falling even further at the end of MBAC-4. This reduction in birch/hazel woodland would have allowed the expansion of heathland, in particular *Calluna vulgaris*, onto the once occupied land. As *Betula* decreases, *Pinus sylvestris* begins to gradually rise, reaching to around 10% going into MBAC-5. The expansion of Scots Pine is usually encouraged by a decrease in the level of bog surface wetness. Indeed, wet indicators are at a minimum with Cyperaceae, Polypodiaceae and *Sphagnum* absent or at very low values throughout the majority of the zone. *Alnus, Quercus, Ulmus, Tilia* and *Salix* are present but also at very low values.

The overall percentage of herbs is very low during this zone, with *Artemisia* less than 5% TLP at the very start before disappearing. Poaceae contributes the highest percentage to the herb total at the start of the zone peaking to 20% TLP at 330cm before declining to less than 5% TLP.

There are small but noticeable increases in the frequency and concentration of charcoal which suggests intensification in the fire regime. This gives increased evidence for the occurrence of fire within the area surrounding Meall Bad á Chrotha. It is still debatable whether this change in the fire regime was due to natural or anthropogenic activity. The decrease in wet indicators and increase in *Pinus sylvestris* does indicate a shift to increasingly dry bog conditions which would encourage the occurrence of spontaneous wildfires; lightning strikes can also be associated with natural fires, particularly when dry conditions prevail. However, it is possible that there were some early settlers somewhere within the catchment which used fire as a means of land clearance and for other domestic purposes. Although, there is no further evidence to support the suggestion that these charcoal occurrences were due to human activity, it should be strongly considered.

MBAC-5: 295 – 205cm

Pine Woodland (*Pinus sylvestris*)

After the very slight and gradual increase in *Pinus sylvestris* during MBAC-4, the abundance of pine pollen found continues to rise throughout the start of this zone, signifying the ‘Pine Rise’. For the first time in the record it begins to show significant increases and peaks. At 290cm, pine rises to over 20% TLP, the threshold put forward by Bennett (1984) to indicate local pine growth. By 280cm it rises to 40% TLP. After a short lived but substantial decline, *Pinus sylvestris* soon recovers showing a rapid increase to just over 35% TLP. Even though there are minor declines, there is still an evident rising trend in the pine pollen percentage, indicating the development, establishment and increasing density of pine woodland. At 210cm, pine peaks to its highest percentage recorded throughout the profile from Meall Bad á Chrotha. This peak rises to between 75 – 80% TLP; the highest percentage of arboreal pollen recorded from the site.

As *Pinus sylvestris* peaks to its highest percentage within the record, *Betula* drops to its lowest within the zone, to less than 5% TLP, at 210cm. A rise in the abundance of *Alnus* is clear at the start of this zone although does not rise above 10% TLP. The *Corylus/Myrica* curve is also present throughout the zone. Similarly to *Betula*, it shows fluctuations although only minor, remaining between 5-10%. Comparable to *Betula* and *Alnus, Corylus/Myrica* also declines to its lowest value as pine peaks to its highest. It is possible that the large amount of pollen produced by pine trees when they are growing in optimum conditions is responsible for depressing the pollen percentages of the other taxa. However, it is also possible that these taxa were out competed and pushed into areas unsuitable for growth by the very dense pine woodland. *Quercus* is also present throughout MBAC-5, although remains below 5% TLP. *Ulmus* and *Salix* are present but are not a continuous aspect throughout out the zone, and never rise above 5% TLP. *Carpinus* and *Fagus* are the rarest arboreal pollen during this zone, with *Carpinus* only occurring once and *Fagus* occurring three times; both only to 1 – 2% TLP.

Ericaceae shows a close relationship with pine during this zone, both showing corresponding fluctuations. There is an overall declining trend in Ericaceous dwarf shrubs as there is an expansion in pine woodland. At 210cm, Ericaceae contributes only around 10% to the total land pollen percentage; as pine peaks to its highest percentage. This relationship between Ericaceae and pine could reflect a reduction in peat forming conditions due to a drier mire surface. However, light demanding Ericaceae could have declined due to the shade created by the dense pine canopy.

The overall abundance of herbs during this zone is low. These very low percentages of Cyperaceae, *Sphagnum* and Polypodiaceae, which thrive in damper conditions, could indicate that conditions on the mire surface were very dry during the pine rise. At 250cm, there is a small rise in Cyperaceae which coincides with a slight decline in pine. This could represent a minor short-lived local change in the hydrological status of the bog surface.

For the first time in the record from Meall Bad á Chrotha, charcoal occurs through the majority of the zone, perhaps encouraging the growth of pine through the removal of its competitors. During the first half of the zone the level of charcoal found is slightly higher, although still at low values. The rising level and higher frequency of charcoal found during this zone provides increasing evidence for a change in the fire regime at this point in time, however the low values suggest that it is a more regional background change rather than local. Assuming human activity is the cause is very unreliable as the charcoal values are still very low and there is a lack of any other evidence which would suggest human disturbance. The dry conditions inferred from the vegetation composition could well have encouraged the occurrence of wildfires; the more continuous appearance of charcoal could result from a more regional distribution of natural disturbance.

MBAC-6: 205-185cm

Open Pine Woodland (*Pinus sylvestris*, Ericaceae)

The pine peak is short lived, as the 'Pine Decline' is also a clear aspect of this MBAC-6. There is a clear sharp decline in the percentage of pine pollen and the overall percentage of arboreal pollen with a substantial thinning of woodland and the establishment of openings. At 190cm, pine contributes between 25-30% to the total land pollen. *Betula* pollen continues to show steady and gradual fluctuations throughout the zone remaining between 5 – 20% TLP. The collapse of this woodland allows for the expansion of Ericaceae once again into the increasingly open areas. Both Poaceae and Cyperaceae increase immediately after the pine peak suggesting an immediate response to the changing conditions. At 200cm, Cyperaceae begins to increase, and peaks to just over 10% TLP at 190cm. The rise in

Cyperaceae and the collapse of pine can be inferred as a transition back to wetter conditions.

Charcoal is present throughout the zone, although still at very low levels. The increase in charcoal occurs at a time of substantial deforestation; therefore, fire as a means of woodland clearance should be taken into consideration; however, the low levels of charcoal indicate that the source was more regional rather than local and more likely attributed to natural disturbance. Furthermore, no other taxa show a substantial decline similarly to pine, further suggesting that fire was unlikely to be the cause for thinning of the pine woodland. The increase in wet indicators suggests that a change in the hydrological status of the bog was a greater influencing factor in the collapse of pine.

MBAC-7: 185 - 165cm

Open Landscape (Cyperaceae, Ericaceae)

Pine declines down to the lowest value recorded since the start of its rise at 330cm, to less than 5% TLP at 180cm with a transition to a predominantly open landscape and the end of pine dominance. There is one final rise in the percentage of pine pollen at 170cm, a rise to between 10 – 15%. This rise is coincident with a drop in the percentage of Cyperaceae to less than 5% suggesting that there could have been a local decrease in water table depth which influenced this short-lived vegetation change. Throughout the zone, Ericaceous dwarf shrubs are dominant, peaking to 60%, declining slightly to 55% at 170cm. This reflects the decline in pine woodland as Ericaceous shrubs, particularly *Calluna vulgaris*, would have expanded into the nearly open spaces, and the reduction in the forest canopy would have provided a suitable environment for this light demanding taxa to thrive in. Poaceae also responded rapidly to the changing environment, expanding into the newly open spaces, contributing about 20% to the total land pollen percentage throughout the zone. The level of charcoal remains the same as in the previous zone.

MBAC-8: 165 – 115cm

Open Landscape (Cyperaceae)

Pinus sylvestris declines down to just over 5% TLP at 160cm and remains at less than 5% for the rest of the zone. It is possible the spread of blanket bog and the increasingly wet conditions inferred from the vegetation prevented the regeneration of pine. *Betula* is the most significant type of arboreal pollen during MBAC-8 remaining between 10-15% TLP before showing a gradual decline. The percentage of *Alnus* increases at the start of the zone with a peak up to 15% at 150cm, the highest recorded percentage of *Alnus* within the profile. However, *Alnus* quickly declines back to down to around 5% TLP, before showing a further decline to significantly low values towards the end of the zone. *Quercus* is present throughout the majority of the zone, although only at 3-4% TLP. *Carpinus* and *Fagus* occur more regularly, compared to any other zone, however to no more than 2-3% TLP.

The collapse of woodland and establishment of an open habitat is reflected in the abundance of herbs and shrubs. Cyperaceae rises to 55% TLP at 140cm, after a decline at the end of MBAC-6. However, this peak is short lived as it declines down to less than 10% TLP at 120cm. Ericaceae shows a gradual rise throughout the first part of the zone but declines down to 10% at 120cm alongside Cyperaceae. These declines are coincident with a sharp spike in the *Corylus/Myrica* curve. It would be unlikely that this peak is associated to the growth of *Corylus* as there are no other increases in arboreal pollen. From the start of the zone up until 130cm, *Corylus/Myrica* fluctuates around 10% TLP, before suddenly peaking to more than 60% TLP at 120cm. *Corylus/Myrica* may have simply out competed Ericaceae and *Cyperaceae* as the conditions were more favourable for its growth, such as a very local increase in bog wetness, or it is possible that the high level of *Myrica* pollen produced depressed the other taxa values. This spike in *Myrica* is likely to have been very local, perhaps an individual plant immediately on the coring location which could have be responsible for the high abundance of *Myrica* pollen grains and its short-lived nature. Its decline is just as sudden and significant falling back down to around 10% TLP immediately following the peak.

The domination of Cyperaceae suggests that there was deterioration in bog conditions with an increase in the level of wetness. This is supported by the very gradually increasing levels of *Sphagnum* with indicate the development of blanket bog. There is a sharp and sudden spike in *Sphagnum* at the 150cm. This is most likely due to a variation in local conditions. This rise in *Sphagnum* and relatively high abundance of Ericaceae *(Calluna vulgaris)* could indicate a very local rapid enhancement of peat forming conditions.

Plantago lanceolata, apart from a short-lived and minor appearance in MBAC-2, appears and remains a clear component of the Meall Bad á Chrotha record for the first time. At 160cm, *Plantago lanceolata* is recorded, however only in low values between 1-3% TLP. Although, the values remain low, this occurrence of *Plantago* could indicate the establishment of a more permanent settlement community somewhere within the area. However, further evidence for human activity is rather limited as there is no substantial changes in any other taxa that may represent human disturbance. *Artemisia* and *Pteridium* remain below 5% TLP, while Poaceae, although more prolific than earlier in the record, still does not represent more than 15% TLP, and shows a slight decline towards the end of the zone. Assuming human activity based on evidence from only one taxon, which still represents a low proportion of TLP, is rather unreliable; therefore, the possibility that this occurrence of *Plantago lanceolata* is natural must be taken into consideration. Although, its continued occurrence and evidence collected by other investigations does support the theory that there may well have been a small scale permanent settlement somewhere within wind dispersal distance that began around this time.

MBAC 9: 115 – 0cm

Open Landscape (Cyperaceae, Ericaceae)

This is the final zone in the record from Meall Bad á Chrotha and it signifies a shift towards the open landscape and vegetation that we see at the site today. Arboreal pollen drops to its lowest levels during this zone following the deforestation period, while herbs peak to their highest. *Betula* declines down to less than 5% TLP at the very top of the profile. Small pockets of *Betula* and *Corylus* could well have still persisted in wetter areas. There is upland *Betula* woodlands present around streams in the catchment today. *Pinus sylvestris* remains below 5% TLP throughout the zone; with the overall percentage of arboreal pollen markedly declining at 10cm with many trees disappearing from the record and the rest dropping down to less than 3-4% TLP. At 110cm, *Corylus/Myrica* is back down to around 10% TLP after peaking, allowing Ericaceae and Cyperaceae to increase back up to higher abundances. Cyperaceae then continues to fluctuate between 30 – 50% TLP until 30cm where it peaks up to 70% TLP at 30cm. Ericaceae peaks to around 50% TLP at 70cm after a gradual increase from the start of the zone. Ericaceous dwarf shrubs such as *Calluna vulgaris* are very tolerant of water logged conditions, therefore invade the mire surface. Cyperaceae and Ericaceae remain the dominant taxa for the remainder of the record. These taxa along with the gradually increasingly levels of *Sphagnum* suggest that peat-formation was taking place in the catchment. Poaceae is also present in the herb layer, fluctuating between 10 – 15% TLP after a decline at the start of the zone.

There is an indication of increased anthropogenic activity at Meall Bad à Chrotha during this zone. *Plantago lanceolata* continues to increase ever so slightly up to around 5% TLP at 10cm. The more regular occurrences and slightly elevated percentages of *Plantago lanceolata* do suggest that some agricultural activity was present in the area at this point in time, although the low percentages indicate that it was not immediately on or near the Meall Bad á Chrotha site. The occurrence of charcoal also becomes more frequent during this zone, being recorded at every depth. The level of charcoal found remains relatively low although still higher than some of the previous occurrences recorded lower in the profile. This intensification in anthropogenic activity and the coincident vegetation changes may well be linked. The final drop in arboreal pollen at the top of the profile could well be due to the clearance of the remaining trees for agricultural or domestic purposes, or by the persistent grazing of livestock within the area.

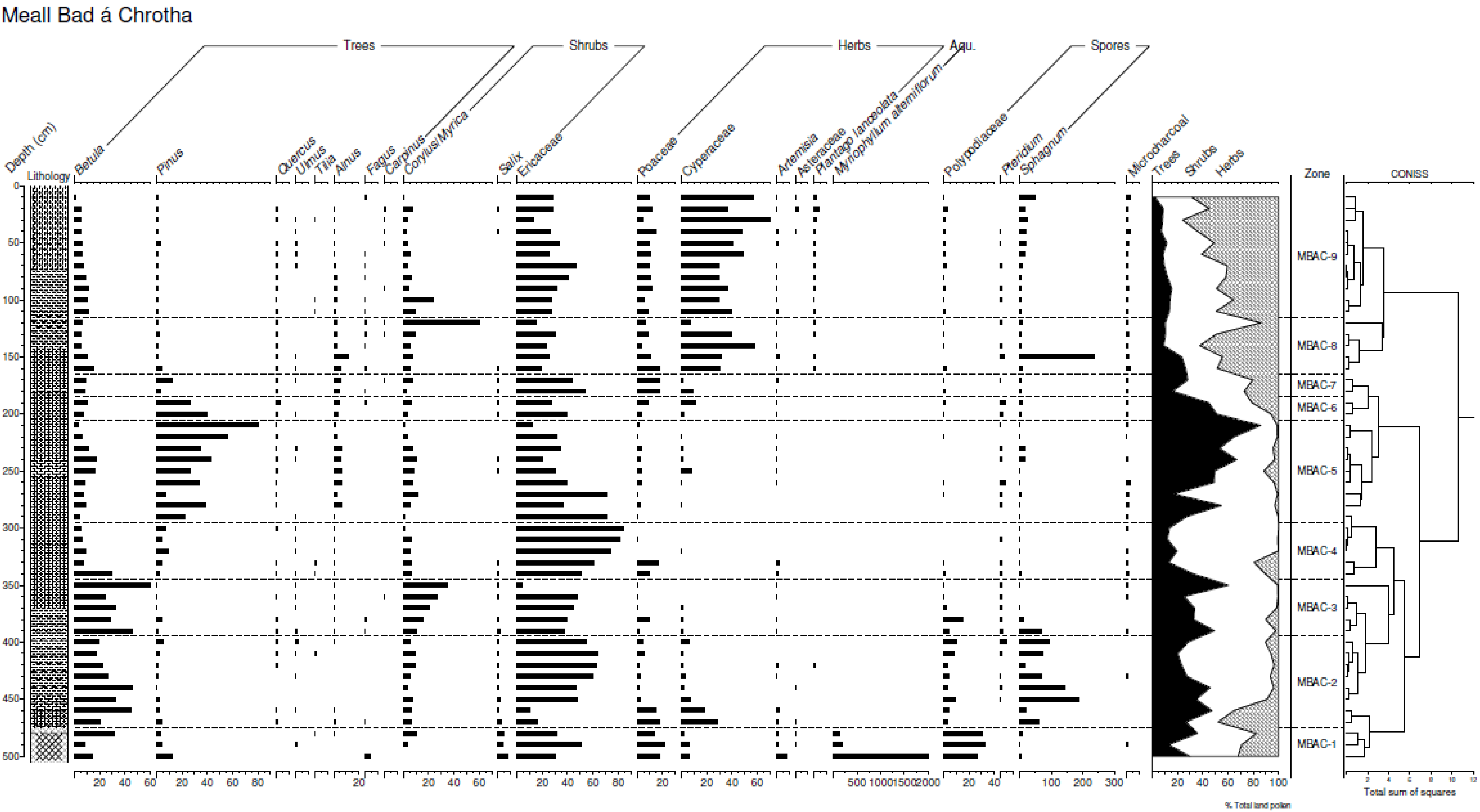


Figure 37. Pollen, spore and charcoal diagram with CONISS and lithology from Meall Bad à Chrotha, Wester Ross, Northwest Scotland. *Sphagnum* and *Myriophyllum alterniflorum* were not included in the total land pollen sum.

5.3 MTWA Reconstruction

The C2 package (Juggins, 2007) was used to produce the mean temperature of the warmest month from the fossil assemblage retrieved from the Meall Bad á Chrotha core. MTWA was the chosen variable as it allows the identification of any significant drops in summer temperature, which could push pine and a number of other taxa, passed their survival threshold; therefore, allowing taxa which require a reduced temperature to invade. Attention must be drawn to the complicated relationship between *Pinus sylvestris* and the MTWA. Caution must be taken when interpreting this relationship as the MTWA fluctuations are in fact primarily driven by the pine pollen curve; therefore, arguing that the growth and collapse of pine match with the temperature variations is a circular argument. Taking this into consideration, the record produced from Meall Bad á Chrotha will be largely compared with independent climate reconstructions. There are more independent investigations which reconstruct MTWA rather than annual average, therefore reconstructing this variable gives more scope for record comparison. The zones on the MTWA graph are based on the pollen evidence to allow easy comparison/interpretation and avoid confusion.

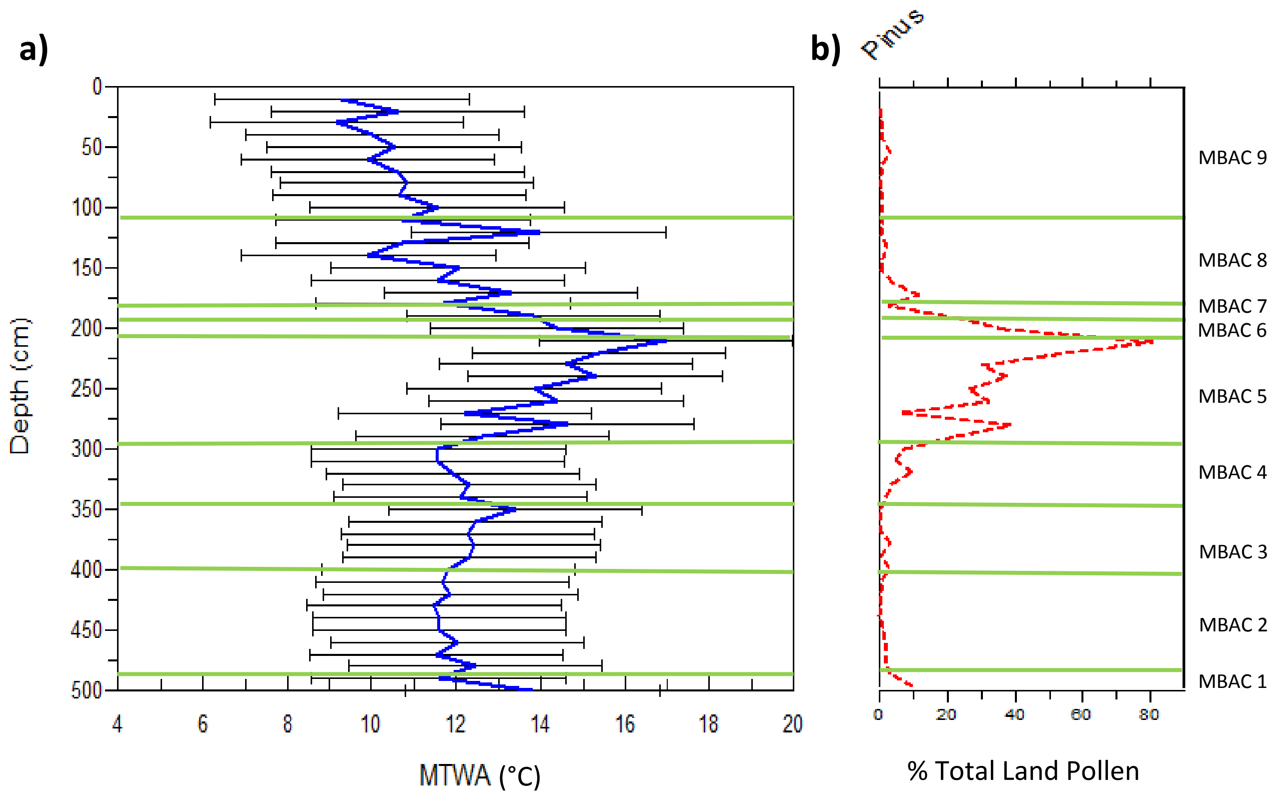


Figure 38a and 38b. a) The Mean Temperature of the warmest month inferred from the application of the fossil assemblage collected from Meall Bad á Chrotha to the model seen in figure 36 produced using a modern pollen-climate data set **b)** The total land pollen percentage of *Pinus sylvestris* to demonstrate the relationship between the taxon and MTWA.

MBAC 1: 400-475cm, MBAC 2:475-395cm and MBAC 3: 395-345cm

Temperature Decline and Early Holocene Warming Trend

After a decline at the very base of the record, the temperature shows an upward trend, rising from 11.5°C to 13.5°C – a trend which is associated with the warming of the early Holocene; the rising temperatures are associated with the increasing abundances of *Betula* and *Corylus* which expanded under the increasing suitable growing conditions.

MBAC 4: 345-295cm

Temperature Depression

A significant depression curve of 2ºC takes the MTWA back down to values witnessed at the base of the record (11.5ºC). This depression shows a close relationship with *Corylus* and *Betula*, with substantial collapses in both taxa as the temperatures drop. A rise in Ericaceae is also evident at this time.

MBAC 5: 295-205cm

Holocene Thermal Maximum

There is a sharp rise in temperature peaking up to 16.5ºC correlating with the pine maxima. It appears the relationship between *Pinus sylvestris* and temperature is very strong and sensitive, with temperature directly influencing pine expansion and contraction, however the MTWA is essentially driven by the pine pollen, as discussed above. MTWA reaches its highest value during this zone indicating the establishment of the Holocene Thermal Maximum.

MBAC 6: 205-185cm, MBAC 7: 185-165cm, MBAC 8: 165-115cm and MBAC 9: 115-0cm.

Holocene Thermal Decline

After the Holocene Thermal Maximum came the Holocene Thermal Decline. This is evident is both the MTWA reconstruction and the fossil pollen assemblage. There is a sharp decline in MTWA from as high as 16.5ºC to as low as 10ºC. This trend is in line with the 'Pine Decline' seen in the pollen record. This overall cooling trend continues towards modern day (except for a couple of very short-lived increases, which can particularly be seen at the end of MBAC 8) – with MTWA falling below those seen during the early Holocene. The reduced MTWA can be inferred by the continued fall in arboreal pollen with a rise in wet indicator taxa such as Cyperaceae from the mid-late Holocene.

Zone	Depth of sample (cm)	Dominant Taxa	Description and Interpretation	MTWA Reconstruction
MBAC-1	500-475	*Myriophyllum Alterniflorum*	Open lacustrine environment - water body of about 1-2 metres in depth	Cool - Fluctuating between 11.5°C to 12.5°C
MBAC-2	475-395	*Betula,* Ericaceae, *Sphagnum*	Open birch woodland – terrestrialisation of water body and improved, drier conditions allowed for birch expansion	Very gradual warming after a minor fall – MTWA between 11.5°C to 12.5°C
MBAC-3	395-345	*Betula, Corylus/Myrcia,* Ericaceae	Birch/hazel woodland – continued climatic improvement allowed for the establishment of hazel and mixed-deciduous woodland expanded	More pronounced warming trend peaking to 13.5°C
MBAC-4	345-295	Ericaceae	Open landscape – decline of birch/hazel woodland, dry bog conditions allowed pine to colonize	Cool depression dropping to 11.5°C
MBAC-5	295-205	*Pinus sylvestris*	Dense pine woodland – dry and warm conditions lead to dense pine woodland	Sharp rising trend of around 5°C peaking to 16.5°C
MBAC-6	205-185	*Pinus sylvestris,* Ericaceae	Open pine woodland - A transition to wetter bog surface levels made conditions unfavourable for growth leading to the 'Pine Decline'	Sharp cooling trend
MBAC-7	185-165	Cyperaceae, Ericaceae	Open landscape – continued wetter, peat forming conditions prevented the regeneration of woodland and *Pinus sylvestris*	Sudden, short-lived rise of about 2°C
MBAC-8	165-115	Cyperaceae	Open landscape – Cyperaceae thrived in the wetter environment, peat-formation continued and anthropogenic activity intensified within the catchment	Continued cooling trend before a sharp rise up to 14°C
MBAC-9	115-0	Cyperaceae, Ericaceae	Open landscape – Cyperaceae continued to be dominant and woodland declined to its lowest levels as peat expanded. Small pockets of *Betula* and *Corylus* persisted in more isolated localities. Settlements became more permanent and agricultural activity intensified within the catchment.	Overall declining trend towards modern day fluctuating around 9.5°C and 10.5°C

Table 16. Interpretative log of the environmental changes inferred from the vegetational and climatological (MTWA) reconstructions for Meall Bad á Chrotha.

Chapter 6: Discussion and Interpretation

Lowe (1993) summarized the vegetational succession of Scotland for the Holocene in three general phases: The 'Pioneer' began at the ending of the Lateglacial period around 11,700 cal BP; following this was the 'afforestation' phase from 10,400 – 5000 cal BP and finally the 'deforestation' phase commencing at 5000 cal BP to present day. These phases can be broadly correlated to the majority of palynological reconstructions from Scotland and indeed Meall Bad á Chrotha.

6.1 Lateglacial-Early Holocene: Pioneer communities and the Afforestation Period

The opening of the record (500cm) from Meall Bad á Chrotha, Wester Ross, Northwest Scotland, indicates that an open lacustrine environment was present. Fine detritus organic mud is the dominant sediment type further suggesting a lake filled the area. The high abundances of the aquatic vegetation, such as *Myriophyllum alterniflorum,* and the high levels of algae and diatoms, are indicative of clear water with water depths of >60 – 100cm (Spencer, 1964). This is commonly marked as the transition into the early Holocene from the final significant cold period of 'Arctic severity' to affect the British Isles which is known as the Loch Lomond Stadial and occurred between 13,000-11,700 cal BP (Shennan et al, 1993; Selby, 2004). The substantial abundance of algae probably represents *Pediastrum*; an easily identifiable and common component in many Lateglacial reconstructions. It also suggests that the water body was relatively nutrient-rich, perhaps receiving its nutrients from the erosion of soil from around the lake and further upslope (Tipping and McCulloch, 2003). The peak in *Myriophyllum alterniflorum* at Cross Lochs, in the Flow Country, eastern Sutherland is dated to 12,716 cal BP. Similar vegetation successions characterised by aquatic communities are recorded at Loch Ashik on the Isle of Skye at 13,185 cal BP (Birks and Williams, 1983) and at Rumach Iochdar, near Arisaig in the Northwest. Shennan et al (1993) associate this peak in *Myriophyllum alterniflorum* to the earliest stage in the transition from the Loch

Lomond Stadial into the Holocene. Coastal locations within Wester Ross, such as Meall Bad á Chrotha, would have suffered with more extreme and exposed conditions during the Lateglacial and early Holocene in comparison to the sheltered mountainous inlands. The vegetational successions of these areas show various resemblances to certain locations within the Isle of Skye due to their comparable exposed nature. Cyperaceae, indicative of wet conditions would have been present in the damp areas around the water body at Meall Bad á Chrotha. *Empetrum nigrum* would have most probably been a dominant Ericaceous dwarf shrub in extensive patches of drier heathland along with abundant grasses and Polypodiaceae. *Artemisia* is present although a relatively rare taxon at the site but it is characteristic of many early Holocene pollen sites from Northern Scotland; as a poor competitor it was one of the first pioneer taxa to colonise the open bare soils in Lateglacial environments, however it was quickly out-competed as the soil became more fertile and more nutrient-demanding vegetation could successfully compete (Birks, 1973; Birks and Williams, 1983; Walker et al, 1988; Walker and Lowe, 1990). *Betula* shrub was present, although in relatively low frequencies, perhaps in the more sheltered locations within the catchment, which is characteristic of Northwest Scotland in comparison to other sites in the British Isles where it was an established feature of the landscape earlier (Pennington, 1975). The climatic deterioration that signified the Loch Lomond Stadial and the subsequent rapid rise in both summer and winter temperatures which characterised the first 500 years of the Holocene are evident in the biotic response at Meall Bad á Chrotha.

The Loch Lomond Stadial was the last substantial cold period within the British Isles. A surge in the level of meltwater into the North Atlantic from the Fennoscandian and Laurentide ice sheets (Ruddiman and McIntyre (1981), resulted in a shift of the Oceanic polar front in a southward direction. At 11,700 cal BP, the polar front moved to its maximum southerly location near the Southwest of Ireland (Ruddiman et al, 1977). Johnson and Maclure (1976) suggested that polar ice packs may well have reached as far as the north of Scotland. The abrupt transition to warmer conditions which is associated with the opening of the Holocene epoch, at 11,700 cal BP, is evident from data collected from the cores in Greenland (Dansgaard et al, 1989; Alley et al, 1993). The warming of the early Holocene is due to the movement of the oceanic polar front in the North Atlantic back northwards (Ruddiman and McIntyre, 1981). Bard et al (1987) suggest the movement of the polar front to be between 35°N and 55°N, coincided near perfectly with the end of the Loch Lomond Stadial. The movement created a substantial surge of warm water within the Atlantic, which ultimately resulted in a 9°C rise in sea surface temperatures in the Nordic Sea, within 50 years of the polar front migration (Koc et al, 1992; Koc and Jansen, 1994). The suggestion that the Loch Lomond Stadial to Holocene transition took less than 20 years was put forward by Alley et al (1993), and data collected from fossil coleopteran assemblages by Atkinson et al (1987) indicate that mean annual temperatures could have risen by 1.7°C every decade and by 7°C within 50 years of transition (Dansgaard et al, 1989).

A sharp increase in the percentage of Cyperaceae is accompanied by a decline in the level of Ericaceae at 470cm. The interpretation of changes in the abundance of Cyperaceae pollen can be problematic due to the challenge of identifying Cyperaceae pollen grains to the taxon level. Not all Cyperaceae taxa are adapted to the same moisture conditions making the interpretation of local palaeohydrological variations particularly challenging. However, a transition to wetter conditions from drier bog conditions can be inferred when a substantial rise in the percentage of Cyperaceae pollen relative to the level of Ericaceae pollen occurs (Anderson, 1998).

The abrupt reduction and disappearance of *Myriophyllum alterniflorum* from the Meall Bad á Chrotha landscape after a sharp decline and an increase in dry indicators signifies the transition to warmer and drier climes at the end of the Lateglacial period and the start of the Holocene. The decline of other taxa which prefer damper conditions such as Cyperaceae and Polypodiaceae also support the evidence to suggest climatic improvement and a transition to drier surface conditions. The lake present would most likely have undergone terrestrialisation as vegetation and sediment filled the area. The MTWA reconstructions show that the temperature at Meall Bad á Chrotha during this period was around 12.5^{o}C after a rise of just under 1^{o}C after as *Myriophyllum alterniflorum* decline. The correlation of this event between numerous sites from varying locations throughout the western seaboard and further north indicates that the disappearance of *Myriophyllum alterniflorum* was a regional occurrence resulting from a large scale hemispheric climatic improvement (Lowe, 1993). Gribun on the Isle of Mull (Walker and Lowe, 1987) and two sites on the Isle of Skye (Lowe and Walker, 1991; Benn et al, 1992) are characteristic of the early Holocene vegetation succession of the western British Isles. At these sites on the Isle of Mull and Skye this decline is followed by an expansion in *Filipendula* at 11,710 cal BP (Selby, 2004). Aalbersberg and Litt (1998) argue that the arrival and expansion of *Filipendula* is indicative of climatic improvement, with an increase in warmth pushing the mean temperature for July above 10°C. A rise in the temperature of the water is also indicated by the occurrence of *Nymphaea* on the Isle of Skye (Selby, 2004).

This transition from the Loch Lomond Stadial into the early Holocene signifies a shift from an open lacustrine environment with abundant Ericaceae (most probably *Empetrum* Heather), sedge and grass communities around the lake to the expansion of *Betula* woodland and the initiation of peat development at Meall Bad á Chrotha. *Empetrum* and *Calluna* are grouped together in a combined Ericaceae curve; therefore other records from the area were used to identify which taxon was more likely to contribute to the vegetation at any particular time at Meall Bad á Chrotha. As temperatures increased much of Scotland became increasingly wooded signifying the start of the afforestation period that characterises vegetation succession during the early Holocene. At Meall Bad á Chrotha, *Betula* was the dominant arboreal pollen with a slightly later expansion of *Corylus/Myrica*. The development of *Betula* and *Corylus* woodlands during the early to mid-Holocene is a characteristic feature of many northern Scottish pollen diagrams, as the increasing warmth provides adequate growing conditions without much competition from other taxa (Charman, 1994; Anderson, 1998). One limitation relating to the quantitative reconstruction from the Meall Bad á Chrotha core is that the pine pollen curve is essentially drives the MTWA plot, therefore as discussed previously, wherever possible, independent climate reconstructions have been used as a comparison. The reconstruction from Meall Bad á Chrotha shows that the MTWA fluctuated between 11.5^{o}C and 12.5^{o}C at this time. At Abernathy forest, further south in the Cairngorm area, the chironomid inferred mean July air temperature for the early Holocene fluctuated between 12^{o}C and 14^{o}C (Matthews et al, 2011), slightly higher than that inferred for the fossil assemblage at Meall Bad á Chrotha. Chironomids are highly sensitive to the smallest of temperature fluctuations therefore, are one of the most powerful proxies when it comes to reconstructing past climate fluctuations (Brooks and Birks, 2001).

The expansion of *Betula* occurred between 10,400 and 9800 cal BP on the mainland (Lowe and Walker, 1977; Vasari, 1977; Birks, 1977, 1980). The woodland remained relatively open with Ericaceous dwarf shrubs, presumably *Calluna vulgaris,* abundant in the under-storey and openings. The pioneer and competition intolerant taxa, such as *Artemisia,* declined during this early phase of the Holocene as more nutrient-demanding vegetation invaded the ground cover and the level of shade increased with the expansion of woodland. Although, Polypodiaceae decreased as *Betula* expanded, it was still present, most likely in woodland openings created by wind throw or tree senescence, in which light-demanding herbs and ferns grew locally for a brief period. A noticeable decline in Ericaceous dwarf shrubs, *Empetrum nigrum*, to only isolated patches is a common indicator of the Holocene transition at 11,700 cal BP and can be seen throughout the western Isles (Fossit, 1994). However, this decline in Ericaceous dwarf shrubs is only short lived as there is an expansion of *Calluna* which was most probably due to the increase in acidity through the process of leaching. Furthermore, the rise in *Sphagnum* spores is indicative of an increasingly acidic soil, despite the fact *Betula* is usually able to preserve and promote the alkalinity of the soil (Miles, 1985). *Sphagnum imbricatum* is a fairly oceanic taxon with high abundance in regions with an Atlantic climate (Godwin and Conway, 1939; Green, 1968). The expansion of *Sphagnum* could well have contributed to the reduction of other taxa such as Polypodiaceae, Cyperaceae and Poaceae. The ability of *Sphagnum* to encourage and promote acidity as well as nutrient deprived, anoxic bog communities means that many other plants are unable to grow in these conditions (Ohlson et al, 2001). The Ericaceae and *Sphagnum* rich community signify the start of blanket bog growth within a catchment (Boomer et al, 2012). The receding levels of the former pond would have left a waterlogged, boggy area, ideal for the initiation of blanket peat formation. Ericaceous dwarf shrubs have a higher tolerance to increased waterlogging (Bannister, 1964); therefore invade the mire surface as other taxa decline due to their inability to grow in the changing conditions.

The arrival of *Corylus/Myrica* is slightly later than that of *Betula* at Meall Bad á Chrotha. On the west coast, the arrival and subsequent rise in *Corylus* and/or *Myrica* is dated to just before 9500 cal BP, with marginally later dates further east and inland (Birks, 1989). The *Corylus/Myrica* curve during the early Holocene will most likely be *Corylus* as the majority of vegetation reconstructions show an increase in *Corylus* at this time as the climate warmed and conditions became drier. A rising trend in the MTWA is evident at Meall Bad á Chrotha at the time of the *Corylus* rise. There is a $2^{o}C$ rise in temperature from the beginning of the *Corylus* expansion to its peak ($13.5^{o}C$). At Abernethy forest, chironomids also show a similar temperature rise around this time up to between $13^{o}C$ and $14^{o}C$ (Matthews et al, 2011). At Reidchalmai, near Golspie in Sutherland, *Corylus/Myrica* (presumably *Corylus*) expanded to form a mixed woodland with *Betula* slightly later, comparable to Meall Bad á Chrotha. The date given for the gradual expansion of *Corylus* is between c. 10,500 and 9700 cal BP at Reidchalmai (Tipping and McCulloch, 2003). The gradual incorporation, a period of more than several hundred years, of *Corylus* into the deciduous woodland is unusual (Birks, 1989; Edwards, 1990; Tipping, 1994; Tallantire, 2002); however, the successful competition between *Corylus* and *Betula* at Reidchalmai was due to the micro-climate and pedological influences (Tipping and McCulloch, 2003). The gradual growth of *Corylus* means it is hard to estimate the date for the local expansion at Reidchalmai, but it is most likely nearer 9700 cal BP, in accordance with other expansions from the region and throughout the north of Scotland (Tipping and McCulloch, 2003). In the Northwest of Scotland, *Betula* was predominantly the most extensive tree, whereas in other parts of Scotland there tended to be a more equal combination of both *Betula* and *Corylus avellana* (Fossitt, 1996). Indeed, at Meall Bad à Chrotha, *Betula* remained the dominant tree over *Corylus* throughout the Holocene.

The possibility that the *Corylus* rise at Meall Bad á Chrotha was associated with disturbance from the Mesolithic population has to be taken into consideration. The finding of hazelnut shells at a Mesolithic midden at Sand in Wester Ross that had been subjected to heat suggests that the woodlands surrounding the Peninsula were used as a food resource base (Hardy and Wickham-Jones, 2003). It has been suggested that the increase in *Corylus avellana* pollen abundance around 9000 cal BP, which can be seen in many pollen sequences from Scotland, is associated with the burning or coppicing of hazel shrubs. These anthropogenic activities were carried out to increase nut productivity, which was an essential source of food, and to encourage timber development which was used for domestic undertakings. Both burning and coppicing promote the production of *Corylus* pollen and increase flowering activity (Smith, 1970). However, there is debate surrounding the influence of anthropogenic disturbance on the growth of *Corylus*. Huntley (1993) argues that the warming of the climate during the early Holocene was more likely to have contributed to the higher representation of *Corylus* in the pollen records and that human disturbance would have been minimal. Although, in many sequences the hazel rise coincides with charcoal occurrences, these occurrences are insignificant in abundance which contradicts the theory that this taxon was an important aspect within the fire regime (Edwards, 1990). Indeed the lack of charcoal in the Meall Bad á Chrotha record during the early Holocene and the MTWA reconstruction support the suggestion that climate change rather than anthropogenic activity initiated any *Corylus* rise.

Comparable to numerous records from Northwestern and northern Scotland the charcoal occurrences found throughout the early part of the record from Meall Bad á Chrotha remain low with no significant peaks or continued presence. The charcoal recorded is most likely 'background' occurrences deposited at the site from a source elsewhere in the catchment (Clark, 1988), there is also no correlation between the charcoal frequencies and any vegetation change. Separating fires of natural origin from human origin is difficult; but the lack of supporting evidence for anthropogenic burning indicates that the charcoal occurrences recorded during the early Holocene were most likely caused by spontaneous wildfires. Wildfires can be caused by lightning strikes in dry tundra like conditions (Riezebos and Slotboom, 1984; Kolstrup, 1992). The lack of variation in woodland dynamics indicate that fire was not being used to control or manage the woodland, therefore supporting the assumption that any early charcoal occurrences were most probably of natural origin somewhere within the catchment. Indeed, an increase in the level of aridity due to a warming climate could well have promoted small, isolated, natural woodland fires (Bradshaw et al, 1996: Brown, 1997).

There is little indication from the Meall Bad á Chrotha record that suggests Mesolithic communities had any profound impact directly at the site; however this does not suggest that there was not any anthropogenic activity occurring in the area at this time. There is a very minor occurrence of *Plantago lanceolata* recorded at 420cm at Meall Bad á Chrotha which is commonly used as an indicator of anthropogenic activity; however, it can become naturally established and colonize maritime grassland communities, and this is the case on the Isle of Lewis from around 9000 years ago (Fossitt, 1996). The minor level of *Plantago lanceolata* found at Meall Bad á Chrotha suggests that this was also a natural occurrence; although, there is substantial evidence that supports the presence of Mesolithic populations in Wester Ross it is highly unlikely that the small level found represents early disturbance in these exposed areas. On the South Erradale peninsula itself, where Meall Bad á Chrotha is located, a Mesolithic site has been identified at Redpoint (Gray, 1960) and in Sheildaig there is also evidence of human occupation during this period (Walker, 1973). Further sites have been identified by Hardy and Wickham-Jones (2003) within the Inner Sound at Loch a Sguirr, An Corran and at Sand on the Applecross Peninsula which is immediately adjacent to South Erradale.

The Inner Sound is an ideal area for the occurrence of anthropogenic activity during the early Holocene. The marine topography allows for movement and settlement; two aspects which are essential for the survival of nomadic hunter gatherers (Hardy and Wickham-Jones, 2003). There are salt and fresh water as well as upland and lowlands areas which provide a wide range of rich bases for communities to exploit. This area is also suitable for easy travel and contact between people, whether it be by water or land; however, these aspects of past anthropogenic activity are notoriously hard to reconstruct as very little trace is left on the landscape.

A number of sites survive from the early Holocene within The Inner Sound. The Mesolithic age midden found at Sand on the Applecross peninsula could well date back to 8200 cal BP (Hardy and Wickham-Jones, 2003). A narrow blade microlithic assemblage as well as a piece of antler harpoon and bone tools were all found in the excavation. The abundance of shellfish found, especially limpets; suggest that they were a primary food source, contesting the suggestion that limpets were only used for bait. Not only limpets but the level of mussels, winkles, dog whelk, cockles and crustacean found indicate that shellfish were eaten directly by the Mesolithic inhabitants of Sand (Hardy and Wickham-Jones, 2003). Sand is directly on the coastline so the sea would have been an essential aspect and resource within the life of any Mesolithic inhabitants. Although, the abundance of animal and fish bones also found in the midden was substantial, demonstrating the large range of resources that the region of Wester Ross and the Inner Sound has to offer. The excavation also supports the use of heat, particularly for cooking purposes. Pot boilers were found in large quantities, along with stones that had been cracked by fire, and charcoal fragments.

The Mesolithic inhabitants of Sand appear to be advanced with evidence to suggest a large range of activities that took place within the area, such as tool and jewellery manufacture. However, the findings at Sand indicate that the midden was only in use for a few brief seasons. This contrasts to other middens excavated on the west coast of Scotland which appear to have been in use for far longer. The short duration of time over which this midden accumulated may be due to the result of a change in conditions, such as period of famine (Hardy and Wickham-Jones, 2003). If the midden sites were changed seasonally then it would be expected that the Applecross Peninsula and surrounding areas, such as the South Erradale Peninsula where Meall Bad á Chrotha is situated, would be scattered with abundant comparable shell middens. Although there are high numbers middens, they are not all Mesolithic, and if seasonality was an ordinary occurrence then the level of middens would be expected to be much greater. This supports the argument that a change of conditions led to the abandonment of this midden. Edwards (1990) suggests the midden at Sand could date back to the cooling period at 8200 cal BP with abandonment only occurring a season or two later (Hardy and Wickham-Jones, 2003). There is also evidence of a rise in sea levels in the area at this time which would have had a severe impact on the communities that live and rely on the coastline (Hardy and Wickham-Jones, 2003). This climatic deterioration at 8200 cal BP provides an interesting interpretation for the change of conditions, such as the onset of famine, which resulted in the short term accumulation of the midden.

The release of fresh water from the melting of ice bodies disrupted the Thermohaline Ocean Circulation creating reduced sea surface temperatures and an increased level of precipitation surrounding the North Atlantic at 8200 cal BP (Dickson et al, 2002). This contributed to the climate change at 8200 cal BP. Any changes in the North Atlantic would directly affect the west of Scotland, along the coastal zones having substantial impacts on any inhabitants of the area. These changes along with a transition in ocean currents may well have had a negative impact on the population of fish and animals (Lamb, 1979; Cushing, 1982). Formerly reliable food sources may have been altered due to the shifts in the patterns of spawning and migration, and more frequent and severe storms may have impacted upon the coastline (Tipping and Tisdall, 2004), with an increase in the strength of winds and wave dimension (Bacon and Carter, 1991; Gunther et al, 1998). These impacts in turn would have had a negative effect on the rate of erosion and retreat in coastal zones, particularly in the north and west of Scotland where the impacts would have been at their strongest (Keatinge and Dickson, 1979; Gilbertson et al, 1999). Indeed, coastal retreat which occurred at 6000 cal BP can be seen at Skara Brae (Vega Leinert et al, 2000). These shifts in coastal geographies would have also led to a reorganization of the animal and plant communities that inhabit the coastal zone (Rowley-Conwy, 1985); a change in these communities which would have been a central food resource to the Mesolithic population could well have been the cause of famine and the subsequent short term use of the midden at Sand (Hardy and Wickham-Jones, 2003). High resolution data collected from proxies such as trees (Ballie, 1995; Helama et al, 2002; Leuschner et al, 2002) and ice cores

(Dansgaard et al, 1993; O'Brien et al, 1995; Taylor et al, 1993; Alley, 2000b) support the argument that these changes in the climate happened within a very short space of time, with a shift in atmospheric circulation occurring within only one human generation (Tipping and Tisdall, 2004). These abrupt and rapid changes in climate would have had an intense effect on the communities who rely on the coastline for their existence. Indeed there is also evidence for the abandonment of the western Islay coastline at 8200 cal BP (Mithen 2000a, b). Wicks and Mithen (2014) further support the assumption that the climatic shift at 8200 cal BP resulted in change throughout the Mesolithic population in western Scotland. This climatic shift was characterised by a significant temperature reduction of about 6 ± 2^{o}C with the peak of the deteriorated climate lasting between 60-70 years out of a 200 year event. Two theories of population change were put forward as a result of this shift by Wicks and Mithen (2014). The first theory supports the findings from western Islay of abandonment of the islands and western mainland to areas less affected by the increasingly cold and arid conditions, such as river valleys in eastern Scotland or southern Britain. This supports the theory put forward by Gonzalez-Samperiz et al (2009) that the migration of Mesolithic hunter-gatherers in Northeast Spain to the more mountainous humid locations was due to the 8200 cal BP event. However, Wicks and Mithen (2014) argue that the second theory is more plausible. The already low population densities and poor reproduction rates may have been further affected by reduced birth rates and higher mortality directly influenced by the temperature and precipitation drop. Even a slight reduction in population would initiate a substantial collapse due to their already low numbers. There are also other factors such as the scarce availability of food, raw materials and medicinal plants that could have indirectly influenced mortality and reproduction rates (Wicks and Mithen, 2014).

The closing of the Mesolithic period occurred roughly around the same time as the 6000 cal BP climate shift (Bond et al, 1997, 2001). This climatic transition coincided with a shift in the food resources of the late Mesolithic to early Neolithic communities. The previous reliance on marine and coastal resources shifted to terrestrial resources, most likely with an increasing focus on domesticated animals

(Bonsall et al, 2002). This is especially evident on the west coast of Scotland (Schulting, 1998; Richards and Hedges, 1999a, b; Bonsall et al, 2002).

There are some early brief occurrences of *Alnus* recorded at Meall Bad á Chrotha in comparison to the earliest occurrence of *Alnus* in the region of 8250 cal BP at Glen Carron (Anderson, 1998). However, these early occurrences are only a minor feature of the landscape not rising above 5% TLP. Reidchalmai in Sutherland demonstrates a similar appearance of *Alnus* dated to 11,000 – 10,000 cal BP. Tipping and McCulloch (2003) suggested that this early appearance could be down to a premature colonisation of *Alnus glutinosa*, the only species of *Alnus* inhabiting Britain presently. Although, the rapid colonisation of *Alnus incana* (grey alder) in the south of Scandinavia very early in the Holocene has also been associated with an early expansion of *Alnus* in Scotland. It is not known for sure whether *Alnus incana* crossed the northern narrow section of the North Sea (Coles, 1998) and briefly grew in the British Isles, however the theory cannot be dismissed as the pollen morphologies of the two taxa are very similar, therefore cannot easily be identified (Tipping and McCulloch, 2003; Erdtman et al, 1961; Moore et al, 1991).

In the south and central areas of Scotland, *Ulmus* had arrived and became a dominant aspect of the landscape by 10,000 cal BP (Walker, 1984; Whittington et al, 1991b). By 7800BP, *Quercus* had joined *Ulmus*, both thriving in the low areas in Aberdeenshire (Vasari and Vasari, 1968). They spread up to the edge of the highlands, successfully growing in the valleys (Donner, 1982). There is also evidence to suggest that both *Ulmus* and *Quercus* occurred in the Inner Hebrides, in sheltered, isolated pockets (Birks and Williams, 1983; Robinson and Dickson, 1988). The northern highlands did not see the development of these trees until around 9500 cal BP (Tipping, 2003). At three sites on the Isle of Skye, Inver Aulavaig, Talisker Bay and Peinchorran their rise was dated a couple of hundred years earlier although at very low frequencies (Selby, 1997). They are very slow at colonising the landscape and demand a nutrient-rich soil unlike *Betula* and *Corylus* which are first to colonize and can spread rapidly on soil which is relatively poor in nutrients. The presence of these taxa at Meall Bad á Chrotha is so low that it can be assumed that they were not growing immediately around the site. These low levels can also be found at Glen Torridon and Glen Carron (Anderson 1998). Lowe (1993) assumed that such low percentages, which are characteristic of many pollen diagrams from areas such as the Northwest, the Scottish islands and the far north, that the trees were not established in such exposed areas. On the other hand, Bennett et al (1990, 1992) proposed that woodlands consisting of elm and oak were present in the south of Uist and perhaps in the Shetland Isles. It can be assumed that if *Ulmus* and *Quercus* were present in these areas that they would be restricted to isolated pockets in sheltered vicinities. Indeed, it is possible that these taxa were present in one of the more sheltered valleys within the same catchment as Meall Bad á Chrotha. Therefore, the low values represent pollen which has travelled from a more isolated source. Both taxa remain poorly represented throughout the Northwest and northern Scotland during the Holocene (Tipping, 2003).

There is a distinct transition to a drier peat surface during MBAC 3 at Meall Bad á Chrotha with a significant reduction in the contribution of wet indicators such as *Sphagnum* and Cyperaceae to the record. The small bog pools most likely dried out during this period along with a decrease in the peat formation. At Farlary, near Golspie in Sutherland, a similar vegetation succession accompanied by a transition to drier bog surfaces is associated with the 8200 cal BP climatic event (Tipping et al, 2007). It is possible that the similarities seen at Meall Bad á Chrotha also signify this climatic excursion, however this is only an assumption as the Meall Bad á Chrotha core has not been radiocarbon dated, therefore there is no reliable chronology. The 8200 cal BP shift to climatic deterioration was hemispheric in scale and the largest individual climatic event within the North Atlantic region during the Holocene. North Sea marine sediments provide evidence for a 2°C temperature reduction at this time (Klitgaard-Kristensen et al, 1998), with numerous other sets of data, collected from various proxies also supporting a shift to a period of reduced temperature and/or a decrease in the level of precipitation (Klitgaard-Kristensen et al, 1998; Nesji and Dahl, 2001). The MTWA reconstruction from Meall Bad á Chrotha does not support the theory that a reduction in temperature characterised this climatic excursion at 8200 cal BP, in fact there is a steady rising trend in the MTWA as the peat surfaces dried. The records from Farlary do not provide any evidence to suggest whether it was an increase in temperature which resulted in a rise in evaporation rates or a decrease in temperature which led to the drying out of the peat surface (Tipping et al, 2007). The development of herbaceous peat and wood after the *Sphagnum* decline at Meall Bad á Chrotha could well date to around 7900 cal BP if the correlation to the record from Farlary is correct. Hallsdottir (1996) proposed that the 8200 cal BP climatic event caused the collapse of birch woodland in northern Iceland. There is no evidence at Meall Bad á Chrotha, Loch Farlary or any other northern Scottish sites which support this proposal (Tipping, 1994; Davies, 1999). In fact, there is evidence at Meall Bad á Chrotha that the reduction in *Sphagnum* allowed for the expansion of woodland, with increases in *Betula* and *Corylus/Myrica* (presumably *Corylus*) along with abundant Ericaceous dwarf shrubs (*Calluna vulgaris*) in the understory. Farlary contrasts with other studies which also record the 8200 cal BP climate shift. The evidence from Farlary suggests that the level of bog surface wetness decreased and shifted to a markedly drier condition for more than 2000 cal years, whereas further studies from different sites suggest a short lived 200 year climatic excursion (Tipping et al, 2007). At Meall Bad á Chrotha, the palynological evidence supports Tipping et al's (2007) theory of a prolonged drier phase as there is a continuation of dry indicators up until the beginning of the pine decline which is estimated to have occurred around 4000 years ago at the site. This suggests a relatively long dry period, perhaps with a few minor, local variations in the hydrology of the catchment in between, but nothing significant enough to show an evident response within vegetation patterns.

A rise in *Alnus* begins at 280cm, remaining around 10% TLP, apart from a slight peak to 15% TLP at 150cm. In many vegetation successions from northern Scotland an *Alnus* rise can be witnessed, however there is a broad timescale as to when this rise in *Alnus* occurred and to what impact the rise had on the landscape in different regions (O'Sullivan, 1975; Walker and Lowe, 1981; Chambers and Elliot, 1989). At the Loch of Winless, in Caithness in the Northeast of Scotland, the initial rise was dated to 7745 cal BP with the maximum of 6% TLP being reached at 2800 cal BP, whilst at Lochan an Druim, in the extreme Northwest, *Alnus* peaked at 4800 cal BP after the earlier rise was dated at 6529 cal BP (Charman, 1994). At Glen Torridon, the rise is dated approximately to 7250 cal BP, while at Glen Carron the rise is slightly earlier 7750 cal BP (Anderson, 1998). A profile from Eilean Subhainn, an island on Loch Maree, Wester Ross, dates the *Alnus* rise later, at around 7500 cal BP (Kerslake, 1982). Indeed, Lowe (1993) proposes that by 7500 cal BP *Alnus* was a feature in most areas of Scotland. Similarly to Meall Bad á Chrotha, the *Alnus* rises recorded at Glen Torridon and Glen Carron begin as the rise in *Pinus sylvestris* occurs. *Alnus* competed successfully in the lowland deciduous woodlands and the highland *Pinus* woodlands (Lowe, 1993). The percentages of *Alnus* do remain fairly low in the majority of pollen diagrams from Scotland; Meall Bad á Chrotha is no exception. This abrupt arrival of *Alnus* in the pollen records throughout Scotland is an indication of a hydrological shift as alder has a high tolerance to wet conditions (Lowe, 1993). *Alnus* is considered to be a pioneer taxon which prefers coastal locations; these factors could well be the reason for the early occurrence of the taxon in particular locations (Birks, 1989; Hirons and Edwards, 1990; Lowe, 1993). However, these low percentages suggest that *Alnus* was restricted to damper soils around river valleys or around lochs such as Loch Bad á Chrotha or Loch Clair which are in relatively close proximity to Meall Bad á Chrotha. *Alnus* seedlings are drought sensitive and unless there is water to transport the seeds, they are not dispersed far from the parent tree (McVean, 1963). Therefore, it is likely that *Alnus* only expanded locally near the lochans. The lack of synchronicity in the arrival and expansion of *Alnus* throughout Scotland suggests the species was influenced by not only regional climate changes but also local factors (O'Sullivan, 1975; Whittington et al, 1991b). However, the possibility that *Alnus* was able to establish itself within the highlands as a result of land disturbance through human activity should be taken into consideration. *Alnus* is not a successful competitor therefore the clearance of other tree species, whether it is through fire or the grazing of animals would allow the establishment of *Alnus* through a reduction in the number of competitors. However, there is little evidence for such activity at Meall Bad á Chrotha. There are small occurrences of charcoal during the early Holocene but these are most likely 'background' levels either from natural fires due to the increase in aridity or lightning strikes, or small isolated fires set by hunter-gatherers as a means of land clearance somewhere within the catchment. The origin of early fires within the highlands remains debatable (Robinson, 1987; Charman, 1992; Edwards, Whittington and Hirons, 1995; Tipping, 1996; Davies, 1999).

6.1.2 The Pine Rise

There are a number of vegetational changes at Meall Bad á Chrotha that are consistent with those from other sites within the Northwest of Scotland, such as Glen Carron and Glen Torridon. This allows the calibrated radiocarbon dates of particular events, such as the pine rise and subsequent decline, or any hydrological shifts, to be transferred to what is believed to be the corresponding events at Meall Bad á Chrotha. However, as is the same with any study of this type, when radiocarbon date analysis has not been carried out, caution must be taken when correlating dates from similar diagrams, as similar events can occur throughout a region, although, not necessarily in synchronicity, particularly if the changes are caused by human activity or local influences have played a role.

Both birch and hazel show a rapid decline at 350cm, immediately before the beginning of the pine rise at Meall Bad á Chrotha, which ultimately led to the development of dense pine woodland. Indeed, the replacement of birch/hazel woodland throughout Scotland by scots pine is a typical and distinct aspect of any early Holocene vegetation succession. Pollen diagrams indicate that this replacement occurred earlier throughout Wester Ross, and particularly the Loch Maree catchment than in any other parts of Scotland (Birks, 1989).

At 290cm, *Pinus sylvestris* exceeds the 20% TLP threshold, which was proposed as an indicator for local pine growth by Bennett (1984). This threshold for the growth of pine was initiated due to the large quantity of pollen that is produced by the taxon. The large quantities produced can suppress the total land pollen percentages of other taxa, although the taxa never declined. This could well be the case for the percentages of other types of arboreal pollen within the Meall Bad á Chrotha record. However, the possibility of out-competition and the decline of tree taxa due to the shade must be considered. Indeed, pine peaked to over 80% TLP which indicates a very dense forest canopy. The percentages of *Betula, Alnus* and *Corylus* show slight declines as pine peaks to its highest value; before rising again as the pine pollen percentages fall. This dense canopy would prevent the establishment of new trees, in particular *Corylus* which is a light-demanding taxon. The percentage of *Quercus* also increases, although very slightly, as pine declines. *Quercus*, similar to *Corylus* is also a very light-demanding taxon.

Within the Loch Maree catchment in Wester Ross, the date for the pine expansion is between 8900 and 8250 ^{14}C years BP (Birks, 1996). However, outside of Loch Maree there are contrasting dates for this expansion which indicate an asynchronous and uneven colonisation of the species throughout not only Wester Ross but Scotland as a whole. 7424 cal BP is given for the increase in pine pollen at Loch Clair (Pennington et al, 1972), while a date of 8707 cal BP is given for the rise at Loch Sionascaig further north in Sutherland. The date given for the spread and establishment of pine at Glen Torridon is 7490 cal BP, although it passed 20% TLP at 7250 cal BP; while at Glen Carron, 7580 cal BP was given for pine expansion (Anderson, 1996, 1998). These dates indicate that even though there was some variability in the timing of the pine rise in different catchments, by 7000 cal BP the majority of Wester Ross, including Meall Bad á Chrotha, would most probably have been densely forested by *Pinus sylvestris.*

This expansion of *Pinus sylvestris* and the associated rise in MTWA most likely represent the establishment of the 'Mid-Holocene Thermal Maximum'. It appears that pine growth and MTWA have a very close relationship demonstrating the sensitivity of pine to the smallest of fluctuations in climatic parameters. However, arguing that changes in pine pollen match with the temperature changes is a circular argument as the MTWA plot is driven by the pine pollen curve, therefore the fossil assemblage will be primarily compared with independent climate reconstructions. At Meall Bad á Chrotha pine shows a significant rising trend as there is a sharp increase in MTWA. The MTWA rises from 11.5^{o}C to around 16^{o}C at Meall Bad á Chrotha, around 4-5^{o}C higher than today's values. Seppa and Birks (2001) collected data from northern Finland that suggests July temperatures exceeded 13^{o}C during the Holocene thermal maximum, around 1.6-1.7^{o}C higher than modern day values. In southern and central Sweden temperatures during the mid-Holocene maxima were around 2-2.5^{o}C higher than present (Antonsson, 2006);

while at Dalmuttladdo, northern Norway Bjune et al (2004) also suggests that July's mean temperature was around 2^{o}C higher than present. The pine expansion at Dalmutladdo began at 7000 cal BP when temperatures rose producing the warm and dry conditions required for pine growth (Bartholin and Karlen, 1983; Briffa et al, 1988; Hicks, 2001). Indeed, summer temperature and the subsequent reduction in waterlogging is almost certainly the central controlling factor in the germination, growth and survival of Scots pine (Vorren et al, 1996). This thermal maximum allowed for the movement of the maximum tree-line upslope; Antonsson (2006) suggests that the tree-line expanded around 250-300 metres upslope in southern and central Sweden due to the temperature rise. There are contrasting dates for this peak in summer temperatures, although most lie between 7000 and 6000 cal BP within northern Europe (Davies et al, 2003). However, Birks and Seppa (2001) argue using pollen data that the maximum occurred between 7900 and 6700 cal BP. It is possible that the timing of the Holocene Thermal Maximum varied between localities due to local climatic influences.

At Meall Bad á Chrotha, the warming trend shown in the MTWA reconstruction is supported by a decrease in the level of wet indicators and drier bog surface conditions immediately before the pine rise. Reduced levels of precipitation as well as enhanced evaporation rates from the warmer temperatures may have also contributed to the drier surface conditions. Cyperaceae which is a dominant taxon in damper environments disappears from the record at the start of the pine rise and remains at very low abundances throughout the period of pine dominance. *Sphagnum* also decreases to its lowest levels just before the pine rise further suggesting a drying out of the peat surface. Indeed, the growth of pine is encouraged by a reduction in waterlogged conditions (Bell and Tallis, 1974). Extensive areas of *Sphagnum* in a catchment can prevent the establishment and growth of pine seedlings as they become bogged down due to the ability of *Sphagnum* to grow higher than any tree seedlings (Ohlson et al, 2001). On the other hand, if enough pine trees do succeed and grow to an adequate size within an environment where *Sphagnum* is dominant, it is possible that the establishment of these pines will eradicate the *Sphagnum* community through the alteration of hydrology and nutrient levels (Holling, 1992).

To help infer any palaeohydrological fluctuations at Eilean Subhainn, an island on Loch Maree, a C:N ratio was produced using the measurement of total carbon and nitrogen. This ratio curve and light transmittance data indicate that reasonably dry bog conditions prevailed from the early to mid-Holocene (Anderson, 1998). There is also a substantial rise in dry indicators at Cross Lochs in eastern Sutherland at the time of the pine rise. A relationship between pine and Ericaceous dwarf shrubs is evident; a decline in Ericaceae can be matched with the pine rise at Meall Bad á Chrotha. This relationship can be seen in many Holocene palynological reconstructions from Northwest Europe. The substantial decline in Ericaceae pollen at the same time as *Pinus sylvestris* shows a rise could be due to the increased shading from the pine canopies which would ultimately reduce the flowering of Ericaceae present in the under-storey (Charman, 1990). However, it is more likely that this relationship between pine and Ericaceae reflects a reduction in peat-forming conditions due to the drier mire surface. The peat is more humified during times of decreased bog surface wetness as more decomposition occurs when the mire surface is dry (Blackford, 2000).

At Meall Bad á Chrotha, there is a substantial decline in the level of pine pollen at 270cm which coincides with a rise in the percentage of Ericaceous dwarf shrubs. There is also evidence to suggest that there was a higher frequency of small scale fires immediately before and during this decline in pine pollen abundance. Scots pine began to rise at Loch Farlary, near Golspie in Sutherland, during an intensely dry period between 7600and 7500 cal BP (Tipping et al, 2007). However, between 7000 and 6500 cal BP, a period of increased aridity is recorded. This increased aridity may well be responsible for the fragmentation of deciduous woodland and changing the severity and regularity of fire, which in turn made way for the development and extension of heath (Tipping et al, 2007). Although, the low charcoal levels at Meall Bad á Chrotha do not indicate that the fires were immediate to the site, it is possible that similarly to Loch Farlary, although on a smaller scale, that this intense dry period caused a reduction in *Pinus sylvestris* somewhere within the catchment which encouraged the expansion of heath through alterations in the fire regime. On the other hand, the MTWA reconstructions at Meall Bad á Chrotha show a sharp decline of about 2^{o}C during this period. This could well indicate that there was a short lived climatic deterioration which resulted in a brief reduction in pine growth. This could correlate to the wet shift inferred at Eilean Subhainn, Loch Maree at 6500-6040 cal BP which lasted for 200 years (Anderson, 1998). However, this disturbance of pine woodland at Meall Bad á Chrotha was very short lived with pine soon peaking back up to higher values in accordance with a decline in the abundance of Ericaceous dwarf shrubs.

The pine woodlands of the Scottish highlands may well have been encouraged and shaped through the occurrence of natural fires (Durno and McVean, 1959; McVean, 1963; Rackham, 1986). Hancock et al (2005) produced data to suggest that the burning of Ericaceous communities promoted the establishment of pine seedlings by as much as four times more than if the ground had not been burnt previously. The record from Meall Bad á Chrotha contradicts the hypothesis that fire was responsible for the dynamics of *Pinus sylvestris* throughout the Holocene in northern Scotland. It would be assumed that if fire was needed for the establishment of suitable conditions in which pine seeds can successfully regenerate, then the level of charcoal would peak immediately before a rise in *Pinus sylvestris* (Froyd, 2006). There is a very slight increase in the level and frequency of charcoal found at Meall Bad á Chrotha at the commencement of the pine rise, which could indicate a minor influence of fire on the expansion of pine at the site. However, the levels of charcoal found remain so low, fire could not have played any more than very minor role in pine regeneration; the drying out of the peat surface would most likely have played the dominant, primary role. Furthermore, if fire was necessary for the continuation of pine supremacy, through the expulsion of pine competitors which are more intolerant to fire, then it would be assumed that past peaks in the population of *Pinus sylvestris* would correspond with increased levels of charcoal (Froyd, 2006). This is not the case at Meall Bad á

Chrotha with charcoal reaching its lowest levels and disappearing for the record during the height of pine dominance.

Although, there is a lack of any substantial evidence to indicate that late Mesolithic communities inhabited the land on Meall Bad á Chrotha, there is evidence of anthropogenic activity at this period in the region and even on the South Erradale peninsula with a midden discovered at Redpoint. Therefore, the possibility that the charcoal occurrences recorded are due to human interference must be considered. There is no distinct evidence for woodland clearance through the use of fire at Meall Bad á Chrotha, although Charman (1990) suggests that burning vegetation would have been likely from as early as 8714 cal BP, and in eastern Sutherland, the early spread of blanket peat indicates that burning by Mesolithic or early Neolithic populations was happening well before 6693 cal BP. There is a substantial and abrupt drop in the abundance of *Betula* and *Corylus/Myrica* immediately before the start of the pine rise. This does coincide with an increase in the frequency and level of charcoal concentrations, which may indicate an intensification of the fire regime in the area. However, the quantity of charcoal found is very minor; therefore, fire intensity would have still been low and not immediately local to Meall Bad á Chrotha. At Cross Lochs in Sutherland, the abundances of *Betula* were linked to the fire regime before 7000 cal BP (Charman, 1990, 1994). Indeed, both animals and vegetation could have been managed and controlled through the use of fire by Mesolithic communities (Smith, 1970; Jacobi et al, 1976; Mellars, 1976; Simmons et al, 1981). The burning of shrubs would allow animals to graze the land on a small scale. However, there is no evidence of a relationship between the fire regime and shrub abundance at Meall Bad á Chrotha, and although there is a decline in the level of both *Betula* and *Corylus/Myrica*, it is highly unlikely that this was related to the fire regime. The decline of these tree species and their subsequent replacement by pine woodland is a common aspect in the majority of vegetation reconstructions throughout Scotland. The synchronicity and widespread nature of this vegetation succession supports the theory that it came about as a response to the early Holocene climatic warming. It would not be unreasonable to say however, that there were small Mesolithic populations in the area focused in isolated pockets along the coastline, such as at Redpoint, which perhaps did use fire as a method of land clearance although on a very local scale, perhaps contributing to the 'background' charcoal levels witnessed at Meall Bad á Chrotha.

The MTWA and vegetation reconstructions from Meall Bad á Chrotha and independent climate reconstructions all suggest that the early Holocene was a period of increasing warmth and improved conditions, which was possibly responsible for the dynamics and consequent spread of *Pinus sylvestris.* This can be seen not only at Meall Bad á Chrotha but throughout Wester Ross and the majority of northern Scotland; although, there is some variation and asynchronicity in the timing of the pine rise within different catchments which is most likely due to more local influences or a time lag between the climatic event and vegetational response. However, it can be assumed that by 7000 years ago Meall Bad á Chrotha was densely forested by *Pinus sylvestris.*

6.2 Mid-Holocene: Deforestation and the Pine Decline

The transition from dense pine woodland to an open landscape dominated by shrubs and herbs is evident at Meall Bad á Chrotha. The collapse of woodland during the mid-Holocene is a significant period in Scotland's vegetational history, particularly as it gave way to the landscape we see today. There have been many suggestions as to what caused this collapse, predominantly the rapid decline in *Pinus sylvestris* populations; however, there is substantial evidence within the Meall Bad á Chrotha record to suggest that there was a shift from the warm and dry climatic stability that prevailed during the Holocene Thermal Maximum, to a period of cold and wet instable conditions during the Holocene Thermal Decline.

The pine decline at Meall Bad á Chrotha is rapid and abrupt with a fall from over 80% TLP at 210cm, to 5% TLP at 180cm. Increasing levels of Ericaceae dwarf shrubs, Poaceae and Cyperaceae suggest a transition to increasingly open, wet, peat-forming conditions. The increase in Cyperaceae after the pine decline can be seen in many vegetation successions throughout Scotland, including Cross Lochs (Charman, 1990), Glen Torridon, Glen Carron (Anderson, 1998) and Eilean Subhainn

(Kerslake, 1982). An increase in the level of Cyperaceae is an indicator of a shift to wetter conditions as Cyperaceae typically grows in waterlogged habitats. After the high temperatures associated with the Holocene thermal maximum there is a substantial declining trend in summer temperatures at Meall Bad á Chrotha. This climatic deterioration is known as the 'Holocene thermal decline' or the late Holocene 'neoglacial' (Rosen et al, 2001; Seppa and Birks, 2001, 2002; Korhola et al, 2002). This cooling trend, which occurs as there is a drop in pine pollen percentages, continues to fall below early Holocene temperatures, fluctuating between 9.5-10.5^{o}C during the late Holocene in comparison to the 11.5-12.5^{o}C seen earlier in the record. This supports Bigler et al's (2002) proposal that the coldest period of the Holocene was indeed the last 4000 years. There is around a 4-5^{o}C drop in summer temperatures from the mid-Holocene thermal maximum peak to the temperatures associated with the late Holocene. In Swedish Lapland, a similar decline can be seen with a drop of 2.5-4^{o}C in annual temperature (Shemesh et al, 2001).

In areas where growing conditions are already marginal, such as Meall Bad à Chrotha, any slight change in climatic factors would quickly push any vulnerable taxa, such as *Pinus sylvestris*, past their survival threshold. It is clear that pine is extremely sensitive to the most minimal temperature fluctuations, particularly changes in summer temperature. This can be seen in the distribution of scots pine today which is restricted by temperatures during the summer months (Karlen, 1976). In order for pine seeds to ripen a mean temperature of more than 10.5°C is needed for at least four months during the summer (Carlisle and Brown, 1968). However, if summer temperatures are too high, then when winter temperatures begin, the pine seedlings will not be adapted to frost conditions. Temperature reductions can be just as damaging to pine growth (Kullman, 1983b). The production, germination and establishment of seedlings can be severely inhibited if temperatures are too low (Kullman, 1983b). Although, the main effect of any temperature reduction would be on evapotranspiration rates which would in turn affect the hydrological state of the peat surface; therefore, increasing the level of waterlogging and making conditions unfavourable for pine establishment and growth (Gear, 1989). This is also the case for a rise in precipitation levels. Around 4000 cal BP there was a marked changed in the altitudinal and latitudinal range of *Pinus sylvestris* throughout not only Northwest Scotland but much of Northwest Europe (Gear, 1989). This change in distribution and transition to wetter conditions can be seen by the preservation of pine macrofossils in Scottish blanket peats (Pears, 1972; Birks, 1975; Dubois and Ferguson, 1985). The covering of blanket peat over the buried pine stumps indicates that a climatic deterioration resulted in enhanced peat growth preventing pine regeneration (Lowe, 1993). The preservation of plant remains, such as pine macrofossils, occurs when an increase in water table height reduces the amount of time for aerobic decay of plant matter to occur in the acrotelm layer, therefore lowering the rates of decomposition (Ingram, 1978).

There is a fairly broad range of dates for the pine decline both regionally and locally. The decline at Glen Torridon is dated to 4410 cal BP, while at Glen Carron it is 3830 cal BP (Anderson, 1998). Birks (1972) gave a fairly early date for the decline, 4780 cal BP at Loch Maree, while Kerslake (1982) gave a much later date of 4043 cal BP. This is most likely due to local factors influencing the growing limits of the taxon therefore thresholds are past at varying times. It is also possible that in some cases the time lag between the climatic event and the vegetational response will alter at different locations due to the contrasting soil type, underlying substrate and topography. However, throughout Wester Ross and northern Scotland, many peat sequences support the theory of a two phase decline in the abundance of pine pollen. Glen Carron and Eilean Subhainn both exhibit this two phase decline (Anderson, 1998), as well as sites in the west of Sutherland, Lochan Dubh (Kerslake, 1982) and Loch Sionascaig (Pennington et al, 1972). There appears to be a reasonably early initial decline in pine pollen which is followed by stagnation in the declining trend or even a regeneration of the taxon. This is ultimately followed by a final collapse in pine values to critically low values. In both the Glen Torridon and the Glen Carron cores a shift from wet to dry and back to wet conditions can be detected between 5000 and 4000 cal BP (Anderson, 1998).

It has been suggested that this hydrological pattern may well be intimately linked to the fluctuations and ultimate fate of Scots pine woodland during the mid-Holocene (Anderson, 1998). Evidence for climatic deterioration comes from two deep sea sediment cores from the North Atlantic; one near Ireland and one off the Greenland coast (Bond et al, 1997). The cores expose the numerous ice rafted debris events that occurred throughout the Holocene at 10,300, 11,100, 9400, 8100, 5900, 4200, 2800 and 1400 cal BP (Bond et al, 1997). These events would have produced short-lived reductions in sea surface temperature and marked a period when cold, ice packed waters migrated southward from a Greenland source (Anderson, 1998). Evidence for such events and cooling in sea surface temperature primarily come from the abundance of planktonic foraminiferal. In particular, a significant rise in the taxon *Globigerina quinqueloba* indicates an intensely cool period, as this taxon is dominant today in the Arctic waters to the north of Iceland (Bond et al, 1997). Warm water taxa of planktonic foraminiferal such as *N. pachyderma* show prominent decreases during the ice rafted debris events of the mid-late Holocene. There is evidence to suggest that an event such as this occurred at c. 5900 cal BP. This date would approximately coincide with a transition to reduced sea surface temperatures recorded by diatoms at c. 5700 cal BP in the Nordic seas (Koç et al, 1993; Koç and Jansen, 1994; Koç et al, 1996). It has been suggested that this climatic deterioration around 5000 cal BP combined with the expansion of blanket peat and a rise in the level of soil paludification pushed *Pinus sylvestris* past its survival limits, therefore initiating the first and most significant fall in pine values (Anderson, 1998).

However, a transition back to drier bog surface conditions around 4500 cal BP put a halt to this initial pine decline. In both the Eilean Subhainn and Glen Torridon core this change back to improved conditions came after the central decline. Neither of these sites demonstrates a significant rise in pine pollen during this dry phase. This indicates that the shift to drier conditions was due to a change in climate as opposed to local conditions such as an expansion of pine woodland in close proximity to the bogs which in turn would produce higher rates of evapotranspiration and a subsequent local reduction in the depth of the water table (Anderson, 1998). Although, at Glen Carron there is evidence for a rise in pine pollen abundance during these improved climatic conditions. Indeed, a rise would be expected as pine thrives on drier bog surfaces. It is possible that at Eilean Subhainn and Glen Torridon, the development of blanket peat had already taken hold of these catchments, making it very difficult for anything other than very minor rises in pine to occur during the dry phase; perhaps only regenerating on the last remaining dry patches available and only producing low pollen quantities due to its marginal growing conditions (Anderson, 1998). There is also data to support short-lived increases in pine pollen percentages just before its concluding decline around 4000 cal BP from south Eastern Skye (Birks and Williams, 1983) and Sutherland (Charman, 1994). Indeed, the data from Eilean Subhainn and Glen Carron, as well as other sites in the west of Scotland, suggest that this dry shift around 4500 cal BP, slowed the decline, until a further climatic deterioration and consequent wet shift, pushed the remaining pine trees past their survival limits at c. 4000 cal BP. Charman (1990) also suggests that a regional wet shift occurred at 4372 cal BP, which coincides with the pine decline further north at Cross Lochs and the widespread decline to the south and southwest (Birks, 1975; Pennington et al, 1972). In comparison to the records from Glen Carron and Eilean Subhainn, there is no evidence at Meall Bad á Chrotha of a two phase decline in *Pinus sylvestris*. Although, there is a slight increase at 170cm after the main pine decline, this increase does not reach the 20% TLP threshold proposed as an indicator for local pine growth by Bennett (1984). However, there is a coincident fall in the percentage of Cyperaceae and a minor rise in the MTWA which could indicate a possible short-lived shift back to the warmer and drier bog conditions witnessed before the climatic deterioration.

This second shift to a period of increased bog surface wetness can be correlated to the ice rafted debris event that occurred at 4200 cal BP (Bond event number 3) resulting in a reduction in sea surface temperatures dropping to a minimum just after 4000 cal BP (Bond et al, 1997). These two periods of reduced sea surface temperatures divided by a short-lived period of warmer sea surface temperatures in the North Atlantic (Bond et al, 1997) correlate to the wet, to dry and back to wet bog surface conditions that are inferred from the sequences within Wester Ross (Anderson, 1998). However, there are some discrepancies when correlating the timing of these ice rafted debris events to the vegetation sequences and hydrological changes recorded in Wester Ross. For instance, the timing of the ice rafted debris event at 4200 cal BP coincides with a shift to drier peat surfaces not an increase in wetness. This can be explained by a time lag between the ice rafted debris event and the response of the peatland lasted a few hundred years. Therefore, it is highly plausible that this wet shift dated between 4020 and 3630 cal BP, which is inferred from the Wester Ross peat cores (Anderson, 1996, 1998) and can be seen in the vegetational succession at Meall Bad á Chrotha, was related to the event in the North Atlantic at c. 4200 cal BP. This is supported by Bond et al (1997) who provide evidence from planktonic foraminifera that the reduced sea surface temperatures occurred a few hundred years after the ice rafted debris event itself which in turn would prolong the response expected from the peatland after the event. Indeed, this time lag can also be seen with the initial event at 5900 cal BP, which resulted in the first and central pine decline (Anderson, 1998). Throughout Scotland, changes in bog vegetation have been detected and linked to increased precipitation and a decrease in evaporation caused by the North Atlantic Oscillation (Crawford, 2000).

A multiple proxy investigation carried out by Jessen et al (2005) at Lake Igelstön, southern Sweden, also shows two phases of climatic excursions after the Holocene Thermal Maximum. An increasing level of organic-producing algae and a decline in carbonate-producing algae, as well as higher erosion rates within the catchment supports the stable oxygen isotope data ($\delta^{18}O_{sed}$) which suggests a major climatic transition to a period of wetter and cooler conditions. The date given for the initiation of this climatic downturn was between 4600 to 4450 cal BP, with the first phase dated between 4450 to 4350 cal BP. This phase is characterised by an overall decrease in pollen productivity and in particular collapse in the *Corylus* abundance at Lake Igelsjön, a reduction in Calcium carbonate content and a high rise in lake levels indicated by the $\delta^{18}O_{sed}$ data (Jessen et al, 2005). Following this is a period of 250 years in which there is a cessation in the climatic deterioration trend. However, this is relatively short-lived with a return back to the increasingly wet and cool conditions between 4100 to 3800 cal BP. A reduction in $\delta^{18}O_{sed}$ indicates that there was a rise in effective moisture levels peaking at 4010 cal BP, which was followed by a drop in Organic carbon content at 3800 cal BP (Jessen et al, 2005). At 4050 cal BP the overall percentages of pollen show a rapid decline, similarly to the first phase, a significant restructuring of woodland composition and extent occurred. All the data collected from the differing proxies indicate that there were two distinct phases leading towards a period of decreased temperatures and a rise in effective moisture (Jessen et al, 2005). The most extreme period of wet and cold conditions occurred at approximately 3800 cal BP for 100 years, as inferred from lake productivity data. However, the stable oxygen isotope data suggest that this period of enhanced wettest and reduction in temperatures occurred for far longer, for around 4100 to 3300 cal BP, with the minimum being reached nearer the 3300 cal BP mark. There is further evidence for a climatic event from northern Norway which dates a rapid and major cooling to 3800 cal BP (Lauritzen and Lundberg, 1999), also there were substantial glacial re-advance events in the south of Sweden at 4000 cal BP (Nesje et al, 2001). In northern Sweden pine trees responded rapidly to the climatic cooling and the initiation of unsuitable growing conditions with a reduction in the altitudinal tree-line from as early as 5200 cal BP; a mean growing season temperature drop of 1.5^{o}C can be inferred from the lowering of the tree-line by 175m by 4500 cal BP (Barnekow, 2000). These investigations support the findings from Lake Igelsjön which suggest the transition to a period of increased effective precipitation and/or reduction in temperature and an overall transition to substantial climatic instability. The data collected from Norway and Sweden indicates that there was a reduction in atmospheric CO_2; although this occurred later than would be expected if it was the force behind this climatic transition. Instead, this decrease in atmospheric CO_2 most probably occurred as a response to the changes in the terrestrial biosphere after the climatic downturn (Jessen et al, 2005). Rather the Holocene Thermal Decline is most likely linked to orbital insolation and prolonged variations in incoming solar insolation (Jessen et al, 2005; Bradley, 2003).

Wanner et al (2011) also provide evidence for a series of cold phases throughout the Holocene at 8200, 6300, 4700, 2700, 1550 and 550 cal BP. The Holocene can be divided into three central stages (Wanner et al, 2011). Between 11,700 and 7000 cal BP, the northern hemisphere endured a temperate climate with high summer insolation (Renssen et al, 2009). The second phase is defined as the 'Holocene Thermal Maximum' between 7000 and 4200 cal BP, when high temperatures prevailed throughout the northern hemisphere (Klimenko et al, 1996; Alverson et al, 2003). The final period is referred to as the 'Neoglacial'. Declining summer temperatures and glacier re-advances due to lower insolation throughout the boreal summer in the northern hemisphere characterised this neoglacial period which peaked at 4300 cal BP (Porter and Denton, 1967; Denton and Karlen, 1973).

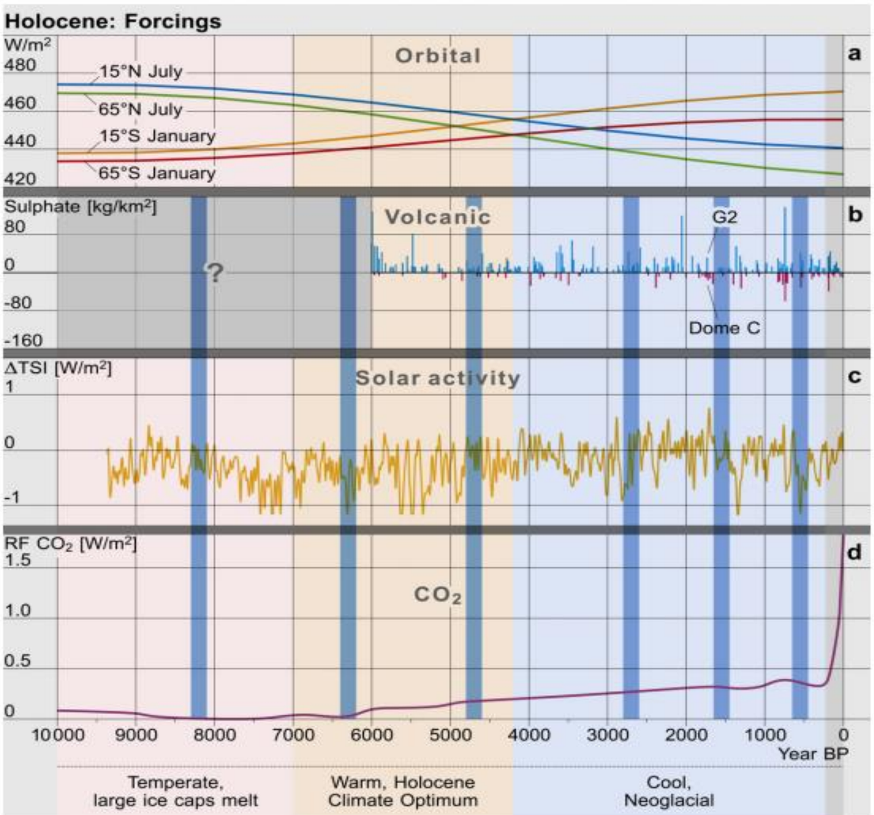


Figure 39. Main forcing's behind climate change during the Holocene. **a)** Orbital changes and solar insolation for two sites in the northern and southern hemisphere (Berger, 1978) **b)** Volcanic activity during the last 6000 years. The vertical bars show sulphate concentrations from ice cores which infer volcanic activity (blue-Greenland, red-Antarctica) **c)** solar activity variations (Steinhilber et al, 2009) **d)** CO_2 fluctuations. Blue bars indicate cold periods (IPCC, 2007) (Wanner et al, 2011).

The phase identified by Wanner et al (2011) between 4800 and 4500 cal BP corresponds with Bond event number 3 at 4200 cal BP (Bond et al, 1997, 2001). Solar activity was average at 4700 cal BP and volcanic activity has been disregarded as an influencing factor in this cold phase as there were no particularly explosive eruptions during this time frame (Wanner et al, 2011). Figure 39 shows the main forcing's behind climate change during the Holocene.

Therefore, Wanner et al (2011) proposed that a lowering of solar insolation along with a downturn in thermohaline circulation were most probably the contributing factors behind this particular climatic excursion which saw the collapse of *Pinus sylvestris* at Meall Bad à Chrotha and throughout northern Scotland.

Olsen et al (2012) also put forward the theory that changes in the North Atlantic circulation played the dominant role in the end of the Holocene thermal maximum which occurred between 5000 and 4500 cal BP. High-resolution geochemical data (Mn/Fe ratio) was produced from a lake in Kangerlussuaq in western Greenland to help reconstruct past variability in the North Atlantic Oscillation and atmospheric circulation patterns. At approximately 4500 cal BP there was a transition from predominantly positive North Atlantic Oscillation conditions to negative. Changes in the Mn/Fe ratio correspond to inflated $CaCO_3$ concentrations which infer a fluctuation in the North Atlantic Oscillation pattern after 4000 cal BP (Olsen et al, 2012). Moros et al (2006) also suggest that changes in the North Atlantic Ocean are behind this mid-Holocene climatic downturn. A surge of fresh water and ice around east Greenland resulted in the relocation and increasing strength of the polar front at the onset of the Neoglacial period at 5500 cal BP and particularly after 4000 cal BP when there was enhanced instability in east Greenland ice rafted debris events. Geirsdottir et al (2013) also propose that variations in the ocean-atmosphere system are the central cause behind climatic cooling and instability after the Holocene Thermal Maximum. Decreases in insolation during the summer months began at 5500 cal BP but a further cooling event was superimposed on this already cooling trend between 4300 to 4000 cal BP, thus resulting in reorganisation and reduction of vegetation throughout the Northern Hemisphere (Geirsdottir et al, 2013). The pine decline witnessed at Meall Bad à Chrotha and throughout northern

Scotland is a prime example of the vegetation response to this climatic deterioration.

It has been suggested that increased storminess could have influenced the collapse of pine populations; although this has long been debated. Changes in vegetation during the mid-Holocene have been linked to an increase in storminess out in the Atlantic Ocean on the isolated island of St Kilda (Walker, 1984). However, a reduction in wind exposure between the period of 9000 and 4500 years ago has been indicated from the continued occurrence of deciduous woodland on the Inner Hebridean Islands (Birks, 1987). The 'Little Ice Age' has been suggested as a time during which the pine populations in the west suffered from the impacts of wind throw (Lamb, 1964; Smout et al, 2005). Subfossil pinewood remains from Clashgour 'A', near Loch Tulla on Rannach Moor in western Scotland, provide evidence that Scots pine adapted to the westerly orientated wind stress (Tipping, 2008). The adaptation of *Pinus sylvestris* to wind exposure occurs in the network of roots under the ground surface (Nicoll and Ray, 1996) and the growth rings, near the base of the tree, increase in thickness (Wilson, 1975). Indeed, around 7400 years ago, winds coming from the west appear to have been as dominant as they currently are (Roy, 1997). The wind is funnelled down the valley of Meall Bad á Chrotha, forcibly from west to east. Similarly, the wind is funnelled down Abhainn Shira valley near Clashgour from west to east (Roy, 1997). The pine populations at Clashgour could well have grown in relatively open woodland which would have increased their vulnerability to wind damage (Tipping, 2008). It was also suggested by Bridge et al (1990) that the pine in the area occurred in scarce density; the data collected from Clashgour suggests the occurrence of < 200 trees per hectare reinforces this theory. However, even though the trees at Clashgour exhibit adaptations to wind stress, the number of trees that fell to the east was higher than what would be expected. The possibility of the fallen trunks being realigned by flowing water has been disregarded. Also, the pinewoods were growing on stable, mineral soil as opposed to a weak peat surface (Tipping, 2008).

Numerous significant storm events with extreme winds coming from the west, most likely defeated them despite their adaptation mechanisms. Also, at Clashgour 'A' there is evidence to suggest that the trees fell after being snapped from above the stump as opposed to below. The evidence combined from Clashgour in western Scotland supports the theory put forward by Tipping (2008) that it was severe winds that resulted in the reduction in scots pine populations at 7400 cal BP, instead of an increase in the level of precipitation which was suggested by Bridge et al (1990). The pine woodland at Rannoch Moor increased in its extent about 7500 cal BP, however between 7300 and 5600 cal BP, these woodlands underwent a substantial reduction, before recovering once again. The final decline in the area of Rannoch Moor is dated at 4500 cal BP. Bridge et al (1990) argued that a rapid increase in the level of precipitation led to the two phase reduction in pine woodland. Stable isotope data collected from the remains of subfossil pine in the Cairngorms (Dubois and Ferguson, 1985) was used to obtain this relationship between the two phases of decline and correlating increases in precipitation by Bridge et al (1990). A likely possibility is that an increase in paludification could have resulted in the weakening of the trees which in turn would have increased their vulnerability to them toppling over in the event of a storm (Tipping, 2008). However, the placement of toppled trunks at Clashgour does not support this theory. These pines show that they were adapted to wind stress from early on in their growth indicating that the winds that could have possibly contributed to their demise were intense and frequent (Tipping, 2008).

Wind throw and the extent to which it contributed to the pine decline in western Scotland remains debatable. The data presented from Clashgour ‘A’ seems to provide adequate evidence which suggests the contribution of wind throw to the pine demise, however a palynological reconstruction from the site shows no declines in the percentage of pine pollen around the time the pine woods were toppled by excessive winds. This suggests that the storms in fact had a minor influence on the pinewoods or that the dispersal of pollen and the strength of the wind, canopies of trees and the pollen sources changed (Bunting and Tipping, 2004). Wind stress could well have played a more substantial part in the collapse of pine throughout Scotland than previously thought, although the stress may have been restricted to the higher altitudes where the wind is more extreme and frequent, particularly on the west coast and in the valleys where the intensity of the wind is inflated due to the funnelling effect. The west facing valley of Meall Bad á Chrotha is high altitude and lends itself to the funnelling effect; therefore it is possible that wind stress could have played a minor part in the pine decline recorded at the site.

Any deterioration in climate or variations in wind strength, precipitation or temperature would have particularly affected Meall Bad á Chrotha due to its high altitude and exposed nature. Tree density and extent on exposed upland areas and wind facing slopes is lower than on low land sheltered areas. The west coast and western isles is predominantly more exposed than further inland on mainland Scotland, as a result, tree composition and extent is considerably different (Birks, 1977, 1988). Fossitt (1996) proposes that altitude rather than latitude has a more controlling influence on the range and density of woodland cover.

However, the suggestion that a climatic deterioration or variation was the fundamental contributing factor in the collapse of pine is poorly supported, as only a hand full of studies have provided evidence other than that of the pollen sequence, for a climatic shift. Although, this theory is supported by the relative synchronicity of the decline throughout much of the northern highlands, it is vital that other possible controls which could have influenced or contributed to the dynamics of *Pinus sylvestris* during the mid-Holocene are taken into consideration.

The majority of Northwest Europe throughout the mid-Holocene was heavily wooded (Bennett, 1989; Tipping, 1994). The relationship between people and these woodlands cannot be ignored. Some catchments may have experienced land use and tree clearance earlier than other localities, therefore influencing pine dynamics before the taxon had a chance to respond to climatic deterioration. The Mesolithic to Neolithic transition which occurred approximately around 6400 and 6000 cal BP (Woodbridge et al, 2012) has been described as a period when people 'turned their back on the sea, to face the land' (Schulting et al, 2002). The reliance on the sea as a food source shifted to land based subsistence such as farming and domesticated animals. However, it has been suggested that on the west coast, this transition occurred over a broad timescale due to the severe climate and isolation of the area which led to the continued use of the sea as a central food source; in comparison to the abrupt and rapid switch to and advancement in farming throughout western Europe during the Mesolithic-Neolithic transition (Schulting et al, 2002).

The mid-Holocene woodland decline seen at Meall Bad á Chrotha is followed by a shift to an open landscape dominated by Cyperaceae and *Calluna* heath. These water-tolerant communities most likely represent the growing patches of blanket peat with dry heathland (Fossitt, 1994a). After the decline of woodland and in particular the collapse of *Pinus sylvestris*, Ericaceous dwarf shrubs, such as *Calluna vulgaris*, would have expanded into the newly open landscape (Pennington et al, 1972). Both Glen Torridon and Glen Carron exhibit similar vegetation shifts. The rise in Cyperaceae is directly coincident with the final pine decline at Meall Bad á Chrotha. A shift to increased bog surface wetness dated between 4070 and 3620 cal BP can be seen in both the Glen Carron and Glen Torridon cores, indicating the onset of a regional wet shift (Anderson, 1998). This regional wet shift relates to the sudden final decline in pine pollen abundance (Anderson, 1998). However, this wet shift is coincident with the arrival of cultural indicators in the pollen records and a rise in charcoal levels in both cores, indicating to some extent the presence of prehistoric land-use and burning within the area. Similarly, at Meall Bad á Chrotha the reduction of pine pollen to critically low values coincides with an increase in the frequency of both *Plantago lanceolata* and charcoal with both becoming continuous components for the remainder of the record. Small scale settlements probably became more permanent within the area around this time which would have resulted in the intensification of subsistence farming and crop production. These indications of anthropogenic activity cause problems when it comes to interpreting any regional wet shifts and vegetational changes. When land clearance is carried out through the use of fire on a significant scale, it can change the dynamics of hydrology within a catchment through the reduction in the permeability of soil and an increase in waterlogging (Anderson, 1998). Throughout upland Britain, the establishment and expansion of blanket peat was most likely encouraged by the hydrological changes associated with the use of fire to aid land clearance (Moore, 1972, 1986a, b). Although it is always important to consider the possibility that anthropogenic activity influenced catchment hydrology particularly in upland areas such as Meall Bad á Chrotha, the likelihood of such activity having an impact is low due to the isolation and inaccessibility of the site. Furthermore, the low abundances of *Plantago lanceolata* and charcoal recorded suggest that the intensity of land use required to initiate such impacts was not happening in close enough proximity to the site. These background levels are likely to have originated from a regional source or perhaps represent the intensification of arable farming within the valley below Meall Bad á Chrotha, where there is ample land suitable for land use, settlement and agricultural practices. If the decline was related to anthropogenic activity then there would widespread, significant evidence for land use and settlement such as extensive grazing by animals, fires or woodland clearance within the Meall Bad à Chrotha record and throughout the rest of northern Scotland. However, the lack of evidence and the inability of prehistoric populations to encourage such a rapid, extensive and synchronous woodland decline make the theory of a climatic downturn more plausible. Furthermore, a continued cooling trend in summer temperatures is evident. The cooler conditions would reduce evaporation rates and therefore change the hydrological status of the bog and encourage the growth of water-tolerant plant communities. It is possible that the inferred development of land use for the minor rise in *Plantago lanceolata* and charcoal occurred as a response to woodland collapse, with populations taking advantage of the newly open environment, as opposed to populations encouraging the woodland decline. However, the evidence for human activity at Meall Bad á Chrotha is rather limited as there is no significant changes in any other taxa which could indicate such activity. Relying on only one taxon, such as *Plantago lanceolata*, especially when the abundance found is still relatively low, is unreliable, therefore, the possibly that this taxon grew naturally within the area must be considered.

There has been much debate as to whether palynological reconstructions at landscape-scale are sensitive enough to represent any disturbances and land use, which may have only been carried out on a small scale, which could well have been characteristic of many landscapes during the mid-Holocene (Davies and Tipping, 1997; Halstead, 2000; Sugita et al, 1997; Willis and Bennett, 1994). Past farming communities typically settled on land which was well drained with adequate soils, which would enable any land use such as the production of crops or grazing. This type of land is found away from the wetland deposits which are required for the preservation of pollen grains necessary for palynological reconstructions (Tipping et al, 2009). These disadvantages associated with the use of palynological analysis in order to determine human activity may have contributed to the low values found at Meall Bad á Chrotha, as further studies carried out within the area and throughout the rest of Scotland show strong evidence for anthropogenic activities and agricultural practices from the mid-Holocene. Indeed, there is evidence for anthropogenic activity more immediately close to the site. Over twenty-two hut circles were identified on the south side of Loch Gairloch, with one being identified at Loch Bad á Chrotha (Dagg, 1998). The majority of these hut circles identified were located on small, cultivable areas of land away from the wet deposits required for pollen preservation.

The theory that a volcanic eruption contributed to the collapse of *Pinus sylvestris* throughout Scotland has long been debated. A rise in the level of bog surface wetness can be caused by a reduction in temperature, evaporation and insolation. A volcanic eruption such as Hekla-4 can cause a volcanic induced climate shift which could, in turn, be responsible for producing these factors which can increase the level of bog surface wetness (Porter, 1981; Sear et al, 1987). Climatic deterioration produced by volcanic dust veils or acid pollution caused by the expulsion of chemicals during an eruption could well be a substantial contributing factor to the pine decline witnessed at Altnabreac (Blackford et al, 1992). The date for the Hekla-4 tephra deposition does correlate to the reduction in pine pollen at a number of sites throughout Scotland; Altnabreac in the region of Caithness exhibits a particularly strong correlation between the collapse of pine pollen and the deposition of volcanic ash, with the percentage of *Pinus* pollen falling to 1% TLP from 17% TLP just above the tephra maximum (Blackford et al, 1992). Tephra particles that have absorbed pollutants or acid rain have been proposed as a potential trigger in the brief decline of pine pollen abundances. Pines are very susceptible to any acidification of the mire surface (Blackford et al, 1992). Greenland ice cores contradict this proposal as they do not show any rise in acidity at 4310 cal BP, although a peak in acidity at 4690 cal BP has been associated with the Hekla-4 eruption (Hall et al, 1994; Hammer, 1984). At Reidchalmai, the Hekla-4 eruption is prominent (Tipping and McCulloch, 2003). A loss-on-ignition signal is determinable as well as an evident rise in glass shards. However, there is no evidence to suggest that any fallout from the eruption contributed to a change in the landscape. Indeed at Reidchalmai, the final decline in *Pinus* pollen occurs at 4310 cal BP, 50 years before the fall of tephra (Tipping and McCulloch, 2003). Blackford et al (1992) claimed that the Hekla-4 eruption initiated the climatic effects that subsequently caused the collapse of *Pinus sylvestris* populations throughout Scotland. It is possible that these volcanic effects influenced the dying off and suffering of trees up to 900km away from the Hekla volcano in the south of Iceland (Blackford et al, 1992).

However, there are numerous data sets that provide more high resolution records that show the pine decline was well underway before the Hekla-4 eruption (Birks, 1994; Hall et al, 1994; Huntley et al, 1997). A shift to a period of increased wetness is evident at all three sites in west Corlea, central Ireland (Caseldine et al, 1998) and there is a growing collection of data to support a shift in the degree of humification at 4000 cal BP throughout much of Northwest Europe (Barber et al, 1994). Caseldine et al (1998) concluded that this heightened surface wetness began immediately before the Hekla-4 eruption; therefore any impact produced by the eruption would have been superimposed on the increased wetness that was already occurring throughout Northwest Europe (Dugmore et al, 1995). There is no evidence to suggest the deposition of volcanic acid at the west Corlea sites; although acid intolerant taxa, such as *Ulmus* and *Fraxinus*, show declines to critically low abundances only a couple of decades after the tephra deposition. These declines however, are more likely to be associated with land clearance through anthropogenic activity, as there are coincident increases in the abundance of herbs which are characteristic in any land clearance practices (Caseldine et al, 1998).

The effects of volcanic eruptions can have pronounced environmental and, in particular, ecological effects not just near the source but on a wider scale (Sheets and Grayson, 1979). If the pine trees were already growing in marginal conditions at the very limits of their ecological range due to increased wetness, then it may be that the eruption at Hekla-4 contributed to the pine decline to some extent at Meall Bad á Chrotha; perhaps by speeding up the rate of decline or pushing the last remaining pine trees past the limits of their range and preventing any regeneration of the taxon.

There has also been the suggestion that fire played an integral role in the dynamics of scots pine throughout the Holocene. If an alteration in fire frequency and/or intensity was the reason for the decline in scots pine throughout Scotland then a substantial variation in the volume of charcoal recorded would be evident within the record. There is an increase in the frequency of charcoal at Meall Bad á Chrotha after the peak in pine pollen however the abundance remains low, suggesting that if fires did occur in the area they were not necessarily immediate to the site. The distance that microscopic charcoal fragments can be transported is substantial (Chandler et al, 1983; Clark, 1988a), therefore low abundances of charcoal could represent the occurrence of fire over a broad area. If there were fires within the area then they were single, isolated and of low intensity; however fires such as this can in fact encourage pine growth (O'Connell, 1990). The use of single fires to encourage the growth of Scots pine is currently being used in northern Scotland on *Calluna vulgaris* dominated moorland. In a five year period after a single burn, the establishment of pine from seed fall, seed survival and cumulative seedling recruitment has been shown to increase (Curlevski et al, 2010; Hancock et al, 2009). Therefore the fire regime on the South Erradale area is unlikely to have made any significant contribution to the pine decline. The increase in the frequency of charcoal, although still minor values, at Meall Bad á Chrotha can most probably be associated with an increase in the anthropogenic activity within the area, perhaps on the peninsula, through domestic fires or from the local burning of vegetation. Charcoal from such fires could well have been transported from elsewhere as the range of charcoal dispersal would have been enhanced due to the increasingly open landscape (Fossitt, 1996). Evidence collected from four locations throughout the highlands of Scotland have provided nothing to support the idea that fire had any influence over the dynamics of scots pine throughout the Holocene (Froyd, 2006). A reconstruction of *Pinus sylvestris* and charcoal from Lochan na h-Inghinn, Reidh-lochan, Loch an Amair and Dubh-Lochan show no correlation, with no evidence of a charcoal increase immediately before or during the pine rise and peaks (Froyd, 2006). High charcoal values have been linked to landscapes which are predominantly open blanket peat, particularly in the west of Scotland (Fossitt, 1994b). Management of peatlands may have been undertaken through the use of fire. This could account for the increase in charcoal concentrations after the mid-Holocene woodland decline, when blanket peat started to take hold of catchments across western Scotland.

There is no evidence to suggest that the pine decline witnessed at Meall Bad á Chrotha was a result of pathogenic attack. The decline of *Ulmus* through pathogenic attack at 5700 cal BP throughout northern Europe is clearly witnessed in every pollen diagram within the taxon range (Watts, 1961). If a similar attack was responsible for the collapse of *Pinus sylvestris* then it would be assumed that the decline would be clear in every palaeovegetational reconstruction within the taxon range, similarly to *Ulmus*. However, there are variations within the pollen diagrams from throughout Europe which supports the conclusion that pathogenic attack did not play a role in the pine decline (Bennett, 1994).

The evidence from Meall Bad á Chrotha and many other independent climate reconstructions indicate that it was reduced temperatures and subsequent changes in bog hydrology which played the controlling factor in the pine decline. This cooling trend is most likely associated with a shifting North Atlantic oscillation at 4200 cal BP and possible decreases in solar insolation (Jessen et al, 2005; Wanner et al, 20011; Olson et al, 2012); which ultimately led to pine becoming locally extinct between 4200 and 3500 cal BP (Gear, 1989). This is not to say however that other factors, such as storminess, volcanic or human activity did not influence vegetation dynamics during the mid-Holocene; although, if they did, then their influence at Meall Bad á Chrotha was minimal.

6.3 Mid-Late Holocene: Return to Open Landscape and Intensification of Anthropogenic Activity

The collapse of woodland, particularly *Pinus sylvestris*, throughout Scotland from around 4000 cal BP allowed the growing human population to expand into areas that were before unsuitable for land use. Following the pine decline at Meall Bad á Chrotha and the overall decline in the level of arboreal pollen, there is a coincident appearance in *Plantago lanceolata* (160cm), albeit at low percentages, accompanied by an increase in Ericaceous dwarf shrubs, (presumably *Calluna vulgaris),* Poaceae and Cyperaceae. This new open landscape dominated by heath, sedge and grass communities would have provided an extensive area for subsistence farmers and hunter gatherers to exploit (Davies, 1999, 2007; Davies et al, 2004; Tipping and Tisdall, 2004). In the words of Baillie (1998) this period during the Holocene was *"Bad for trees – Good for humans"*. Similar vegetation successions can be seen throughout Scotland supporting the theory that there was an expansion in farming communities during the Bronze Age period (4500-2800 cal BP). At Glen Carron the appearance and increase in *Plantago lanceolata* is dated to 3830 cal BP, while at Glen Torridon a very comparable date of 3850 cal BP has been given for the same event (Anderson, 1996). However, Anderson (1998) suggests that these low percentages of *Plantago lanceolata* indicate that cultivation was not intensive immediately close to the sites; although, they still indicate that this period signified the point when burning, grazing and cultivation became permanent features in upland areas. At Meall Bad á Chrotha the appearance of *Plantago lanceolata* occurs just before pine falls to its lowest value suggesting that the response to this decline in woodland and new open landscape was relatively immediate on the South Erradale peninsula. At Reidchalmai the anthropogenic reaction to the declining woodland populations was also relatively immediate (Tipping and McCulloch, 2003); however around Loch Farlary in the uplands pastoral farmers seem to have taken approximately 800 cal years to exploit the landscape (Tipping et al, 2008). Although, this interval between forest decline and the establishment of pastoral farming around Loch Farlary is not a typical trend throughout upland areas; west Glen Affric had been exploited by farming communities since 4300 cal BP (Davies, 2007; Davies and Tipping, 2004).

The continued reduction in arboreal pollen and creation of openings during the later Holocene at Meall Bad à Chrotha could well be due to the sustained grazing of domesticated animals which would push any remaining woodland, which was already at its marginal limits of growth, past its survival threshold. The low percentages of trees remaining are likely to be present in sheltered, isolated pockets within the catchment or Badachro river valley. There seems to be no relationship between the fire frequencies and heath abundances which support the suggestion that the charcoal recorded originated from a different source area. Clark (1988) suggests that small charcoal frequencies could well be from a different part of the region as microscopic charcoal has the ability to travel over long distances. The increase in agricultural indicators supports the idea that these fires were of human origin as opposed to natural; fire may have been used to clear land for grazing animals and cultivation. Throughout Scotland the falling levels of arboreal pollen and heathland along with the increase in fire frequency and intensity support this theory; however, there is very limited evidence from Meall Bad á Chrotha to make this suggestion reliable. If the charcoal recorded is of human origin then the fires were not in the local area and were on a very small scale. At Loch Farlary, there was intensification in grazing pressures after 3500 cal BP, with increases in *Pteridium,* Poaceae and *Plantago lanceolata* (Tipping et al, 2008). After 3040 cal BP there is also evidence for more regular burning. Regular, controlled fires were used as a way of improving the quality of *Calluna* heath for the grazing of domestic animals, which also probably prevented the regeneration of tree species. These grazing pressures witnessed at Loch Farlary could well have contributed to the reduction of trees; by 3000 cal BP the area was almost completely treeless (Tipping et al, 2008).

The sharp peak in *Sphagnum* at the very end of the pine decline at 150cm could suggest rapid paludification on a local level. This coincides with relatively abundant levels of Ericaceae (most likely *Calluna vulgaris*) and an increasing contribution from Cyperaceae and an overall shift to a more water-tolerant community which indicate a transition to a wetter phase with enhanced peat-forming conditions (Ellis and Tallis, 2000). However, a rise in the percentage of *Sphagnum* and a decline in the level of pine pollen can indicate a peak in acidity (Charman, 1990). The interaction between *Pinus sylvestris* and *Sphagnum* has been long debated but it is possible that this *Sphagnum* peak is associated with a very local rise in the level of bog acidification at Meall Bad á Chrotha. If pine is already at its marginal limits of growth, then even a slight short term increase in the level of acidity can push any already vulnerable pine woodlands past its equilibrium threshold and prevent any regeneration of the species (Blackford et al, 1992). However, the central pine decline had already taken place at Meall Bad á Chrotha before this *Sphagnum* peak. Also, it could have only been growing immediately on the coring site which would not have had any impact on pine growth. *Sphagnum* levels remain very low throughout the pine phase, and then peak once pine falls to insignificant values. Ohlson et al (2001) have suggested that the growth of pine can prevent the colonisation of *Sphagnum*; therefore its decline may have enabled *Sphagnum* to recolonize and re-establish itself on the mire surface.

There is a 2°C fall, down to just below 10°C, in the MTWA accompanying the rise in water-tolerant taxa, such as *Sphagnum* and Cyperaceae, which suggests that this vegetation change and shift to a wetter environment could be associated with the climatic event recorded in the North Atlantic at 2800 cal BP, which resulted in local effective precipitation increases (O'Brien et al, 1995; Anderson, 1995, 1998; Van Geel et al, 1996, 1998a, b; Binney, 1997; Bond et al, 1997; Anderson et al, 1998; Hughes et al, 2000; Tisdall, 2000, 2003; Charman et al, 2001, 2006; Langdon and Barber, 2002; Langdon et al, 2003). An ice rafted debris event, similar to the one recorded around 4200 cal BP, is said to have caused a fairly extreme period of climatic deterioration with substantial temperature and rainfall variations (Bond et al, 1997). If the temperatures fell enough then evapotranspiration rates could well have been reduced, therefore the water table would have risen, as opposed to an actual rise in rainfall. Although, increased rainfall could have still contributed to the increase in waterlogging. This is referred to as 'effective precipitation' (Tipping et al, 2008). Lamb (1977) argued that the 2800 cal BP climatic event saw temperatures fall as much as 2°C lower than they were during the climatic optimum which occurred between 8000 and 5000 cal BP. There is evidence for a period of increased wetness in a number of records from throughout Scotland. At Loch Farlary, a transition to increasingly wet mire surfaces is recorded from 2900 cal BP (Tipping et al, 2008), while at Beinn Dearg a date of 3500 and 2800 cal BP has been put forward as a period of increased wetness. Furthermore, Anderson (1998) suggested that two wet shifts occurred in the Northwest of Scotland between 3340-3060 and 2850-2540 cal BP. And an extended phase of greater precipitation levels is recorded in west Glen Affric at Loch Coulavie between 2700 cal BP to around 2300-2000 cal BP (Tisdall, 2000, 2003).

This climatic transition has been suggested to have had a substantial impact on the stability of late Bronze Age settlements (Piggott, 1972; Burgess, 1985, 1989; Barber, 1998; Cowley, 1998; McCullagh and Tipping, 1998; Tipping, 2002). Populations in the Northwest and throughout northern Scotland would have been particularly vulnerable to any climatic deterioration due to the close proximity of the North Atlantic Ocean (Bond et al, 1997; Tipping and Tisdall, 2004). The overall assumption is that these communities were forced to abandon these marginal areas, particularly upland and coastal areas, to more sheltered areas in the lowlands, perhaps even to different regions less affected by this climatic downturn. This assumption is supported by the apparent termination or lessening of crop production at Loch Farlary just after 2700 cal BP. Similar events involving crop production come from Achany Glen at 2900 cal BP (Smith, 1998) and west Glen Affric, which is a very exposed upland area, around 2860 cal BP (Tipping, 2002; Davies, 2003; Davies et al, 2004). Although, there is evidence to suggest a reduction in cultivation at these sites, the assumption that these areas were completely abandoned due to this climatic downturn may be over exaggerated; indeed grazing continued. There is no evidence to suggest that abandonment or a reduction in the production of agricultural crops occurred on the South Erradale peninsula. *Plantago lanceolata* is a continuous presence within the record from 130cm, albeit at very low values, indicating the possibility of a continued small scale anthropogenic presence throughout the later part of the Holocene in the area. However, basing the evidence for human activity on only one taxon is rather unreliable, as *Plantago lanceolata* can occur naturally within a landscape. Although, charcoal also continues to be present within the record at low values, but, yet again, fires can occur naturally. If these occurrences were related to anthropogenic disturbance then it is most likely that these low levels of *Plantago lanceolata* and charcoal originate from the more versatile flat land in the valley below.

Since the end of the mid-Holocene thermal maximum an overall cooling trend in summer temperatures has continued throughout the late Holocene. The decline continues to fall below early Holocene temperatures, fluctuating between 9.5-11^{o}C in comparison to the 11.5-12.5^{o}C seen during the start of the record. In northern Finland, the temperature drops below 12^{o}C at 1900 cal BP and then for the first time since the early Holocene drops below 11.5^{o}C at 800 cal BP (Seppa and Birks, 2001). There is little evidence for any major climatic excursions after what is assumed to be the 2800 cal BP climatic event shown in the Meall Bad á Chrotha record. There is a lack of evidence for the 'climatic optimum' which is recorded in many pollen diagrams throughout the northern hemisphere. Dansgaard et al (1989) used data collected from a Greenland ice core to argue that this period was characterised by temperatures 1 - 2°C greater than they are today, with substantial retreat of the glaciers in the northern hemisphere (Grove, 1979). This event occurred around 1550 and 1100 cal BP at Glen Carron, Glen Torridon and Eilean Subhainn characterized by abundant *Calluna vulgaris* representing the warmer and drier conditions (Anderson, 1998). There is a slight rise (70 – 80cm) in Ericaceous dwarf shrubs following a decline in *Corylus/Myrica* (presumed to be *Myrica* due to the overall decline in arboreal pollen and the relatively wet conditions that prevailed at the time) and small overall decline in Cyperaceae which could suggest a small reduction in the level of bog surface wetness at Meall Bad á Chrotha. However, this change is only slight making it difficult to attribute this variation to a significant climate event rather than a small, short-lived local fluctuation in bog surface wetness.

A short-lived period of wetter bog conditions can be inferred from a sharp peak in the level of Cyperaceae at 30cm at Meall Bad á Chrotha. This correlates with a 1^{o}C drop in summer temperature, which in turn would lower evaporation rates and increase bog surface wetness - this could be attributed to the 'Little Ice Age' which can also be seen at other sites within Wester Ross, in particular, Glen Carron and Glen Torridon (Anderson, 1996, 1998). Anderson (1998) dates the wet shift associated with this climatic deterioration between 940 and 760 cal BP which is fairly early in comparison to other dates given for the event. However, similarly to the Little Optimum it is difficult to say whether this slight variation in the vegetation of Meall Bad á Chrotha was due to a more regional climatic event such as the Little Ice Age or a more local hydrological variation. At Glen Torridon this climatic shift coincided with the disappearance of *Plantago lanceolata* and a reduction of microscopic charcoal which suggests that land use and anthropogenic activity at this time ceased in the area most likely as a result of the worsening climatic conditions (Anderson, 1998). There is no evidence of any abandonment at this time in the Meall Bad á Chrotha record. In fact, at this point, *Plantago lanceolata* begins to increase, although still at low values, suggesting a slight intensification of agriculture somewhere within the catchment, or perhaps closer to the site. However, the evidence is still rather limited based on only one taxon, therefore the possibility of natural occurrence cannot be ruled out.

Throughout western Scotland and the upland areas there is extensive evidence of prehistoric activity in localities that are overlaid with blanket peat and heath today. There remains much debate as to what encouraged the establishment and growth of peat in these areas. There has been the suggestion that the intensification of human activity after the woodland decline led to its development. The use of a plough on the land can cause the soil impaction which in turn reduces soil drainage. Also, the reduction of cultivation can lower the availability of nutrients needed for many plants to grow. This would allow plants that are adapted to poor, wetter soil conditions such as *Sphagnum* to colonise the area and therefore initiate the development of peat through a deficiency in aeration and in turn a reduction in the decomposition of plants (Edwards and Whittington, 1997). However, it is more likely that the development of peat at Meall Bad á Chrotha was due to natural causes such as increasingly wet climatic conditions resulting from greater oceanic influences (Edwards and Whittington, 1997). The upland nature of Meall Bad á Chrotha and its inaccessibility supports the assumption that the vegetational changes and the expansion of peat were unlikely to be related to significant human impact. The low *Plantago lanceolata* and charcoal percentages further support this assumption indicating that there was no intense land use going on in close proximity to the site. However, the likelihood that the vast blanket peats present today on the lower areas of the South Erradale peninsula are related to human impact is much greater due to its suitability for settlement and land use; and there is continuous evidence for anthropogenic activity in the vicinity since the Mesolithic period and there is a permanent population inhabiting the area today.

Chapter 7: Conclusion

The record from Meall Bad à Chrotha, Wester Ross, Northwest Scotland, has allowed for the reconstruction of past vegetation, climate and anthropogenic activity for a time period spanning from the Lateglacial to present day.

The Lateglacial to Holocene transition can be seen by the lacustrine environment and pioneer communities present at the base of the record; these pioneer communities were first to colonise the bare nutrient-poor ground around the lake after deglaciation. The steady rise in summer temperature along with the terrestrialisation of the lake and improvement of conditions allowed for the expansion of more nutrient-demanding vegetation, such as *Betula* and slightly later *Corylus*, forming the mixed deciduous woodland that characterises much of Scotland during the early Holocene. The continued drier bog surfaces and warmer climatic conditions associated with the Holocene Thermal Maximum eventually lead to the colonisation and establishment of *Pinus sylvestris* and the development of a dense pine forest by 7000 cal BP at Meall Bad à Chrotha.

The more regular occurrence of 'background' charcoal before the pine rise could well represent the burning of single fires stimulated by the increasingly dry bog conditions brought on by the rising temperatures. There is the possibility that these fires could be due to small scale, domestic burning by early settlers somewhere within the area; however, the lack of evidence for any further human activity supports the assumption that spontaneous wildfires were more likely the cause of the charcoal rise. It is possible that these isolated single fires did encourage the growth of *Pinus sylvestris* somewhere within the catchment, although, the primary contributing factor in the colonisation and expansion of pine would most likely have been the establishment of a suitable ecological niche for pine to grow in the warmer and drier conditions.

The lack of evidence for any Mesolithic activity in the local vicinity at Meall Bad à Chrotha is not representative of activity elsewhere on the peninsula - there is evidence at Sand of a Mesolithic midden and numerous other sites within the catchment of small populations which relied heavily on the coastal resources the South Erradale and the neighbouring Applecross peninsula had to offer (Hardy and Wickham-Jones, 2003). The high, exposed, inaccessible nature of Meall Bad à Chrotha is far from ideal for the settlement of early populations, who would likely have preferred lower, more sheltered land such that present in the valley below.

The end of the Holocene Thermal Maximum and the start of the Holocene Thermal Decline can be seen across the Northern Hemisphere in forest range and composition. The pine decline is a clear, central aspect in the vegetational response to climatic deterioration at Meall Bad à Chrotha. Unlike Glen Carron and Eilean Subhainn (Anderson, 1998), there is no evidence of a two phase decline which could well suggest that the two phase declines witnessed at these sites were in fact due to local rather than regional factors.

The vegetation succession and the temperature reconstruction at Meall Bad á Chrotha as well as the relative synchronicity of the decline and correlation of climatic shift indicators between studies, from not only different catchments but different regions, support the proposal that it was a regional climatic deterioration that resulted in the collapse of pine woodlands in the studied region, but also across much of northern Scotland. The sensitive relationship between pine growth and any climatic fluctuations is evident in the MTWA reconstruction with pine decreasing at the same time as a distinct cooling trend; however, the issue of pine essentially driving the MTWA plot and circular reasoning has to be remembered. Although, when compared to outside studies the results remain similar – the pine collapse coincides with a temperature drop. A slight change to wetter conditions particularly in areas where growing conditions are already marginal can cause the cessation of germination and pine growth. The Holocene Thermal Decline sees summer temperatures fall below early Holocene values, making the last four millennia the coldest of the Holocene.

The close proximity of the Northwest of Scotland to the North Atlantic Ocean is the dominant controlling factor on the climate of the region and is most likely responsible for the climatic fluctuations witnessed throughout the Holocene at

Meall Bad á Chrotha. The timing of ice rafted debris events and sea surface temperature changes coincide with many hydrological variations throughout Wester Ross, indicating that circulation changes in the North Atlantic Ocean and impacts on regional climate could well be the dominant forcing mechanism responsible for the pine decline. In particular an event ice rafted debris event in the North Atlantic (Bond event no. 3) at 4200 cal BP coincides with the climatic shift that signalled the fate of *Pinus sylvestris* (Bond et al, 1997).

Other factors which could have influenced the pine decline, such as human activity, volcanic eruptions or pathogenic attack have been disregarded here due to the lack of evidence to support these hypotheses; although, that is not to say they did not contribute to more local declines in other areas of Wester Ross and Northern Scotland. If they did contribute to the pine collapse at Meall Bad á Chrotha then their influence would have been minimal as the evidence is stronger for other factors playing the driving role.

There is little evidence at Meall Bad à Chrotha to suggest that human interference changed the structure and composition of vegetation at any point during the Holocene. However, it is highly likely that the surrounding South Erradale peninsula became increasingly subject to anthropogenic activity as the Holocene progressed towards present day, with communities exploiting the newly open landscape. The appearance of *Plantago lanceolata* and the low levels of charcoal after the pine decline are again most likely to be 'background' levels due to Neolithic, Bronze Age and later land-use activity, such as crop production, domesticated animal grazing and small-scale domestic fires in the valley below. The continued reduction of arboreal pollen in the later Holocene could be due to the clearance of the remaining pockets of woodland after the mid-Holocene woodland decline by the grazing of domestic livestock. This transition to a more open landscape after the decline of *Pinus sylvestris* and other tree species gave way for the establishment of peat-forming communities and the treeless environment we see today.

This study supports Bigler et al's (2002) proposal that the last four millennia have been the coldest of the Holocene. However, this pattern is beginning to change as anthropogenically-driven climate change continues to threaten the ecosystems and communities that inhabit our planet causing “severe, persuasive and irreversible” damage (IPCC, 2014). It is vital for studies such as the one carried out at Meall Bad á Chrotha to investigate past climatic fluctuations and to expand our knowledge on the subsequent terrestrial responses. This in turn would allow us to envisage and prepare for any future climatic and environmental effects.

Appendix

Depth	Sphagnum	Ericaceae	Cyperaceae	Pinus	Betula	Corylus	Alnus	Ulmus	Poaceae	Plantago	Polypodiaceae	Artemisia	Asteraceae	Fagus	Quercus	Salix	Carpinus	Tilia	Pteriduim	Total Pollen
10cm	189	108	217	4	4	2	0	0	36	7	1	1	1	2	0	0	0	0	0	383
20cm	56	94	123	2	19	26	3	0	39	12	11	2	7	0	4	1	2	0	0	345
30cm	228	113	618	12	50	27	5	1	40	11	7	0	0	0	5	0	1	1	0	891
40cm	73	101	185	4	22	10	1	1	55	1	2	4	1	0	0	1	1	0	1	390
50cm	92	162	193	17	30	17	2	6	43	2	6	5	0	0	1	0	1	0	1	486
60cm	156	248	486	13	63	52	2	5	108	9	13	0	0	1	4	0	0	0	0	1004
70cm	54	260	166	4	38	19	7	3	52	4	9	4	0	1	2	0	0	0	2	571
80cm	69	448	334	7	100	68	25	0	118	6	5	6	0	1	6	0	0	0	0	1124
90cm	74	375	440	13	135	54	30	0	146	4	7	4	0	0	11	0	2	0	5	1226
100cm	31	299	327	14	114	260	25	0	61	1	1	4	0	1	1	0	0	2	6	1116
110cm	34	315	448	11	124	107	17	0	100	3	6	11	0	3	3	0	0	1	0	1149
120cm	90	143	64	6	60	583	25	0	61	3	2	5	0	4	6	0	1	0	4	967
130cm	25	151	198	11	27	48	10	0	41	1	2	3	0	3	2	0	1	0	0	498
140cm	12	102	253	4	22	22	12	0	14	0	1	3	0	1	2	0	0	0	4	440
150cm	773	82	105	3	35	24	38	1	34	1	0	6	0	0	1	0	0	0	9	339
160cm	52	110	175	24	86	34	34	2	101	1	13	3	0	5	3	1	0	0	0	592
170cm	46	375	12	107	77	63	51	1	148	0	3	15	0	0	5	2	1	0	1	861
180cm	32	336	55	17	53	15	28	0	112	0	4	3	0	2	4	1	0	0	0	630
190cm	81	242	102	236	86	55	42	0	78	0	5	4	0	5	24	3	0	0	33	915
200cm	99	652	37	660	115	58	58	5	55	0	4	4	0	0	12	1	0	0	25	1686
210cm	2	121	0	839	35	14	16	0	11	0	0	0	0	0	2	0	0	0	1	1039
220cm	28	449	2	812	96	42	31	0	9	0	3	3	0	0	3	0	0	0	0	1450
230cm	148	357	7	360	115	77	72	5	34	0	0	3	0	0	2	0	0	0	5	1037
240cm	159	214	7	467	189	108	59	4	24	0	0	2	0	0	4	1	0	0	5	1084
250cm	6	146	37	131	79	42	30	0	18	0	0	1	0	0	1	1	0	0	0	486
260cm	15	379	4	324	83	70	61	0	30	0	4	1	0	0	1	0	0	0	39	996
270cm	33	522	0	54	51	82	22	0	6	0	3	0	0	0	1	0	0	0	8	749
280cm	3	248	1	263	64	36	42	0	19	0	0	0	0	0	1	0	0	0	2	676
290cm	1	389	0	126	21	10	3	1	1	0	0	0	0	0	0	0	0	0	0	551
300cm	4	297	0	26	17	6	2	1	0	0	0	0	0	0	2	0	0	0	0	351
310cm	6	467	0	27	38	37	4	2	3	0	0	0	0	0	0	0	0	0	3	581
320cm	1	420	1	55	52	31	3	2	0	0	0	0	0	0	0	0	0	0	0	564
330cm	6	642	0	39	78	68	10	0	175	0	1	29	0	0	1	3	0	14	4	1064
340cm	75	631	0	29	364	82	10	1	119	0	8	13	0	0	1	1	0	2	7	1268
350cm	12	73	0	6	1030	594	3	1	6	0	4	4	0	0	0	2	0	0	10	1733
360cm	0	414	0	4	213	230	1	2	2	0	3	2	0	0	1	2	2	0	3	879
370cm	6	195	3	2	144	92	1	2	3	0	9	0	0	0	0	0	0	0	3	454
380cm	34	122	2	12	86	49	1	0	28	0	46	1	0	1	1	2	0	0	3	354
390cm	396	208	3	7	255	56	1	5	7	0	23	2	0	1	6	6	0	0	9	589
400cm	300	175	20	17	63	18	1	7	13	0	34	0	0	0	3	3	0	0	15	369
410cm	270	235	2	7	63	35	0	1	20	0	29	0	0	0	1	1	0	3	2	399
420cm	56	203	2	3	71	31	0	0	5	1	9	3	0	0	1	1	0	0	0	330
430cm	353	298	17	6	130	26	0	1	14	0	20	1	0	0	0	3	0	0	1	517
440cm	767	254	12	2	249	16	0	0	6	0	15	0	2	0	0	3	0	0	3	562
450cm	694	173	24	9	120	26	0	0	9	0	33	0	0	0	0	6	0	0	1	401
460cm	124	62	112	10	269	41	5	1	89	0	25	11	0	0	1	3	0	0	0	629
470cm	335	87	150	21	110	34	8	0	92	0	17	8	3	1	3	13	0	0	0	547
480cm	30	100	6	10	101	34	2	0	43	0	98	5	2	0	0	14	0	2	0	417
490cm	9	207	25	17	34	15	0	3	89	0	134	4	1	0	0	15	0	0	0	544
500cm	3	30	6	12	14	0	0	0	18	0	26	8	0	4	0	8	0	0	0	126

Table 17. Pollen counts from Meall Bad á Chrotha

Depth	Charcoal Count
10cm	9
20cm	3
30cm	6
40cm	12
50cm	7
60cm	8
70cm	5
80cm	8
90cm	14
100cm	7
110cm	8
120cm	3
130cm	6
140cm	4
150cm	5
160cm	15
170cm	8
180cm	4
190cm	6
200cm	4
210cm	1
220cm	1
230cm	0
240cm	3
250cm	0
260cm	21
270cm	15
280cm	13
290cm	3
300cm	1
310cm	0
320cm	3
330cm	11
340cm	25
350cm	2
360cm	1
370cm	0
380cm	0
390cm	1
400cm	0
410cm	0
420cm	0
430cm	2
440cm	0
450cm	0
460cm	0
470cm	0
480cm	0
490cm	1
500cm	0

Table 18. Charcoal counts from Meall Bad á Chrotha.

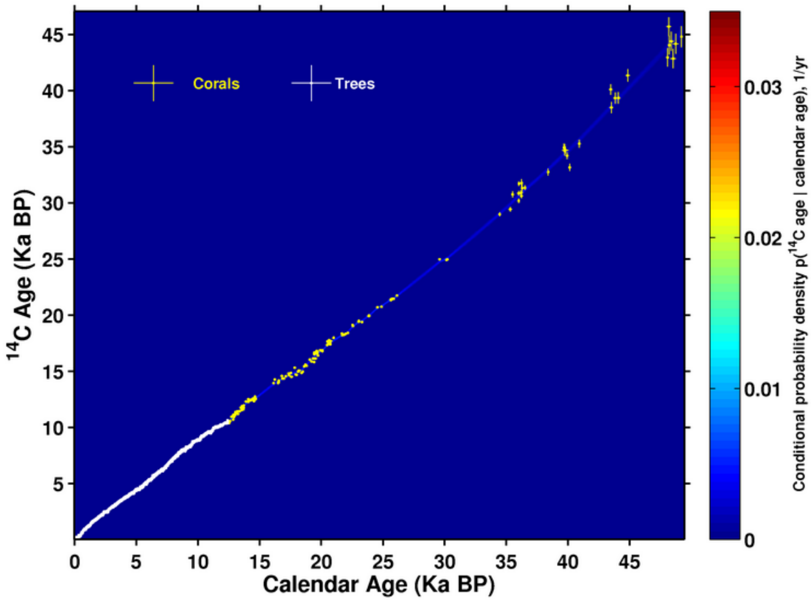


Figure 40. The Fairbank0107 calibration curve was used to convert radiocarbon dates into calendar or calibrated years BP (Fairbanks et al, 2005).

References

1. Aalbersberg.G., Litt.T., 1998, Multi-Proxy Climatic Reconstructions for the Eemian and Early Weichselian, *Journal of Quaternary Science*, Vol.13., pp.367-390.
2. Agee.J.K., 1998, Fire and Pine Ecosystems, In: *Ecology and Biogeography of Pinus* (eds) Richardson.D.M., Cambridge University Press, pp.193-208.
3. Ali.M., Oda.H., Hayashida.A., Takemura.K., Torii.M., 1999, Holocene Palaeomagnetic Secular Variation at Lake Biwa, Central Japan, *Geophysical Journal International*, Vol.136., pp.218-228.
4. Allen.P., 2007, Pollen Analysis from Lanton Quarry, Northumberland*, Archaeological Research Services Ltd.*
5. Alley.R.B., 2000, The Younger Dryas Cold Interval as Viewed from Central Greenland, *Quaternary Science Review,* Vol.19., pp.213-226.
6. Alley.R.B., Meese.D.A., Shuman.C.A., Gow.A.J., Taylor.K.C., Grootes.P.M., White.J.W.C., Ram.M., Waddington.E.D., Mayewski.P.A., Zielinski.G.A., 1993, Abrupt Increase in Greenland Snow Accumulation at the End of the Younger Dryas Event, *Nature,* Vol.361., pp.527-529.
7. Alverson.K.D., Bradley.R.S., Pedersen.T.F., 2003, *Paleoclimate, Global Change and the Future,* Springer, New York.
8. Andersen.S.H., 1998, A Survey of the Late Palaeolithic of Denmark and Southern Sweden, *British Archaeological Reports*, pp.523-566.
9. Anderson.D.A., 1995, *An Abrupt mid-Holocene Decline of Pinus sylvestris in Glen Torridon, Northern Scotland: Implications for Palaeoclimatic Change,* Oxford, School of Geography Research Papers, Vol.52., vol.1-29.
10. Anderson.D.A., 1998, A Reconstruction of Holocene Climatic Changes from Peat Bogs in Northwest Scotland, *Boreas,* Vol.27., pp.208-224.
11. Anderson.D.E., 1996, *Abrupt Holocene Climatic Change Recorded in Terrestrial Peat Sequences from Wester Ross, Scotland,* D.Phil. Thesis, University of Oxford.
12. Anderson.D.E., 1997, Younger Dryas Research and its Implications for Understanding Abrupt Climate Change, *Progress in Physical Geography*, Vol.21., pp.230-249.
13. Anderson.D.E., 1998, A Reconstruction of Holocene Climatic Changes from Peat Bogs in Northwest Scotland, *Boreas,* Vol.27., pp.208-224.

14. Anderson.D.E., Binney.H.A., Smith.M.A., 1998, Evidence for Abrupt Climate Change in Northern Scotland between 3900 and 3500 Calendar years BP*, The Holocene*, Vol.8., pp.97-103.
15. Anderson.R., 2010, *Restoring Afforested Peat Bogs: Results of Current Research*, Forestry Commission Research Note.
16. Antonsson.K., 2006, Holocene Climate in Central and Southern Sweden, Quantitative Reconstructions from Fossil Data, *Acta Universitatis Upsaliensis Uppsala,* Vol.168., pp.9-37.
17. Armitt.I., Finlayson.B., 1992, Hunter-Gatherers Transformed: The Transition to Agriculture in Northern and Western Europe, *Antiquity*, Vol.66., pp.664-676.
18. Artz.R., Donnelly.D., Anderson.R., Mitchell.R., Chapman.S., Smith.J., Smith.R., Cummins.R., Balana.B., Cuthbert.A., 2012, *Managing and Restoring Blanket Big to Benefit Biodiversity and Carbon Balance – A Scoping Study,* Scottish Natural Heritage Commissioned Report.
19. Ashmore.P.J., 2001, A List of Archaeological Radiocarbon Dates, *Discovery and Excavation in Scotland*, Vol.2., pp.122-128.
20. Atkinson.T.C., Briffa.K.R., Coope.G.R., 1987, Seasonal Temperatures in Britain During the Past 20,000 years Reconstructed Using Beetle Remains, *Nature*, Vol.325., pp.587-592.
21. Atkinson.T.C., Briffa.K.R., Coope.G.R., 1987, Seasonal Temperatures in Britain During the Past 22,000 Years, Reconstructed Using Beetle Remains, *Nature*, Vol.325., pp.587-592.
22. Baillie.M.G.L., 1995, *A Slice Though Time,* Dendrochronology and Precision Dating, Batsford Ltd, London.
23. Baillie.M.G.L., 1998, Bad for Trees – Bad for Humans?, In: *Life on the Edge: Human Settlement and Marginality* (eds) Mills.C.M., Coles.G., Oxbow Monograph, Oxford, Vol.100., pp.13-19.
24. Ballantyne.C.K., and Harris.C., 1994, *The Periglaciation of Great Britain*, Cambridge University Press, Cambridge.
25. Ballantyne.C.K., Schnabel.C., Xu.S., 2009, Re-Advance of the Last British Ice Sheet during the Greenland Interstade (GI-1): The Wester Ross Re-Advance, Northwest Scotland, *Quaternary Science Reviews,* Vol.28., pp.783-789.
26. Ballie.M.G.L., 1989, Hekla 3: How Big Was it? *Endeavour*, Vol.13, pp.78-811.

27. Ballie.M.G.L., Munro.M.A.R., 1988, Irish Trees Rings: Santorini and Volcanic Dust Veils, *Nature,* Vol.332, pp.344-346.

28. Ballin.T.B., Saville.A., 2003, An Ahrensburgian-Type Tanged Point from Shieldaig, Wester Ross, Scotland, and its Implications, *Oxford Journal of Archaeology*, Vol.22., pp.115-131.

29. Bannister.P., 1964, The Water Relations of Certain Heath Plants in Relation to Their Ecological Amplitude, II, Field Studies, *Journal of Ecology*, Vol.52., pp.481-497.

30. Bar Matthews.M., Ayalon.A., Kaufman.A., 1997, Late Quaternary Paleoclimate in the Eastern Mediterranean Region from Stable Isotope Analysis of Speleothems at Soreq Cave, Israel, *Quaternary Research*, Vol.47., pp.155-168.

31. Barber.J., 1998, *The Archaeological Investigation of a Prehistoric Landscape: Excavations on Arran 1978-1981*, Edinburgh, Scottish Trust for Archaeological Research.

32. Barber.K., Chambers.F.M., Maddy.D., Stoneman.R., Brew.S., 1994, A Sensitive High Resolution Record of Late-Holocene Climatic Change from a Raised Bog in Northern England, *The Holocene*, Vol.4., pp.198-205.

33. Barber.K.E., Chambers.F.M., Maddy.D., 1994a, Sensitive High Resolution Records of Holocene Palaeoclimate from Ombrotrophic Bogs, In: *Palaeoclimate of the Last Glacial/Interglacial Cycle* Funnell, (eds) B.M. & Kay, R.L.F., pp.57-60., Special Publication no. 94/2, NERC Earth Science Directorate.

34. Barber.K.E., Chambers.F.M., Maddy.D., Stoneman.R., Brew.J.S., 1994b, A Sensitive High Resolution Record of Late Holocene Climate Change From a Raised Bog in Northern England, *The Holocene*, Vol.4., pp198-205.

35. Bard.E., Arnold.M., Maurice.P., Duprat.J., Moyes.J., Duplessy.J.C., 1987, Retreat Velocity of the North Atlantic Polar Front During the Last Deglaciation Determined by 14C Accelerator Mass Spectrometry, *Nature*, Vol.328., pp.791-794.

36. Bard.E., Raisbeck.G.M., Yiou.F., Jouzel.J., 1997, Solar Modulation of Cosmogenic Nuclide Production Over the Last Millennium: Comparison Betweeen ^{14}C and ^{10}Be Records, *Earth and Planetary Science Letters*, Vol.150., pp.453-462.

37. Barnekow.L., 2000, Holocene Regional and Local Vegetation History and Lake-Level Changes in the Tornetrask Area, Northern Sweden, *Journal of Paleolimnology*, Vol.23., pp.399-420.

38. Barton..N., 1997, *Stone Age Britain*, London, Batsford/English Heritage.

39. Barton.N., Roberts.A.J., Roe.D.A., 1991, *The Lateglacial in Northwest Europe: Human Adaptation and Environmental Change at the end of the Pleistocene,* CBA Research Report, No.77.
40. Beer.J., 2000, Long-Term Indirect Indices of Solar Variability, *Space Science Reviews*, Vol.94., pp.53-66.
41. Beer.J., Muscheler.R., Wagner.G., Laj.C., Kissel.C., Kubik.P.W., Synal.H.O., 2002, Cosmogenic Nucleides During Isotope Stages 2 and 3, *Quaternary Science Reviews*, Vol.21., pp.1129-1139.
42. Bell.J.N.B., Tallis.J.H., 1974, The Response of Empetrum nigrum to Different Mire Water Regimes with Special Reference to Wybunbury Moss, Cheshire and Featherbed Moss, Derbyshire, *Journal of Ecology,* Vol.62., pp.75-95.
43. Bell.M., and Walker.M.J.C., 2005, *Late Quaternary Environmental Change – Physical and Human Perspectives,* Pearson.
44. Benn.D.I., 1997, Glacier Fluctuations in Western Scotland, *Quaternary International*, Vol.38-39., pp.137-147.
45. Benn.D.I., Lowe.J.J., Walker.M.J.C., 1992, Glacier Response to Climatic Change During the Loch Lomond Stadial and Early Flandrian: Geomorphological and Palynological Evidence from the Isle of Skye, Scotland, *Journal of Quaternary Science*, Vol.7., pp.125-144.
46. Bennett.K.D., 1984, *The Post Glacial History of Pinus Sylvestris in the British Isles,* Quaternary Science Review, Vol.3., pp.133-155.
47. Bennett.K.D., 1986, The Rate of Spread and Population Increase of Forest Trees during the Postglacial, Philosophical Transactions of the Royal Society of London Series B, Vol.314., pp.523-531.
48. Bennett.K.D., 1989, A Provisional Map of Forest Types for the British Isles 5000 Years Ago, *Journal of Quaternary Science,* Vol.4., pp.141-144.
49. Bennett.K.D., 1995, Post-Glacial Dynamics of Pine (*Pinus sylvestris L.*) and Pinewoods in Scotland, In: *Our Pinewood Heritage* (eds) Aldhous.J.R. pp.23-39, Bell and Bain, Glasgow.
50. Bennett.K.D., 2005b, Map of distribution of Scottish pine remains. Public communication
51. Bennett.K.D., Boreham.S., Sharp.M.J., Switsur.V.R., 1992, Holocene History of Environment, Vegetation and Human Settlement on Cotta Ness, Lunnasting, Shetland, *Journal of Ecology*, Vol.80., pp.241-273.

52. Bennett.K.D., Boreham.S., Sharp.M.J., Switsur.V.R., 1992, Holocene History of Environment, Vegetation and Human Settlement on Catta Ness, Lunnasting, Shetland, Journal of Ecology, Vol.80., pp.241-273.

53. Bennett.K.D., Fossitt.J.A., Sharp.M., Switsur.V.R., 1990, Holocene Vegetational and Environmental History of Loch Lang, South Uist, Western Isles, Scotland, *New Phytologist*, Vol.114., pp.281-298.

54. Bennett.K.D., Tzedakis.P.C., Willis.K.J., 1991, Quaternary Refugia of North European Trees, *Journal of Biogeography,* Vol.18., pp.103-115.

55. Berglund.B.E., 2003, *Handbook of Holocene Palaeoecology and Palaeohydrology,* The Blackburn Press, Sweden.

56. Bever.J.D., Westover.K.M., Antonovics.J., 1997, Incorporating the Soil Community into Plant Population Dynamics: The Utility of the Feedback Approach, *Journal of Ecology,* Vol.85., pp.561-573.

57. Bigler.C., Larocque.I., Peglar.S.M., Birks.H.J.B., Hall.R.I., 2002, Quantitative Multiproxy Assessment of Long-Term Patterns of Holocene Environmental Change from a Small Lake Near Abisko, Northern Sweden, *The Holocene,* Vol.12., pp.481-496.

58. Binney.H.A., 1997, *Holocene Environmental Change in the Scottish Highlands: Multiproxy Evidence from Blanket Peats,* Unpublished Ph.D., thesis, London Guildhall University.

59. Birks.H. J. B., 1981, Late Wisconsin Vegetational and Climatic History at Kylen Lake, Northeastern Minnesota, *Quaternary Research*, Vol.16., pp.322–355.

60. Birks.H.H. and Mathewes.R.W., 1978, Studies in the Vegetational History of Scotland. V. Late Devensian and early Flandrian Pollen and Macrofossil Stratigraphy at Abernethy Forest, Inverness-shire, *New Phytologist*, Vol.80., pp.455–84.

61. Birks.H.H., 1970, Studies in the Vegetational History of Scotland: I. A Pollen Diagram from Abernathy Forest, Inverness-Shire, *Journal of Ecology*, Vol.58., No.3, pp.827-846.

62. Birks.H.H., 1972, Studies in the Vegetational History of Scotland. III. A radiocarbon-dated Pollen Diagram from Loch Maree, Ross and Cromarty, *New Phytologist*, Vol.71., pp.731-754.

63. Birks.H.H., 1975, Studies in the Vegetational History of Scotland. IV. Pine stumps in Scottish Blanket Peats, Philosophical Transactions of the Royal Society of London, *Biological Sciences*, Vol.270., pp.181-226.

64. Birks.H.H., 1984, Late Quaternary Pollen and Plant Macrofossil Stratigraphy at Lochan an Druim, Northwest Scotland, In: *Lake Sediments and Environment History* (eds) Haworth.E.Y., Lund.J.W.G., Leicester, Leicester University Press, pp.377-405.

65. Birks.H.J.B. and Birks.H.H., 1990, *Quaternary Palaeoecology,* Edward Arnold, London.

66. Birks.H.J.B., 1973, *Past and Present Vegetation of the Isle of Skye, A palaeoecological Study*, Cambridge University Press, London.

67. Birks.H.J.B., 1977, *The Flandrian Forest History of Scotland: A Preliminary Synthesis, British Quaternary Studies – Recent Advances* (eds) Shotton.F.W., pp.119-136, Clarendon Press, Oxford.

68. Birks.H.J.B., 1980, *Quaternary Vegetational History of West Scotland, 5th, International Palynological Conference , Guidebook for Excursion C8,* Cambridge, The Botany School University of Cambridge, pp.70.

69. Birks.H.J.B., 1987, *The Past and Present Vegetation of Oronsay and Colonsay and Adjacent Areas,* Mellars, pp.71-7.

70. Birks.H.J.B., 1988, *Long Term Ecological Change in the British Uplands, Ecological Change in the Uplands*, (eds), Usher.M.B., Thompson.D.B.A., pp.37-56, Blackwell, Oxford.

71. Birks.H.J.B., 1989, Holocene Isochrone Maps and Patterns of Tree-Spreading in the British Isles, *Journal of Biogeography*, Vol.18., pp.103-115.

72. Birks.H.J.B., 1990, Changes in Vegetation and Climate During the Holocene of Europe, In: Boer, M.M. & DeGroot, R.S. (eds): *Landscape – Ecological Impact of Climate Change*, pp.133-158., IOS Press, Amsterdam.

73. Birks.H.J.B., 1994, Did Icelandic Volcanic Eruptions Influence the Post-Glacial Vegetational History of the British Isles? *Trends in Ecology Evolution*, Vol.9., pp.312-314.

74. Birks.H.J.B., 2007, Plant Macrofossil Introduction, Encyclopaedia of Quaternary Paleoecology, *Elsevier Scientific Publishing*, pp.2266-2288

75. Birks.H.J.B., Williams.W., 1983, Late-Quaternary Vegetational History of the Inner Hebrides, *Proceedings of the Royal Society of Edinburgh*, Vol.83B., pp.269-292.

76. Bishop.W.W., Coope.G.R., 1977, Stratigraphical and Faunal Evidence for Lateglacial and Early Flandrian Environments in South-West Scotland, In: J.M.Gray and Lowe.J.J., *Studies in the Scottish Lateglacial Environment*, Oxford: Pergamon Press, pp.61-88.

77. Björck.S., Muscheler.R., Kromer.B., Andresen.C.S., Heinemeier.J., Johnsen.S.J., Conley.D., Koc.N., Spurk.M., Veski.S., 2001, High Resolution Analyses of an Early Holocene Climate Event May Imply Decreased Solar Forcing as an Important Climate Trigger, Geology, Vol.29, pp.1107-1110.
78. Bjune.A.E., Birks.H.J.B., Seppa.H., 2004, Holocene Vegetation and Climate History on a Continental-Oceanic Transect in Northern Fennoscandia Based on Pollen and Plant Macrofossils, *Boreas*, Vol.33., pp.211-223.
79. Blaauw.M., van Geel.Bas., van der Plicht.J., 2004, Solar Forcing of Climate Change during the Mid-Holocene: Indications from Raised Bogs in The Netherlands, *The Holocene*, Vol.14., pp.35-44.
80. Blackford.J.J. and Chambers.F.M., 1995., Proxy Climate Record for the Last 1000 Years from Irish Blanket Peat and a Possible Link to Solar Variability, *Earth and Planetary Science Letters*, Vol.133., pp.145-150.
81. Blackford.J.J., 2000, Palaeoclimatic Records from Peat Bogs, *Trends in Ecology and Evolution,* Vol.15., pp.193-198.
82. Blackford.J.J., Edwards.K.J., Dugmore.A.J., Cook.G.T., Buckland.P.C., 1992, Icelandic Volcanic Ash and the Mid-Holocene Scots Pine (*Pinus sylvestris*) Pollen Decline in Northern Scotland, *The Holocene*, Vol.2., pp.260-265.
83. Blundell.A. and Barber.K., 2005, A 2800-year Palaeoclimatic Record from Tore Hill Moss, Strathspey, Scotland: The Need for a Multi-Proxy Approach to Peat-Based Climatic Reconstructions, *Quaternary Science Reviews.* Vol.24., pp.1261-1277.
84. Bocquet-Appel.J.P., Naji.S., Vander Linden.M., Kozlowski.J.K., 2012, Understanding the Rates of Expansion of the Farming System in Europe, *Journal of Archaeological Science*, Vol.39., pp.531-546.
85. Boggie.R., 1972, Effect on Water-Table Height on Root Development of Pinus cortorta on Deep Peat in Scotland, *Oikos*, Vol.23., pp.304-312.
86. Bond.G., Kromer.B., Beer.J., Muscheler.R., Evans.M.N., Showers.W., Hoffman.S., Lotti-Bond.R., Hajdas.I., Bonani.G., 2001, Persistent Solar Influence on North Atlantic Climate During the Holocene, *Quaternary Science Reviews*, Vol.21., pp.1385-1394.
87. Bond.G., Showers.W., Cheseby.M., Lotti.R., Almasi.P., deMenocal.P., Priore.P., Cullen.H., Hajdas.I., Bonani.G., 1997, A Pervasive Millenial-Scale Cycle in the North Atlantic Holocene and Glacial Climates, *Science*, Vol.278., pp.1257-1266.

88. Bonsall.C., Anderson.D.E., Macklin.M.G., 2002, The Mesolithic-Neolithic Transition in Western Scotland and its European Context, *Documenta Praehistorica,* Vol.29., pp.214-249.
89. Bonsall.C., Smith.C., 1990, Bone and Antler Technology in the British Late Upper Palaeolithic and Mesolithic: The Impact of Accelerator Dating, In: Vermeersch.P., Van Peer.P., (eds) *Contributions to the Mesolithic in Europe*, pp.359-368.
90. Bonsall.S., Tolan-Smith.C., Saville.A., 1995, Direct Dating of Mesolithic Antler and Boone Artefacts from Great Britain: New Results for Bevelled Tools and Red Deer Antler Mattocks, *Mesolithic Miscellany,* Vol.16., pp.2-10.
91. Boomer.I., Von Grafenstein.U., Moss.A., 2012, Lateglacial to Early Holocene Multiproxy Record from Loch Assynt, Northwest Scotland, *Proceedings of the Geologists Association*, Vol.123., pp.109-116.
92. Bradley.D.E, 1957, *The Study of Pollen Grain Surfaces in the Electron Microscope,* Research Laboratory, Associated Electrical Industries LtD, Aldermarston, Berks, pp.226-229.
93. Bradley.R., 1984, *The Social Foundations of Prehistoric Britain: Themes and Variations in the Archaeology of Power,* London, Longman.
94. Bradley.R.S., 1988, The Explosive Volcanic Eruption Signal in Northern Hemisphere Continental Temperature Records, *Climate Change*, Vol.12, pp.221-243.
95. Bradley.R.S., 2003, Climate Forcing During the Holocene, In: *Global Change in the Holocene,* Mackay.A.W., Battarbee.R., Birks.J., Oldfield.F., (eds), Arnold, London, pp.10-19.
96. Bradley.R.S., Jones.P.D., 1993, 'Little Ice Age' Summer Temperature Variations: Their Nature and Relevance to Recent Global Warming Trends, *Holocene*, Vol.3., pp.367-376.
97. Bradshaw.R.H.W., 1993, Forest Response to Holocene Climatic Change: Equilibrium or Non-Equilibrium, In: *Climate Change and Human Impact on the Landcape: Studies in Palaeoecology and Environmental Archaeology* (eds) Chambers.F.M., Chapman and Hall, pp.57-65.
98. Bradshaw.R.H.W., Tolonen.K., Tolonen.M., 1996, Holocen Records of Fire from the Boreal and Temperate Zones of Europe, In: 1996, *Sediment Records of Biomass Burning and Global Change* (eds) Clark.J.S., et al Springer, Berlin, pp.347-365.

99. Bradwell. T., Fabel.D., Stoker.M.S., Mathers.H., McHargue.L., Howe.J.A., 2008, Ice Caps Existed Throughout the Lateglacial Interstadial in Northern Scotland, *Journal of Quaternary Science*, Vol.23., pp.401-407.

100. Brenchley.P., 2006, *the Geology of England and Wales*, Geological Society of London.

101. Bridge.M.C., Haggart.B.A., Lowe.J.J., 1990, The History and Significance of Sub fossil Remains of Pinus Sylvestris in Blanket Peats from Scotland, *Journal of Ecology,* Vol.78., no.1, pp.77-99.

102. Briffa.K.R., Jones.P.D., Pilcher.J.R., Hughes.M.K., 1988, Reconstructing Summer Temperatures in Northern Fennoscandia back to A.D.1700 using Tree-Ring Data from Scots-Pine, *Arctic and Alpine Research,* Vol.20., pp.385-394.

103. Briffa.K.R., Melvin.T.M., 2008, A Closer Look at Regional Chronology Standardisation of Tree-Ring Records: Justification of the Need, a Warning of Some Pitfalls and Suggested Improvements, in: *Its Application, Dendroclimatology: Progress and Prospects, Developments in Paleoenvironmental Research,* (eds) M.K.Hughes., H.F.Diaz., T.W.Swetnam, New York: Springer.

104. Broecker.W.S., 2006, Was the Younger Dryas Triggered by a Flood?, *Science*, Vol.312., pp.1146-1148.

105. Brooks.S.J., Birks.H.J.B., 2001, Chironomid-Inferred Air Temperatures from Lateglacial and Holocene Sites in Northwest Europe: Progress and Problems, *Quaternary Science Reviews*, Vol.20., pp.1723-1741.

106. Brown.A.G., 1997, Clearances and Clearings: Deforestation in Mesolithic/Neolithic Britain, Oxford *Journal of Archaology,* Vol.16., pp.133-146.

107. Bruneau.P.M.C., Johnson.S.M., 2014, Scotland's Peatland – Definitions and Information Resources, *Scottish Natural Heritage Commissioned Report*, No.701.

108. Bunting.M.J., Tipping.R., 2004, Complex Hydroseral Vegetation Succession and 'Dryland' Pollen Signals: A Case Study from Northwest Scotland, *The Holocene*, Vol.14., pp.53-63.

109. Burgess.C., 1985, Population, Climate and Upland Settlement, In: *Upland Settlement in Britain: The Second Millenium BC and After*, (eds) Stratt.D., Burgess.C., British Archaeological Reports, British Series, Vol.143., pp.195-230.

110. Burgess.C., 1989, Volcaneos, Catastrophe and the Global Crisis of the Late Secvond Millennium BC, *Current Archaeology,* Vol.117., pp.325-329.

111. Caput.C., Belot.Y., Auclair.D., Decourt.N., 1978, Absorption of Sulphur Dioxide by Pine Needles Leading to Acute Injury, *Environmental Pollution*, Vol.16., pp.3-15.

112. Carlisle.A., Brown.A.H.F., 1968, Biological Flora of the British Isles: *Pinus sylvestris* L. *Journal of Ecology,* Vol.56., pp.269-307.

113. Case.H.J., 1969, Neolithic Explanations, *Antiquity*, Vol.43., pp.176-186.

114. Caseldine.C., Hatton.J., Huber.U., Chiverrell.R., Woolley.N., 1998, Assessing the Impact of Volcanic Activity on Mid-Holocene Climate in Ireland: The Need for Replicate Data, *The Holocene*, Vol.8., pp.105-111.

115. Chambers, F.M., Booth, R.K., De Vleeschouwer, F., Lamentowicz, M., Le Roux, G., Mauquoy, D., Nichols, J.E., van Geel, B., 2012, Development and refinement of proxy-climate indicators from peats, *Quaternary International,* Vol.268., pp.21–33.

116. Chambers.F., Blackford.J.J., 2001, Mid to Late Holocene Climatic Changes: A Test of Periodicity and Solar Forcing in Proxy Climate Data from Blanket Peat Bogs, *Journal of Quaternary Science*, Vol.16., pp.392-438.

117. Chambers.F.M., Elliot.L., 1989, Spread and Expansion of Alnus Mill in the British Isles: Timing, Agencies and Possible Vectors, *Journal of Biogeography*, Vol.16., pp.541-550.

118. Chambers.F.M., Ogle.M.I., Blackford.J.J., 1999, Palaeoenvironmental Evidence for Solar Forcing of Holocene Climate: Linkages to Solar Science, *Progress in Physical Geography*, Vol.23., pp.181-204.

119. Chandler.C., Cheney.P., Thomas.P., Trabaud.L., Williams.D., 1983, *Fire in Forestry, Volume 1: Forest Fire Behavior and Effects,* Wiley.

120. Chapman.D.S., Bartlett.M.G., Harris.R.N., 2004, Comment on "Ground vs. Surface Air Temperatures Trends: Implications for Borehole Surface Temperature Reconstructions" by Mann.M.E. and Schmidt.G., *Geophysical Research Letters*, Vol.31.

121. Chapman.M.R., Shackleton.N.J., 2000, Evidence of 550-Year and 1000-Year Cyclicities in North Atlantic Circulation Patterns During the Holocene, *The Holocene*, Vol.10., pp.287-291.

122. Charman.D.J., 1990, *Origins and Development of the Flow Country Blanket Mire, Northern Scotland, with particular reference to Patterned Fens*, Ph.D. thesis, University of Southampton, UK.

123. Charman.D.J., 1994, Late-Glacial and Holocene Vegetation History of The Flow Country, Northern Scotland, *New Phytologist,* Vol.127., pp.155-168.

124. Charman.D.J., Blundell.A., Chiverrall.R.C, Hendon.D., Langdon.PG., 2006, Compilation of Non-Annually Resolved Holocene Proxy Climate Records: Stacked Holocene Peatland Palaeo-water Table Reconstructions from Northern Britain, *Quaternary Science Reviews*, Vol.25., pp.336-350.

125. Charman.D.J., Caseldine.C., Baker.A., Gearey.B., Hatton.J., Proctor.C., 2001, Paleohydrological Records from Peat Profiles and Speleothems in Sutherland, Northwest Scotland, *Quaternary Research,* Vol.55., pp.223-234.

126. Charmna.D., 2002, Blanket Mire Formation at Cross Lochs, Sutherland, Northern Scotland, *Boreas,* Vol.21., pp.53-72.

127. Clark.J.S., 1988, Particle Motion and the Theory of Charcoal Analysis: Source Area, Transport, Deposition and Sampling, *Quaternary Research,* Vol.30., pp.81-91.

128. Clark.R.L., 1982, Point Count Estimation of Charcoal in Pollen Preparations and Thin Sections of Sediment, *Pollen et Spores*, Vol.24., pp.523-535.

129. Coles.B.J., 1998, Doggerland: A Speculative Survey, *Proceedings of the Prehistoric Society*, Vol.64., pp.45-81.

130. Collard.M., Buchanan.B., Hamilton.M.J., O'Brien.M.J., 2010, Spatiotemporal Dynamics of the Clovis-Folsom Transition, *Journal of Archaeological Science,* Vol.37., pp.2513-2519.

131. Committee on Surface Temperature Reconstructions for the Last 2,000 Years., Council, N.R., 2006, *Surface Temperature Reconstructions for the Last 2,000 Years*. The National Academies Press.

132. Cook.E.R., Briffa.K.R., Meko.D.M., Graybill.D.A., Funkhouser.G., 1995, The 'Segment Length Curse' in Long Tree-Rings Chronology Development for Paleoclimatic Studies, *The Holocene,* Vol.5., pp.229-237.

133. Coope.G.R., 1975, Climatic Functions in Northwest Europe Since the Last Interglacial Indicated by Fossil Assemblages of Coleoptera, Ice Ages: Ancient and Modern, (ed) Wright.A.E. and Moseley.F., *Geology Journal*, Vol.6., pp.153-168.

134. Coope.G.R., 1977, Fossil Coleopteran Assemblages as Sensitive Indicators of Climatic Changes During the Devensian (Last) Cold Stage, *Philosophical Transactions of the Royal Society of London* B280, pp.313-340.

135. Coope.G.R., Joachim.M.J., 1980, Lateglacial Environmental Changes Interpreted from Fossil Coleptera from St Bees, Cumbria, Northwest England, In: Lowe.J.J., Gray.J.M., Robinson.J.E., *Studies in the Lateglacial of Northwest Europe*, Oxford, Pergamon Press, pp.55-68.

136. Coope.G.R., Pennington.W., 1977, The Windermere Interstadial of the Late Devensian, *Philosophical Transactions of the Royal Society of London* Vol.280., pp.337-39.

137. Cowley.D.C., 1998, Identifying Marginality in the First and Second Millennia BC in Strath of Kildonan, Sutherland, In: *Life on the Edge: Human Settlement and Marginality* (eds) Mills.C.M., Coles.G., Oxbow, Oxford, pp.165-171.

138. Crawford.R.M.M., 2000, Ecological Hazards of Oceanic Environments, *New Phytologist,* Vol.147., pp.257-281.

139. Crowley.T.J., North.G.R., 1991, *Paleoclimatology,* Oxford, Clarendon Press.

140. Curlevski.N.J.A., Xu.Z.H., Anderson.I.C., Cairney.J.W.G., 2010, Diversoty fo Soil and Rhiosphere Fungi Under Araucaria bidwillii (Bunya pine) at an Australian Tropical Montane Rainforest Site, *Fungal Divers,* Vol.40., pp.1.-11.

141. Cushing.D.H., 1982, *Climate and Fisheries*, London, Academic Press.

142. Dagg.C., 1998, *Proposed Woodland Grant Scheme at An Torr, Badachro, Gairloch,* Archaeological Survey.

143. Dagg.C., 2010, *Shieldaig Farm, Gairloch, Wester Ross, Proposed New Planting Areas,* Archaeological Desk-Based Evaluation, pp.1-11.

144. Daniell.J.R.G., 1997, *The Late-Holocene Palaeoecology of Scots pine (Pinus sylvestris L.) in North-West Scotland*, Doctoral Thesis, Durham University.

145. Dansgaard.W., Johnsen.S.J., Clausen.H.B., Dahl-Jensen.D., Gundestrup.N.S., Hammer.C.U., Hvidberg.C.S., Steffensen.J.P., Sveinbjörnsdottir.A.E., Jouzel.J., Bond.G., 1993, Evidence for General Instability of Past Climate from a 250-kyr Ice Core Record, *Nature*, Vol.364., pp.218-220.

146. Dansgaard.W., Oeschger.H., 1989, Past Environmental Long-Term Records from the Artic, In *The Environmental Record in Glaciers and Ice Sheets,* (eds) H.Oeschger and C.Langway, Jr., John Wiley, New York, pp.1989.

147. Dansgaard.W., White.J.C., Johnsen.S.J., 1989, The Abrupt Termination of the Younger Dryas Climate Event, *Nature*, Vol.339., pp.532-534.

148. Davies.A.L., 1999, *High Spatial Resolution Holocene Vegetation and Land-Use History in West Glen Affric and Kintail, Northern Scotland,* Unpublished Ph.D. thesis, University of Sterling.

149. Davies.A.L., 2003, Carnach Mor and Camban: Woodland History and Land-use in Alluvial Settings, In: *The Quaternary of Glen Affric and Kintail, Field Guide*, (eds) Tipping.R.M., Quaternary Research Association, London, pp.75-84.

150. Davies.A.L., 2007, Upland Agriculture and Environmental Risk: A New Model of Upland Land-Use Based on High Spatial Resolution Palynological Data from West Affric, Northwest Scotland, *Journal of Archaeological Science*, Vol.34., pp.1-11.

151. Davies.A.L., Tipping.R., 2004, Sensing Small Scale Human Activity in the Palaeocological Record: Fine Spatial Resolution Pollen Analysis from West Glen Affric, Northern Scotland, *The Holocene*, Vol.14., pp.233-245.

152. Davies.A.L., Tisdall.E., Tipping.R., 2004, Holocene Climatic Variability and Human Settlement in the Scottish Highlands: Fragility and Robustness. In: *Atlantic Connections and Adaptations* (eds) Housley.R.A., Coles.G.M., Oxbow, Oxford, pp.2-11.

153. Davies.B.A.S., Brewer.S., Stevenson.A.C., Guiot.J., Data Contributors, 2003, The Temperature of Europe During the Holocene Reconstructed from Pollen Data, *Quaternary Science Reviews*, Vol.22., pp.1701-1716.

154. Davies.B.A.S., Brewer.S., Stevenson.A.C., Guiotc.J., 2003, The Temperature of Europe during the Holocene Reconstructed from Pollen Data, *Quaternary Science Reviews,* Vol.22., 1701-1716.

155. Davies.T.D., Abrahams.P.W., Trander.M., Blackwood.I., Brimblecombe.P., Vincent.C.E., 1984, Black Acidic Snow in the Remote Scottish Highlands, *Nature,* Vol.312., pp.58-61.

156. De.Vleeschouwer.F., Chambers.F.M., Swindles.G.T., 2010, Coring and Sub-sampling of Peatlands for Palaeoenvironmental Research*, Mires and Peat*, Vol.7., pp.1-10.

157. Denison.S., 2001, Earliest Evidence Found of Settlers in Scotland, *British Archaeology*, Vol.60, pp.1-4.

158. Dennell.R., 1983, *European Economic Prehistory: A New Approach*, London, Academic Press.

159. Denton.G.H., Hughes.T.J., 1981, *The Last Great Ice Sheets,* Wiley, New York.

160. Denton.G.H., Karlen.W., 1973, Holocene Climatic Variations – Their Pattern and Possible Cause, *Quaternary Research,* Vol.3., pp.155-205.

161. Desprat.S., Combourieu-Nebout.N., Essallami.L., Sicre.M.A., Dormoy.I., Peyron.O., Siani.G., Bout Roumazeilles,V., Turon.J.L., 2013, Deglacial and Holocene Vegetation and Climatic Changes in the Southern Central Mediterranean from a Direct Land-Sea Correlation, *Climate of the Past*, Vol.9., pp.767-787.

162. Dickson.B., Yashayaev.I., Meincke.J., Turrell.B., Dye.S., Holfort.J., 2002, Rapid Freshening of the Deep North Atlantic Ocean over the Past Four Decades, *Nature*, Vol.416., pp.832-837.

163. Dobbertin.M., Wermelinger.B., Bigler.C., Burgi.M., Carron.M., Forster.B., Gimmi.U., Rigling.A., 2007, Linking Increasing Drought Stress to Scots Pine Mortality and Bark Beetle Infestations, *Science World Journal*, Vol.7., pp.231-239.

164. Donner.J., 1982, Fluctuations in Water Level of the Baltic Ice Lake, *Annales Academiae Scientiarum Fennicae, AIII,* Vol.134., pp.13-28.

165. Dubois.A.D., Ferguson.D.K., 1985, The Climatic History of Pine in the Cairngorms based on Radiocarbon Dates and Stable Isotope Analysis, with an Account of the Events Leading up to its Colonisation, *Review of Palaeobotany and Palynology*, Vol.46., pp.55-80.

166. Dugmore.A.D., Cook.G.T., Shore.S., Newton.A.J., Edwards.K.J., Larsen.G., 1995, Radiocarbon Dating Tephra Layers in Britain and Iceland, *Radiocarbons,* Vol.37., pp.379-388.

167. Dugmore.A.J., 1989, Icelandic Volcanic Ash in Scotland, *Scottish Geographical Magazine*, Vol.105., pp.168-172.

168. Dumfies and Galloway Biodiversity Partnership, 2009, *Conifer Plantations,*pp.183-195.

169. Durno.S.E., McVean.D.N., 1959, Forest History of the Beinn Eighe Nature Reserve, *New Phytologist,* Vol.58., pp.228-236.

170. Edwards.K.E., 1981, The Separation of Corylus and Myrica Pollen in Modern and Fossil Samples, *Pollen Spores*, Vol.23., pp.205-218.

171. Edwards.K.J. 1990, Fire and the Scottish Mesolithic: Evidence from Microscopic Charcoal, In: *Contributions to the Mesolithic in Europe,* (eds) Vermeesch.P.M., Van Peer.P., Leuven: University Press, pp.71-79.

172. Edwards.K.J., 1982, Man, Space and the Woodland Edge – Speculations on the Dectection and Interpretation of Human Impacts in Pollen Profiles, In: *Archaeological Aspects of Woodland Ecology,* (eds) Bell.M., Limbrey.S., BAR, Oxford, pp.5-22.

173. Edwards.K.J., Mithen.S., 1995, The Colonization of the Hebridean Islands of Western Scotland: Evidence from the Palynological and Archaeological Records, *World Archaeology*, Vol.26., pp.348-365.

174. Edwards.K.J., Whittington.G., 1997, Vegetation Change, In: *Scotland: Environment and Archaeology, 8000BC-AD1000*, (eds) Edwards.K.J., Ralston.I.B.M., Wiley, Chichester, pp.63-82.

175. Edwards.K.J., Whittington.G., Hirons.K.R., 1995, The Relationship Between Fire and Long-Tern Wet Heath Deveoplment in South Uist, Scotland, In: *Heath and Moorlands: Cultural Landscapes,* (eds) Thompson.D.B.A., Hester.A.J., Usher.M.B., HMSO: Edinburgh, pp.240-248.

176. Einarsson.T., 1986, Tephrochronology, In: *Handbook of Holocene Paleoecology and Paleohydrology,* (eds) Berglund, B.E., Chichester, John Wiley and Sons, pp.329-342.

177. Ellis.C.J., 2005, Blanket Mire Development, *Climate Change and Human Land-Use,* Royal Botanic Garden Edinburgh.

178. Ellis.C.J., Tallis.J.H., 2000, Climatic Control of Blanket Mire Development at Kentra Moss, Northwest Scotland, *Journal of Ecology*, Vol.88., pp.869-889.

179. Ennos.R.A., Sincalir.W.T., Perks.M.T., 1997, Genetic Insights into the Evolution of Scots Pine, *Pinus sylvestris L.,* in Scotland, *Botanical Journal of Scotland*, Vol.49., pp.257-265.

180. Ennos.R.A., Worrell.R., Malcolm.D.C., 1998, The Genetic Management of Native Species in Scotland, *Foresty,* Vol.71., pp.1-23.

181. EPICA Community Members, 2004, Eight Glacial Cycles from an Antarctic Ice Core, *Nature,* Vol.429, pp.623-628.

182. Erdtman.G., Berglund.B.E., Praglowski.J., 1961, *An Introduction to a Scandinavian Pollen Flora*, Stockholm: Almquist and Wiksells.

183. Everest.J.D., Bradwell.T., Fogwill.C.J., Kubik.P.W., 2006, Cosmosgenic ^{10}Be Age Constraints for the Wester Ross Re-Advance Moraine: Insights into British Ice Sheet Behaviour, *Geografisker Annaler,* Vol.88a., pp.9-18.

184. Faegri.K., and Iversen.J., 1989, *Textbook of Pollen Analysis*, 4th Edition, Wiley.

185. Fairbanks.R., Mortlock.R., Chiu.T., Cao.L., Kaplan.A., Guilderson.T., Fairbanks.T., Bloom.A., 2005, Marine Radiocarbon Calibration Curve Spanning 0 to 50,000 Years B.P. Based on Paired ^{230}Th/^{234}U/^{238}U and ^{14}C Dates on Pristine Corals, *Quaternary Science Reviews*, Vol.24., pp.1781-1796.

186. Finlayson.B., Edwards.K.J., 1997, The Mesolithic, In: *Scotland: Environment and Archaeology, 8000BC –AD1000,* (eds) Edwards.K.J., and Ralston.I.B.M., Chichester, John Wiley, pp.109-125.

187. Forestry Commission, 1998, *Caledonian Pinewood Inventory*, Edinburgh.

188. Forestry Commission, 2014 Available at: maps.forestry.gov.uk

189. Forrest.G.I., 1980, Genotypic Variation Among Native Scots Pine Populations in Scotland Based on Monoterpene Analysis, *Forestry*, Vol.53., pp.101-128.

190. Fossitt.J.A., 1994a, Lateglacial and Holocene Vegetational History of Western Donegal, Ireland, Biology and the Environment, *Proceedings of the Royal Irish Academy*, Vol.94b., pp.1-31.

191. Fossitt.J.A., 1994b, Modern Pollen Rain in the Northwest of the British Isles, *Holocene*, Vol.4., pp.465-476.

192. Fossitt.J.A., 1996, Late Quaternary Vegetation History of the Western Isles of Scotland, New Phytologist, Vol.132., pp.171-196.

193. Fowler.P.D., 1984, Transfer to Terrestrial Surfaces, *Philosophical Transactions of the Royal Society of London*, Vol.305., pp.281-297.

194. Froyd.C.A., 2006, Holocene Fire in the Scottish Highlands: Evidence from Macroscopic Charcoal Records, *The Holocene*, Vol.16., pp.235-249.

195. Futty.D.W., Towers.W., 1982, *Soil and Land Capability for Agritculture: Northern Scotland, Soil Survey of Scotland,* Macaulay Institute for Soil Research, Aberdeen.

196. Gear.A.J., 1989, *Holocene Vegetation History and The Palaeoecology of Pinus Sylvestris in Northern Scotland,* Durham Theses, Durham University.

197. Gear.A.J., and Huntley.B., 1991, Rapid Changes in the Range Limits of Scots Pine 4000 Years Ago, *Science*, Vol.251., pp.544-547.

198. Geirsdóttir.A., Miller.G., Larsen.D., Ólafsdóttir.S., 2013, Abrupt Holocene Climate Transitions in the Northern North Atlantic Region Recorded by Synchronized Lacustrine Records in Iceland, *Quaternary Science Reviews,* Vol.70., pp.48-62.

199. Gilbertson.D.D., Schwenninger.J.L., Kemp.R.A., Rhodes.E.J., 1999, Sand-Drift and Soil Formation Along an Exposed North Atlantic Coastline: 14,000 years of Diverse Geomorphological, Climatic and Human Impacts, *Journal of Archaeological Science*, Vol.26., pp.439-469.

200. Giralt.S., Burjachs.F., Roca.J.R., Julia.R., 1999, Lateglacial to Early Holocene Environmental Adjustment in the Mediterranean Semi-Arid Zone of the Salinas Playa-Lake Alicante, Spain, *Journal of Paleolimnology,* Vol.21., pp.449-460.

201. Godwin.H., 1956, *The History of British Flora,* Cambridge University Press.

202. Godwin.H., 1975, *History of the British Flora: A Factual Basis for Phytogeography*, Cambridge, Cambridge University Press.

203. Godwin.H., Conway.V.M., 1939, The Ecology of a Raised Bog Near Tregaron, Cardiganshire, *Journal of Ecology,* Vol.27., pp.313-363.

204. Gogorza.C.S., Sinito.A.M., Tommaso.I.D., Vilas.J.F., Creer.K.M., Nunez.H., 2000, Geomagnetic Secular Variations 0-12kyr as Recorded by Sediments from Lake Moreno (Southern Argentina), *Journal of South American Earth Sciences*, Vol.13., pp.627-645.

205. Gonzalez-Samperiz.P., Utrilla.P., Mazo.C., Valero-Garces.B., Sopena.M.C., Morellon.M., Moreno.A., Martinez-Bea.M., 2009, Patterns of Human Occupation During the Holocene in the Central Ebro Basin (Northeast Spain) in Response to the 8.2ka Climatic Event, *Quaternary Research*, Vol.71., pp.121-132.

206. Goslar.T., 2002, *^{14}C as an Indicator of Solar Variability*, Discussion Paper ESF-HOLIVAR Workshop, Lammi Finland.

207. Goslar.T., Arnold.M., Tisnerat-Laborde.N., Czernik.J., Wieckowski.K., 2000, Variations of Younger Dryas Atmospheric Radiocarbon Explicable Without Ocean Circulation Changes, *Nature,* Vol.403., pp.877-880.

208. Gratton.J., Charman.D.J., 1994, Non-Climatic Factors and the Environmental Impact of Volcanic Volatiles: Implications of the Lake Fissure Eruption of AD 1783, *The Holocene*, Vol.4., pp.101-106.

209. Gray.A.F., 1960, A Collection of Stone Artefacts from Redpoint, Loch Torridon, Ross-Shire, *Proceedings of the Society of Antiquaries of Scotland*, Vol.93., pp.236-237.

210. Green.B.H., 1968, Factors Influencing the Spatial and Temporal Distribution of Sphagnum imbricatum Hornsch, *Journal of Ecology*, Vol.56., pp.47-58.

211. Grimm.E., 1991, *TILIA and TILIA GRAPH*, Illinois State Museum, Springfield.

212. Grove.J.M., 1979, The Glacial History of the Holocene, *Progress in Physical Geography*, Vol.3., pp.1-54.

213. Gunther.H., Rosenthal.W., Stawarz.M., Carretero.J.C., Gomez.M., Lozano.I., Serrano.O., Reistad.M., 1998, The Wave Climate of the Northeast Atlantic Over the Period 1955-1994: the WASA Was Hindcast, *The Global Atmosphere and Ocean System*, Vol.6., pp.121-163.

214. Hall.V.A., Pilcher.J.R., McCormac.F.G., 1993, Tephra- Dated Lowland Landscape History of the North of Ireland, *New Phytologist,* Vol.125., pp.193-202.

215. Hall.V.A., Pilcher.J.R., McCormac.F.G., 1994, Icelandic Volcanic Ash and the Mid-Holocene Scots Pine (Pinus sylvestris) Decline in the North of Ireland: No Correlation, *The Holocene*, Vol.4., pp.79-83.

216. Hallsdottir.M., 1996, Synthesis of the Holocene History of Vegetation in Northern Iceland, *Palaeoklimaforschung*, Vol.20., pp.203-214.

217. Halstead.P., 2000, Land Use in Postglacial Greece: Cultural Causes and Environmental Effects, In: *Landscape and Land Use in Postglacial Greece*, (eds) Halstead.T.C., Frederick.C., Sheffield Academic Press, Sheffield , pp.110-128.

218. Hammer.C.U., 1984, Traces of Icelandic Eruptions in the Greenland Ice Sheet, *Jokull,* Vol.34., pp.51-65.

219. Hammond.R.F., Van Der Krogt.G., Osinga.T., 1990, Vegetation and Water Tables on Two Raised Bog Remnants in County Kildare, In: *Ecology and Conservation of Irish Peatlands,* (eds) Doyle.G.J., Dublin, Royal Irish Academy, pp.121-134.

220. Hancock.M.H., Egan.S., Summer.R., Cowie.N., Amphlett.A., Rao.S., Hamilton.A., 2005, The Effect of Experimental Prescribed Fire on the Establishment of Scots Pine *Pinus sylvestris* Seedlings on Heather Calluna vulgaris Moorland, *Forest Ecology and Management,* Vol.212., pp.199-213.

221. Hancock.M.H., Egan.S., Summers.R., Cowie.N., Amphlett.A., Rao.S., Hamilton.A., 2005, The Effect of Experimental Prescribed Fire on the Establishment of Scots Pine Pinus sylvestris Seedlings on Heather Calluna vulgaris Moorland, *Forest Ecology and Management*, Vol.212., pp.139-144.

222. Hancock.M.H., Summers.R.W., Amphlett.A., Willi.J., 2009, Testing Prescribed Fire as a Tool to Promote Scots Pine *Pinus sylvestris* Regeneration, European Journal of Forest Research, Vol.128., pp.319-333.

223. Hardy.K., Wickham-Jones.C., 2003, Scotlands First Settlers: An Investigation into Settlement, Territoriality and Mobility during the Mesolithic in the Inner Sound, Scotland, First Results, In: 2003, *Mesolithic on the Move: Papers Presented at the Sixth International Conference on the Mesolithic in Europe*, (eds) Larson.L. et al Stockholm 2000, Oxbow Books, Oxford, pp.369-381.

224. Hargreaves.K.J., Milne.R., Cannell.M.G.R., 2003, Carbon Balance of Afforested Peatland in Scotland, *Forestry,* Vol.76., No.3., pp.299-317.

225. Harris.R.N., Chapman.D.S., 2001, Mid-Latitude (30-60N) Climatic Warming Inferred by Combining Borehole Temperatures With Surface Air Temperatures, *Geophysical Research Letters*, Vol.28., pp.747-750.

226. Harrison.S.P., Digerfeldt.G., 1991, European Lakes as Palaeohydrological and Palaeoclimatic Indicators, *Quaternary Science Reviews*, Vol.12., pp.233-248.

227. Helama.S., Linderholm.M., Timonen.J., Merilainen.J., Eronen.M., 2002, The Supra-Long ScotS Pine Tree Ring Record for Finnish Lapland: 2, Interannual to Centennial Variability in Summer Temperatures for 7500 years, *The Holocene*, Vol.12., pp.681-688.

228. Hicks.S., 2001, The Use of Annual Arboreal Pollen Deposition Values for Delimiting Tree-Lines in the Landscape and Exploring Models of Pollen Dispersal, *Review of Palaeobotany and Palynology,* Vol.117., pp.1-29.

229. Hirons.K.R., Edwards.K.J., 1990, Pollen and Related Studies at Kinloch, Isle of Rhum, Scotland, with Particular Reference to Possible Early Human Impacts on Vegetation, *New Phytologist*, Vol.116., pp.715-727.

230. Hódar.J.A., Castro.J., Zamora.R., 2003, Pine Processionary Caterpilla *Thaumetopoea pityocampa* as a New Threat for Relict Mediterranean Scots Pine Forests Under Climatic Warming, *Biology Conservation*, Vol.110., pp.123-129.

231. Hodell.D.A., Brenner.M., Curtis.J.H., Guilderson.T., 2001, Solar Forcing of Drought Frequency in the Maya Lowlands, *Science,* Vol.292., pp.1367-1370.

232. Holling.C.S., 1992, Cross-Scale Morphology, Geometry and Dynamics of Ecosystems, *Ecological Monographs*, Vol.62., pp.447-502.

233. Hosfield.R.T., Straker.V., Gardiner.P., Brown.A.G., Davies.P., Fyfe.R., Jones.J., Tinsley.H., 2008, *Palaeolithic and Mesolithic, The Archaeology of South West England, South West Archaeological Research Framework: Research Assessment and Research Agenda,* Somerset County Council, Taunton, pp.23-62.

234. Hoyt.D.V., Schatten.K.H., 1997, *The Role of the Sun in Climate Change,* Oxford, Oxford University Press, pp.279.

235. http://digimap.edina.ac.uk (Accessed 10 May 2014)

236. Huang.S., Pollack.H.N., Shen.P.Y., 2000, Temperature Trends Over the Past Five Centuries Reconstructed from Borehole Temperatures, *Nature,* Vol.403., pp.756-758.

237. Hubbard.A., 1999, High-Resolution Modelling of the Advance of the Younger Dryas Ice Sheet and its Climate in Scotland, *Quaternary Research*, Vol.52., pp.27-43.

238. Hughen.K.A., Overpeck.J.T., Lehman.S.J., Kashgarians.M., Southon.J., Peterson.L.C., Alley.R., Sigman.D.M., 1998, Deglacial Changes in Ocean Circulation from an Extended Radiocarbon Calibration, *Nature,* Vol.391., pp.65-68.

239. Hughen.K.A., Southon.J.R., Lehman.S.J., Overpeck.J.T., 2000, Synchronous Radiocarbon and Climate Shifts During the Last Deglaciation, *Science*, Vol.290., pp.1951-1954.

240. Hughes.P.D.M.,Mauquoy.D., Barber.K.E., Langdon.P.D., 2000, Mire-Development Pathways and Palaeoclimatic Records from a Full Holocene Peat Archive at Walton Moss, Cumbria, England, *The Holocene,* Vol.10., pp.465-479.

241. Hunt.J.B., 1993, Report on the 2nd UK Tephra Workshop, *Quaternary Newsletter*, Vol.63., pp.41-44.

242. Hunt.J.B., 1997, A Simulation of the Possible Consequences of a Volcanic Eruption on the General Circulation of the Atmosphere, *Monthly Weather Review,* Vol.105., pp.247-260.

243. Huntley.B., 1990, European Post-Glacial Forests: Compositional Change, *Journal of Vegetational Science*, Vol.1., pp.507-518.

244. Huntley.B., 1993, Rapid Early-Holocene Migration and High Abundance of Hazel (*Corylus avellana* L.): Alternative Hypotheses, In: *Climate Change and Human Impact on the Landscape* (ed), Chambers.F., London, Chapman and Hall, pp.205-216.

245. Huntley.B., Birks.H.J.B., 1983, *An Atlas of Past and Present Pollen Maps for Europe: 0-13,000 Years Ago*, Cambridge University Press, Cambridge.

246. Huntley.B., Daniell.J.R.G., Allen.J.R.M., 1997, Scottish Vegetation History: The Highlands, *Botanical Journal of Scotland,* Vol.49., pp.163-176.

247. Indermühle.A., Monnin.E., Stauffer.B., Stocker.T.F., 2000, Atmospheric CO_2 Concentration from 60 to 20kyrBP from the Taylor Dome Ice Core, Antarctica, *Geophysical Research Letters*, Vol.27., pp.735-738.

248. Ingram.H.A.P., 1978, Soil Layers in Mires: Function and Terminology, *Journal of Soil Science*, Vol.29., pp.224-227.

249. IPCC, 2014, *Climate Change 2014: Impacts, Adaptation, and Vulnerability. Part A: Global and Sectoral Aspects. Contribution of Working Group II to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, (eds) Field.C.B., Barros.V.R., Dokken.D.J., MachK.J., Mastrandrea.M.D., Bilir.T.E., Chatterjee.M., Ebi.K.L., Estrada.Y.O., Genova.R.C., Girma.B., Kissel.E.S., Levy.A.N., MacCracken.S., Mastrandrea.P.R., White.L., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

250. Isarin.R.F.B., Renssen.H., 1999, Reconstructing and Modelling Late Weichselian Climates: the Younger Dryas in Europe as a Case Study, *Earth-Science Reviews,* Vol.48., pp.1-38.

251. Isarin.R.F.B., Renssen.H., Vandenberghe.J., 1998, The Impact of the North Atlantic Ocean on the Younger Dryas Climate in North Western and Central Europe, *Journal of Quaternary Science,* Vol.13., pp.447-453.

252. Jacobi.R.M., Tallis.J.H., Mellars.P.A., 1976, The Southern Pennine Mesolithic and Ecological Record, *Journal of Archaeological Science,* Vol.3., pp.307-320.

253. Jansen, E., J. Overpeck, K.R. Briffa, J.-C. Duplessy, F. Joos, V. Masson-Delmotte, D. Olago, B. Otto-Bliesner, W.R. Peltier, S.Rahmstorf, R. Ramesh, D. Raynaud, D. Rind, O. Solomina, R. Villalba and D. Zhang, 2007: Palaeoclimate. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA

254. Jessen.C., Rundgren.M., Bjorck.S., Hammarlund.D., 2005, Abrupt Climatic Changes and an unstable Transition into a Late Holocene Thermal Decline: a Multiproxy Lacustrine Record from Southern Sweden, *Journal of Quaternary Science*, Vol.20., pp.346-362.

255. Johnson.I.L., Bonsall.C., 1999, Mesolithic Adaptations on Offshore Islands: the Aleutians and Western Scotland, In: *Den Bogen Spannen…Festschrift fur Bernhard Gramsch,* Cziesla.E., (eds) Kersting.T., Pratsch.S., pp.99-106.

256. Johnson.R.G., McClure.B.T., 1976, A Model for Northern Hemisphere Continental Ice Sheet Variation, *Quaternary Research,* Vol.6., pp.325-353.

257. Joint Nature Conservation Committee, 2011, *Report no.445: Towards an Assessment of the State of UK Peatlands.*

258. Joosten.J.H.J., 1995, Between Diluvium and Deluga: The Origin of the Younger Dryas Concept, *Geologie en Mijnbouw*, Vol.74, pp.237-240.

259. Juggins. S., 2007, *C2 Version 1.5 User guide: Software for Ecological and Palaeoecological Data Analysis and Visualisation*, Newcastle University, Newcastle upon Tyne, UK, pp.73.

260. Karlen.W., 1976, Lacustrine Sediments and Tree-Limit Variations as Indicators of Holocene Climatic Fluctuations in Lapland: Northern Sweden, *Geografiska Annaler*, Vol.58A., pp.1-34.

261. Karlen.W., Kuylenstierna.J., 1996, On Solar Forcing of Holocene Climate: Evidence from Scandinavia, *The Holocene*, Vol.6., pp.359-365.

262. Keatinge.T.H., Dickson.J.H., 1979, Mid-Flandrian Changes in Mainland Orkney, *New Phytologist,* Vol.82., pp.585-612.

263. Keigwin.L.D., Boyle.E.A., 2000, Detecting Holocene Changes in Thermohaline Circulation, *Proceedings of the National Academy of Science,* Vol.97., pp.1343-1346.

264. Kelly.P.M., Sear.C.B., 1984, Climatic Impact of Explosive Volcanic Eruptions, *Nature,* Vol.311., pp.740-743.

265. Kerslake.P., 1982, *Vegetational History of Wooded Islands in Scottish Lochs*, Unpublished Ph.D. Thesis, University of Cambridge.

266. Kinloch.B.B., Westfall.R.D., Forrest.G.I., 1986, Caledonian Scots Pine: Origins and Genetic Structure, *New Phytologist*., Vol.104., pp.703-729.

267. Kinnes.I.A., 1985, Circumstance Not Context: The Neolithic of Scotland as seem from the Outside, *Proceedings of the Society of Antiquaries of Scotland,* Vol.108., pp.80-93.

268. Kirk.W., Godwin.H., 1963, A Lateglacial Site at Loch Droma, Ross and Cromarty. *Transactions of the Royal Society of Edinburgh,* Vol.65., pp.225-249.

269. Klimenko.V.V., Klimanov.V.A., Fedorov.M.V., 1996, The History of the Mean Temperature of the Northern Hemisphere over the Last 11,000 Years, *Trans. Russ. Acad. Sci.,* Vol.4., pp.626-629.

270. Klitgaard-Kristensen.D., Sejrup.P., Haflidason.H., Johnsen.S., Spurk.M.A., 1998, The Regional 8200 cal yr B.P. Cooling Event in Northwest Europe, Induced by Final Stages of the Laurentide Ice-Sheet Deglaciation, *Journal of Quaternary Science*, Vol.13., pp.165-169.

271. Koç.N. and Jansen.E., 1994, Response of the High-Latitude Northern Hemisphere to Orbital Climate Forcing: Evidence from the Nordic Seas, *Geology,* Vol.22., pp.523-526.

272. Koc.N., Jansen.E., 1992, A High Resolution Diatom Record of the Last Deglaciation from Southeast Norwegian Sea: Documentation of Rapid Climatic Changes, *Paleoceanography,* Vol.7., pp.499-520.

273. Koc.N., Jansen.E., Haflidason.H., 1993, Paleoceanographic Reconstructions of Surface Ocean Conditions in Greenland, Iceland, and Norwegian Seas Through the Lst 14 ka Based on Diatoms, *Quaternary Science Review*, Vol.12., pp.115-140.

274. Koç.N., Jansen.E., Hald.M., Labeyrie.L., 1996, Late Glacial Holocene Sea Surface Temperatures and Gradients between the North Atlantic and the Norwegian Sea; Implications for the Nordic Heat Pump., In: *Late Quaternary Palaeoceanography of the North Atlantic Margins,* (eds) Andrews.J.T., Austin.W.E.N., Bergsten.H., and Jennings.A.E., Geological Society Special Publication., Vol.111., pp.177-185.

275. Koerner.R.M., Fisher.D., 1982, Acid Snow in the Canadian High Arctic, *Nature*, Vol.295., pp.137-138.

276. Kolstrup.E., 1992, Danish Pollen Records Radio-carbon Dated to Between 50000 and 57000 yr BP, *Journal of Quaternary Science*, Vol.7., pp.163-172.

277. Korhola.A., Vasko.K., Toivonen.H.T.T., Olander.H., 2002, Holocene Temperature Changes in Northern Fennoscandia Reconstructed from Chironomids using Bayesian Modelling, *Quaternary Science Reviews,* Vol.21., pp.1841-1860.

278. Kullman.L., 1983b, Short-Term Population Trends of Isolated Tree-Limit Stands of *Pinus sylvestris L.* in Central Sweden, Artic and Alpine Research, Vol.15., pp.369-382.

279. La Marche.V.C., Hirschboeck.K.K., 1984, Frost Rings in Trees as Records of Major Volcanic Eruptions, *Nature*, Vol.307., pp.121-126.

280. Lageard.J.G.A., Chambers.F.M., Thomas.P.A., 1999, Climatic Significance of the Marginalization of Scots Pine (Pinus Sylvestris L.) c.2500BC at White Moss, South Cheshire, UK., *The Holocene*, Vol.9., pp.321-331.

281. Laine.J., Vasander.H., Sallantaus.T., 1995, Ecological Effects of Peatland Drainage for Forestry, *Environmental Reviews*, Vol.3., pp.286-303.

282. Laj.C., Kissel.C., Mazaud.A., Michel.E., Muscheler.R., Beer.J., 2002, Geomagnetic Field Intensity, North Atlantic Deep Water Circulation and Atmospheric $\Delta^{14}C$ during the Last 50kyr, *Earth and Planetary Science Letters,* Vol.200., pp.177-190.

283. Lamb.H.H., 1964, Trees and Climatic History in Scotland, *Quarterly Journal of the Royal Meteorological Society,* Vol.90., pp.382-394.

284. Lamb.H.H., 1977, Climate History and the Future, Vol.2., *Climate: Past, Present and Future,* Methuen, New York.

285. Lamb.H.H., 1979, Climatic Variation and Changes in the Wind and Ocean Circulation: the Little Ice Age in the Northeast Atlantic, *Quaternary Research*, Vol.11., pp.1-20.

286. Lamb.H.H., 1982, *Climate History and the Modern World,* Vol.387., pp.387., Methuen, London.

287. Landmann.G., Reimer.A., 1996, Climatically Induced Lake Level Changes at Lake Van, Turkey, During the Pleistocene/Holocene Transition, *Global Biogeochemical Cycles,* Vol.10., pp.797-808.

288. Langdon.P.G., Barber.K., 2002, The 'AD 860' Tephra in Scotland: New Data from Langlands Moss, East Kilbride, Strathclyde, *Quaternary Newsletter*, Vol.97., pp.11-18.

289. Langdon.P.G., Barber.K.E., Hughes.P.D.M., 2003, A 7500 Year Peat Based Palaeoclimatic Record Reconstruction and Evidence for an 1100 year Cyclicity in Bog Surface Wetness from Temple Hill Ross, Pentland Hills, SE England, *Quaternary Science Reviews*, Vol.22., pp.259-274.

290. Larsen.G. and Thorarinsson.S., 1977, H-4 and Other Acid Hekla Tephra Layers, *Jokull*, Vol.27., pp.28-46.

291. Lauritzen.S.E., Lundberg.J., 1999, Calibration of the Speleothem Delta Function: An Absolute Temperature Record for the Holocene in Northern Norway, *The Holocene,* Vol.9., pp.659-669.

292. Lawson.T.J., Bonsall.C., 1986, The Palaeolithic of Scotland: A Reconsideration of Evidence from Reindeer Cave, Assynt, in Collcutt.S.N., (eds), *The Palaeolithic of Britain and its Nearest Neighbours: Recent Trends,* Sheffield, University of Sheffield, pp.85-89.

293. Le Treut.H., Somerville.R., Cubasch.U., Ding.Y., Mauritzen.C., Mokssit.A., Peterson.T., and Prather.M., 2007, Historical Overview of Climate Change, In: *Climate Change 2007: The Physical Basis of Science, Contribution of Working Group 1 to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Solomon.S., Qin.D., Manning.M., Chen.Z., Marquis.M., Averyt.K.B., Tignor.M., and Miller.H.L., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

294. Lee.J.A., Press.M.C., Woodin.S., Ferguson.P., 1986, *Responses to Acidic Deposition in Ombrotrophic Mires*, in: Hutchinson.T.C. and Meema.K., Effects of Acidic

Deposition on Forests, *Wetlands and Agricultural Ecosystems,* Berlin, Springer-Verlag, pp.549-560.

295. Legg.C.J., McHaffie.H., Amphlett.A., Worrell.R., 2003, The Status of Wooded Bogs at Abernathy, Strathspey, in: *Restoring Natural Forest Habitat*, Conference Proceedings, Munlochy, pp.12-16.

296. Leuscher.H.H., Sass-Klaasen.U., Jansma.E., Baillie.M.G.L., Spurk.M., 2002, Subfossil European Bog Oaks: Population Dynamics and Long-Term Growth Depressions as Indicators of the Holocene Hydro-Regime and Climate, *The Holocene*, Vol.12., pp.695-706.

297. Leuschner.H.H., Bauerochse.A., Metzler.A., 2007, Environmental Change, Bog History and Human Impact Around 2900BC in NW Germany – Preliminary Results from Dendroecological Study of a Sub-Fossil Pine Woodland at Campemoor, Dümmer Basin*, Vegetation History and Archaeobotany,* Vol.16., pp.183-195.

298. Li, B., Nychka.D.W., and Amman.C.M., 2010, The Value of Multi-Proxy Reconstruction of Past Climate, *Journal of American Statistical Association*, Vol.105., No.491, pp.883-911.

299. Linderholm.H,. Moberg.A., Grudd.H., 2002, Peatland Pines as Climatic Indicators? A Regional Comparison of the Climatic Influence on Scots Pine Growth in Sweden, *Can. J. For. Research,* Vol.32, pp.1400-1410.

300. Lindsey.R.A., Charman.D.J., Everingham.F., O'Reilly.R.M., Palmer.M.A., Rowell.T.A., Stroud.D.A., 1988, *The Flow Country: The Peatlands of Caithness and Sutherland*, Peterborough, Nature Conversancy Council.

301. Lindsey.R.A., Immirzi.C.P., 1996, *An Inventory of Lowland Raised Bogs in Great Britain,* Scottish Natural Heritage, Battleby.

302. Lisiecki.L.E., Raymo.M.E., 2005, A Pliocene-Pleistocene Stack of 57 Globally Distributed Benthic $\delta^{18}C$ Records, *Paleoceanography,* Vol.20., pp.12-17.

303. Livens.R.G., 1956, Three tanged Flint Points from Scotland, *Proceedings of the Society of Antiquaries of Scotland*, Vol.89, pp.438-443.

304. Lowe.J., Walker.M., 2014, *Reconstructing Quaternary Environments*, 3RD Edition, Routledge, London.

305. Lowe.J.J., 1993, Isolating the Climatic Factors in Early and Mid-Holocene Palaeobotanical Records from Scotland, In: Chambers.F.M., (eds) *Climate Change and Human Impact on the Landscape,* Chapman and Hall, London, pp.67-82.

306. Lowe.J.J., Ammann.B., Birks.H.H., Bjorck.S., Coope.G.R., Cwynar.L., De Beaulieu.J.L., Mott.R.J., Peteet.D.M., Walker.M.J.C., 1994, Climatic Changes in Areas Adjacent to the North Atlantic During the Late Glacial-Interglacial Transition (14-9ka BP): A Contribution to IGCP-253, *Journal of Quaternary Science*, Vol.9., pp.185-198.

307. Lowe.J.J., Walker.M.J.C., 1977, The Reconstrcution of the Lateglacial Environment in the Southern and Eastern Grampian Highlands, In: *Studies in the Scottish Lateglacial Environment* (eds) Gray.J.M., Lowe.J.J., Oxford, Pergamon Press, pp.101-118.

308. Lowe.J.J., Walker.M.J.C., 1986, Lateglacial and Early Falndrian Environmental History on the Isle of Mull, Inner Hebrides, Scotland, Transactions of the Royal Society of Edinburgh, *Earth Sciences*, Vol.77., pp.1-20.

309. Lowe.J.J., Walker.M.J.C., 1991, Vegetational History of the Isle of Skye: II, The Flandrian, In: The Quaternary of the Isle of Skye: Field Guide (eds) Ballantyne.C.K., Benn.D.I., Lowe.J.J., Walker.M.J.C., Cambridge, *Quaternary Research Association*, Vol.119-142.

310. Lowell.T.V., Heusser.C.J., Andersen.B.G., Moreno.P.I., Hauser.A., Heusser.L.E., Schluchter.C., Marchant.D.R., Denton.G.H., 1995, Interhemispheric Correlation of Pleistocene Glacial Events, *Science,* Vol.269., pp.1541-1549.

311. Magny.M., 2004, Holocene Climate Variability as Reflected by Mid-European Lake Level Fluctuations, and its Probable Impact on Prehistoric Human Settlements, *Quaternary International*, Vol.113., pp.65-79.

312. Malmer.N., 1986, Vegetational Gradients in Relation to Environmental Conditions in Northwest European Mires, *Canadian Journal of Botany*, Vol.64., pp.375-383.

313. Mangerud.J., Anderson.S.T., Berglund.B.E., Donner.J.J., 1974, Quaternary Stratigraphy of Norden, A Proposal for Terminology and Classification, *Boreas,* Vol.3., pp.109-128.

314. Marchal.O., Stocker.T.F., Muscheler.R., 2001, Atmospheric Radiocarbon During the Younger Dryas: Production, Ventiliation, or Both?, *Earth and Planetary Science Letters,* Vol.185., pp.383-395.

315. Marsden.K., Ebmeier.S.K., 2012, *Peatlands and Climate Change*, Scottish Parliament Information Centre Briefing.

316. Mass.C.F., Portman.D.A., 1989, Major Volcanic Eruptions and Climate: A Critical Evaluation, Journal of Climate, Vol.2, pp.566-593.

317. MatiasL., Jump.A.S., 2012, Interactions between Growth, Demography and Biotic Interactions in Determining Species and Range Limits in a Warming World: The Case of *Pinus sylvestris, Forest Ecology Management,* Vol.282., pp.10-22.

318. Matthews.I.P., Birks.H.H., Bourne.A., Brooks.S.J., Lowe.J.J., MacLeod.A., Pyne-O'Donnell.S.D.F., 2011, New Age Estimates and Climatostratigraphic Correlations for the Borrobol and Penifiler Tephras: Evidence from Abernethy Forest, Scotland, *Journal of Quaternary Science,* Vol.26., pp.247-252.

319. Mauquoy.D., Barber.K., 1999, Evidence for Climatic Deteriorations Associated with the Decline in *Sphagnum imbricatum* Horsch, in Six Ombrotrophic Mires from Northern England and the Scottish Borders, *The Holocene*, Vol.9., pp.423-437.

320. Mauquoy.D., Engelkes.T., Groot.M.H.M., Markesteijn.F., Oudejans.M.G., Van Der Plicht.J., Van Geel.B., 2002a, High Resolution Records of Late Holocene Climate Change and Carbon Accumulation in two Northwest European Ombrotrophic Peat Bogs, *Palaeogeography, Palaclimatology, Palaeoecology*, Vol.186., pp.275-310.

321. Mauquoy.D., Van Geel.B., Blaauw.M., Van Der Plicht.J., 2002B, Evidence from Northwest European Bogs Show 'Little Ice Age' Climatic Changes Driven by Variations in Solar Activity, *The Holocene,* Vol.12., pp.1-6.

322. Mayewski.P.A., Rohling.E.E., Stager.J.C, Karlen.W., Maasch.K.A., Meeker.L.D., 2004, Holocene Climate Variability, *Quaternary Research,* Vol.62., pp.243-255.

323. McCormack.D., 2011, *The Style and Timing of the Last Deglaciation of Wester Ross, Northwest Scotland,* Thesis, University of Manchester

324. McCullagh.R.P.J., Tipping.R., 1998, *The Lairg Project 1988-1996, The Evolution of an Archaeological Landscape in Northern Scotland*, Scottish Trust for Archaeological Research, Edinburgh.

325. McDermott.F., Frisia.S., Huang.Y., Longinelli.A., Spiro.B., Heaton.T.H.E., Hawkesworth.C.J., Borsato.A., Keppens.E., Fairchild.I.J., Van Der Borg.K., Verheyden.S., Selmo.E., 1999, Holocene Climate Variability in Europe: Evidence from $\delta^{18}O$, Textural and Extension-Rate Variations in Three Speleothems, *Quaternary Science Reviews*, Vol.18., pp.1021-1038.

326. McVean.D.N., 1963, Growth and Mineral Nutrition of Scots Pine Seedlings on Some Common Peat Types, *Journal of Ecology*, Vol.51., pp.657-670.

327. McVean.D.N., and Ratcliffe.D.A., 1962, *Plant Communities of the Scottish Highlands,* London, HMSO Nature Conservancy.

328. Mellars.P.A., 1976, Fire Ecology, Animal Populations and Man: A Study of some Ecological Relationships in Prehistory, *Proceedings of the Prehistoric Society*, Vol.42., pp.15-42.

329. Mercer.J., 1980, Lussa Wood I: The Late Glacial and Early Post-Glacial Occupation of Jura, *Proceedings of the Antiquaries of Scotland*, Vol.110., pp.1-31.

330. Merrill.R.T., McElhinny.M.W., McFadden.P.L., 1996, The Magnetic Field of the Earth, Paleomagnetism, the Core, and the Deep Mantle, *International Geophysics Series*, Vol.63., San Diego, Academic Press, pp.531.

331. Meteorological Office, 1952. *Climatological Atlas of the British Isles.* London, H.M.S.O.

332. Mighall.T.M., Lageard.J.G.A., Chambers.F.M., Field.M.H., Mahi.P., 2004, Mineral Deficiency and the Presence of *Pinus sylvestris* on Mires During the Mid to Late Holocene: Palaeoecological Data from Cadogan's Bog, Mizen Peninsula, Co. Cork, Southwest Ireland, *The Holocene*, Vol.14., pp.95-109.

333. Miles.J., 1985, The Pedogenic Effects of Different Species and Vegetation Types and the Implications of Succession, *Journal of Soil Science*, Vol.36., pp.571-584.

334. Mithen.S., 2000, *Hunter-Gatherer Landscape Archaeology: The Southern Hebrides Mesolithic Project 1988-98*, Vol.1.

335. Mithen.S., 2000, The Scottish Mesolithic: Problems, Prospects and the Rationale of the Southern Hebrides Mesolithic Project, In: *Hunter Gatherer Landscape Archaeology: the Southern Hebrides Mesolithic Project 1988-1998,* (ed) Mithen.S., Cambridge, McDonald Institute for Archaeological Research, pp.9-37.

336. Moir.A.K., Leroy.S.A.G., Brown.D., Collins.P.E.F., 2010, Dendrochronological Evidence for a Lower Water-Table on Peatland around 3200-3000BC from Subfossil Pine in Northern Scotland, *The Holocene,* Vol.20., pp.931-942.

337. Montanarella L., Jones, R. J. A. and Hiederer, R., 2006, the Distribution of Peatland in Europe. *Mires and Peat*. Vol.1., pp.1-7.

338. Montoya.E., Rull.V., Van Geel.B., 2010, Non-pollen Palynomorphs from Surface Sediments Along an Altitudinal Transect of the Venezuelan Andes, *Palaeogeography, Palaeoclimatology, Palaeoecology,* Vol.297., pp.169-183.

339. Moore.P.D., 1972, The Initiation of Peat Formation and the Development of Peat Deposits in Mid-Wales., In: *Proceedings of the 4th International Peat Congress*, Vol.89., International Peat Society, Otaniemi, Finland.

340. Moore.P.D., 1986A, Hydrological Changes in Mires, In: Berglund.B.E. (eds) *Handbook of Holocene Palaeoecology and Palaeohydology,* pp.91-105, Wiley and Sons, Chichester.

341. Moore.P.D., 1986b, Man and Mire: A Long and Wet Relationship, *Transactions of the Botanical Society of Edinburgh,* Vol.45., pp.77-95.

342. Moore.P.D., Webb.J.A., Collinson.M.E., 1991, *Pollen Analysis*, 2nd Edition, Blackwell, London.

343. Moros.M., Andrews.J.T., Eberl.D.D., Jansen.E., 2006, Holocene History of Drift Ice in the Northern North Atlantic: Evidence for Different Spatial and Temporal Modes, *Paleoceanography,* Vol.21., pp.2-17.

344. Morrison.A., Bonsall..C., 1989, The Early Post Glacial Settlement of Scotland: A Review. In: *The Mesolithic in Europe,* (eds) Bonsall.C., Papers Presented at the Third International Symposium, Edinburgh 1985, John Donald, pp.134-142.

345. Muscheler.R., Beer.J., Wagner.G., Finkel.R.C., 2000, Changes in Deep-Water Formation During the Younger Dryas Event Inferred from ^{10}Be and ^{14}C Records, *Nature,* Vol.403., pp.567-570.

346. National Climatic Data Centre, *National Oceanic and Atmospheric Administration*, 2014, Department of Commerce, Unite States of America.

347. Neff.U., Burns.S.J., Mangini.A., Mudelsee.M., Fleitmann.D., Matter.A., 2001, Strong Coherence Between Solar Variability and the Monsoon in Oman Between 9 and 6kyr Ago, *Nature,* Vol.411., pp.290-293.

348. Nesji.A., Matthews.J.A., Olaf Dahl.S., Berrisford.M.S., Andersson.C., 2001, Holocene Glacier Flucuatations of Flatebreen and Winter Precipitation Changes in the Jostedalsbreen Region, Western Norway, Based on Glaciolacustrine Sediment Records, *The Holocene*, Vol.11., pp.267-280.

349. Newell.R.E., 1981, Further Studies of the Atmospheric Temperature Change Produced by the Mt Agung Eruption in 1963, *Journal of Volcanology and Geothermal Research,* Vol.11., pp.61-66.

350. Nicholls.N., 1988, Low Latitude Volcanic Eruptions and the El Nino Southern Oscillation, *Journal of Climatology*, Vol.8., pp.91-95.

351. Nicholls.N., 1990, Low Latitude Volcanic Eruptions and the El Nino Southern Oscillation, *International Journal of Climatology,* Vol.10., pp.425-429.

352. Nicoll.B.C., Ray.D., 1996, Adaptive Growth of Tree Root Systems in Response to Wind Action and Site Conditions, *Tree Physiol,* Vol.16., pp.891-898.

353. Nilssen.E., Vorren.K.D., 1991, Peat Humification and Climate History, *Norsk Geologist Tidsskrift,* Vol.71., pp.215-217.

354. O'Brien.S.R., Mayewski.P.A., Meeker.L.D., Meese.D.A., Twicckler.M.S., Whitlow.S.I., 1995, Complexity of Holocene Climate as Reconstructed from Greenland Ice Core, *Science,* Vol.270., pp.1962-1964.

355. O'Brien.S.R., Mayewski.P.A., Meeker.L.D., Meese.D.A., Twickler.M.S., Whitlow.S.I., 1995, Complexity of Holocene Climate as Reconstructed from a Greenland Ice Core, *Science*, Vol.270., pp.1962-1964.

356. O'Connell.C., 1990, Ecology and History of a Bog Pine Woodland at Derryclure, County Offaly, Ireland, *British Ecological Society Bulletin*, Vol.21., pp.116-121.

357. O'Sullivan.P.E., 1975, Early and Middle Flandrian Pollen Zonation in the Eastern Highlands of Scotland, *Boreas,* Vol.4., pp.197-207.

358. Ohlson.M., Økland.R.H., Nordbakken.J.F., Dahlberg.B., 2001, Fatal Interactions Between Scots Pine and *Sphagnum* Mosses in Bog Ecosystems, *OIKOS,* Vol.94., pp.425-432.

359. Ojala.A.E.K., Saarinen.T., 2002, Palaeosecular Variation of the Earth's Magnetic Field During the Last 10,000 Years Based on the Annually *Laminated Sediment of Lake Nautajarvi, Central Finland, The Holocene*, Vol.12., pp.391-400.

360. Okland.R.H., 1990, A Phytoecological Study of the Mire Northern Kisselbergmosen, SE Norway, II, Identification of Gradients by Detrended (Canonical) Correspondence Analysis, *Nordic Journal of Botany*, Vol.10., pp.79-108.

361. Okland.R.H., Okland.T., Rydgren.K., 2001, A Scandinavian Perspective on Ecological Gradients in Northwest European Mires: Reply to Wheeler and Proctor, *Journal of Ecology*, Vol.89., pp.481-486.

362. Oldfield.F., Alversen.K., 2003, The Societal Relevance of Palenvironmental Research, *Paleoclimate Global Change and the Future,* Springer, 1-11.

363. Olsen.J., Anderson.N.J., Knudsen.F., 2012, Variability of the North Atlantic Oscillation over the Past 5,200 Years, *Nature Geoscience*, Vol.5., pp.808-812.

364. Óskarsson,N., 1980, The Interaction between Volcanic Gases and Tephra: Fluorine Adhering to Tephra of the 1970 Hekla Eruption, *Journal of Volcanology and Geothermal Research*, Vol.8., pp.251-266.

365. Patterson.G., Anderson.R., 2000, *Forests and Peatland Habitats,* Forestry Commission.

366. Peacock.J.D., Harkness.D.D., 1990, Radiocarbon Ages and the Full-Glacial to Holocene Transition in Seas Adjacent to Scotland and Southern Scandinavia: A Review, Transactions of the Royal Society of Edinburgh, *Earth Sciences*, Vol.81., pp.385-396.

367. Pears.N.V., 1972, Interpretation Problems in the Study of Tree-Line Fluctuations, In: Taylor.J.A., (ed) *Research Papers of Forest Meteorology*, An Aberystwyth Symposium, The Cambridge News Ltd, Aberystwyth, pp.31-45.

368. Pearson.G.W., and Stuiver.M., 1986, High-Precision Calibration of the Radiocarbon Time Scale, 5000-2500BC, *Radiocarbon*, Vol.28., pp.839-862.

369. Pennington.W., 1975, A Chronostratigraphic Comparison of Late-Weichselian and Late Devensian Sub-Divisions, Illustrated by Two Radiocarbon-Dated Profiles from Western Britain, *Boreas*, Vol.4., pp.157-171.

370. Pennington.W., Haworth.E.Y., Bonny.A.P., Lishman.J.P., 1972, Lake Sediments in Northern Scotland, *Philosophical Transactions of the Royal Society of London*, Vol.264, pp.191-294.

371. Peteet.D.M., 1995, Global Younger Dryas?, *Quaternary International*, Vol.28., pp.93-104.

372. Petit.J.R., Jouzel.J., Raynaud.D., Barkov.N.I., Barnola.J.M., Basile.I., Benders.M., Chappellaz.J., Davis.M., Delaygue.G., Delmotte.M., Kotlyakov.V.M., Legrand.M., Lipenkov.V.Y., Lorius.C., Pepin.L., Ritz.C., Saltzman.E., Stievenard.M., 1999, Climate and Atmospheric History of the Past 420,000 Years from the Vostok Ice Core, *Antarctica, Nature,* Vol.399., pp.429-436.

373. Piggott.S., 1972, A Note on Climatic Deterioration in the 1st Millennium BC in Britian, *Scottish Archaeological Forum*, Vol.4., pp.109-113.

374. Pilcher.J.R., Hall.V.A., 1992, Towards a Tephrochronology for the Holocene of the North of Ireland, *The Holocene,* Vol.2., pp.255-259.

375. Porter. S.C., Denton.G.H., 1967, Chronology of Neoglaciation in the North American Cordillera, *American Journal of Science*, Vol.265., pp.177-210.

376. Porter.S.C., 1981, Recent Glacier Variations and Volcanic Eruptions, *Nature*, Vol.291., pp.134-142.

377. Rackham.O., 1986, *The History of the Countryside*, J.M.Dent, London.

378. Ratcliffe.D.A., 1977, *A Nature Conservation Review*, Vol.2., Cambridge, Cambridge University Press, pp.401 and 320.

379. Reed.J.M., Stevenson.A.C., Juggins.S., 2001, A Multi-Proxy Record of Holocene Climatic Change in Southwestern Spain: the Laguna Medina, Cadiz, *Holocene*, Vol.11., pp.707-719.

380. Reimer.P.J., Baillie.M.G.L., Bard.E., Bayliss.A., Beck.J.W., Blackwell.P.G., Buck.C.E., Burr.G.S., Cutler.K.B., Damon.P.E., Edwards.R.L., Fairbanks.R.G., Friedrich.M., Guilderson.T.P., Herring.C., Hughen.K.A., Kromer.K.A., McCormac.F.G., Manning.S.W., Ramsey.C.B., Reimer.R.W., Remmele.S., Southon.J.R., Stuiver.M., Talamo.S., Taylor.F.W., Van Der Plicht.J., Weyhenmeyer.C.E., 2004 IntCal04 Terrestrial Radiocarbon Age Calibration, 0-26 cal kyr BP, *Radiocarbon*, Vol.46., pp.1029-1058.

381. Renssen.H., Seppa.H., Heiri.O., Roche.D.M., Goosse.H., Fichefet.T., 2009, The Spatial and Temporal Complexity of the Holocene Thermal Maximum, *Nature Geoscience*, Vol.2., pp.411-414.

382. Renssen.H., Van Geel.B., Van Der Plicht.J., Magny.M., 2000, Reduced Solar Activity as a Trigger for the Start of the Younger Dryas? *Quaternary International*, Vol.68-71, pp.373-383.

383. Richards.M.P., Hedges.R.E.M., 1999a, A Neolithic Revolution? New Evidence of Diet in the British Neolithic, *Antiquity*, Vol.73., pp.891-897.

384. Richards.M.P., Hedges.R.E.M., 1999b, Stable Isotope Evidence for Similarities in the Types of Marine Foods used by Late Mesolithic Humans on the Atlantic Coast of Europe, *Journal of Archaeological Science*, Vol.26., pp.717-722.

385. Riezebos.P.A., Slotboom.R.T., 1984, Three-Fold Subdivision of the Allerød Chronozone, *Boreas,* Vol.13., pp.347-353.

386. Roberts.N., 1998, *The Holocene An Environmental History,* 2nd Edition, Oxford, Blackwell, pp.316.

387. Roberts.N., Reed.J.M., Leng.M.J., Kuzucuoglu.C., Fontugne.M., Bertaux.J., Woldring.H., Bottema.S., Black.S., Hunt.E., Karabiyikoglu.M., 2001, The Tempo of Holocene Climatic Change in the Eastern Mediterranean Region; New High Resolution Crater-Lake Sediment Data from Central Turkey, *Holocene,* Vol.11., pp.721-736.

388. Robinson.D.E., 1987, Invetsigations into the Aukhorn Peat Mounds, Keiss, Caithness: Pollen, Plant Macrofossil and Charcoal Analyses, *New Phytologist*, Vol.106., pp.185-200.

389. Robinson.D.E., Dickson.J.H., 1988, Vegetational History and Land Use: A Radiocarbon-Dated Pollen Diagram from Machrie Moor, Arran, Scotland, *New Phytologist*, Vol.109., pp.223-251.

390. Robinson.M., Ballantyne.C.K., 1979, Evidence for the Glacial Advance Predating the Loch Lomond Advance in Wester Ross, *Scottish Journal of Geology*, Vol.15., pp.271-277.

391. Roca.J.R., Julia.R., 1997, Late Glacial and Holocene Lacustrine Evolution Based on Ostracode Assemblages on Southeastern Spain, *Giobios*, Vol.30., pp.823-830.

392. Rose.W.I., 1977, Scavenging of Volcanic Aerosol by Ash: Atmospheric and Volcanological Implications, *Geology*, Vol.5., pp.621-624.

393. Rosen.P., Segerstrom.U., Eriksson.L., Renberg.I., Birks.H.J.B., 2001, Holocene Climatic Change Reconstructed from Diatoms, Chironomids, Pollen and Near-Infrared Spectroscopy at an Alpine Lake (Sjuodjijaure) in Northern Sweden, *The Holocene*, Vol.11., pp.551-562.

394. Rowley-Conwy.P., 1985, The Origin of Agriculture in Denmark: A Review of some Theories, Journal of Danish Archaeology, Vol.4., pp.188-195.

395. Roy.M., 1997, The Highlands and Islands of Scotland, In: *Regional Climates of the British Isles*, (eds) Wheeler.D., Mayes.J., Routledge, London, pp.228-253.

396. RSPB Scotland, 2011, *Realising the Benefits of Peatlands: Overcoming Policy Barriers to Peatland Restoration,* pp.3-18.

397. Ruddiman.W.F., 1977, Late Quaternary Surface Ocean Kinematics and Climatic Change in the High Latitude North Atlantic, *Journal of Geophysical Research,* Vol.82., pp.3877-3887.

398. Ruddiman.W.F., McIntyre.A., 1981, The Mode and Mechanism of the Last Deglaciation: Oceanic Evidence, *Quaternary Research,* Vol.16., pp.125-134.

399. Ruddiman.W.F., Sancetta.C.D., McIntyre.A., 1977, Glacial/Interglacial Response Rate of Subpolar North Atlantic Waters to Climatic Change: The Record in Oceanic Sediments, *Philosophical Transactions of the Royal Society of America Memoir*, Vol.145., pp.111-145.

400. Schulting.R.J., 1998, Slighting the Sea: Stable Isotope Evidence for the Transition to Framing in Northwestern Europe, *Documenta Praehistorica,* Vol.15., pp.203-218.

401. Schulting.R.J., Richards.M.P., 2002, The Wet, the Wild and the Domesticated: the Mesolithic-Neolithic Transition on the West Coast of Scotland*, Journal of European Archaeology*, Vol.5., pp.147-189.

402. Sear.C.B., Kelly.P.M., Jones.P.D., Goodess.C.M., 1987., Global Surface Temperature Responses to Major Volcanic Eruptions, *Nature,* Vol.330., pp.365-367.

403. Selby.K., 1997, *Late Devensian and Holocene Relative Sea Level Changes on the Isle of Skye, Scotland,* Coventry University, PhD Thesis.

404. Selby.K.A., 2004, Lateglacial and Holocene Vegetation Change on the Isle of Skye: New Data from Three Coastal Locations, *Veget Hist Archaeobot*, Vol.13., pp.233-247.

405. Self.S., Rampino.M.R., Barbera.J.J., 1981, The Possible Effects of Large 19th and 20th Century Volcanic Eruptions on Zonal and Hemispheric Surface Temperatures, *Journal of Volcanology and Geothermal* Research, Vol.11., pp.41-60.

406. Seppä.H., Birks.H.J.B., 2001, July Mean Temperature and Annual Precipitation Trends During the Holocene in the Fennoscandian Tree-Line Area: Pollen-Based Climate Reconstructions, *The Holocene*, Vol.11., pp.527-539.

407. Seppa.H., Birks.H.J.B., 2002, Holocene Climate Reconstructions from the Fennoscandian Tree-Line Area Based on Pollen Data from Toskaljavri, *Quaternary Research*, Vol.57., pp.191-199.

408. Shaviv.N.J., 2002, Cosmic Ray Diffusion from the Galactic Spiral Arms, Iron Meteorites, and a Possible Climatic Connection, *Physical Review Letters*, Vol.89., pp.51-102.

409. Sheets.P.D., and Grayson.D.K., 1979, *Volcanic Activity and Human Ecology,* London, Academic Press.

410. Shemesh.A., Rosqvist.G., Rietti-Shati.M., Rubensdotter.L., Bigler.C., Yam.R., Karlen.W., 2001, Holocene Climatic Change in Swedish Lapland Inferred from an Oxygen-Isotope Record of Lacustrine Biogenic Silica, *The Holocene*, Vol.11., pp.447-454.

411. Shennan.I., Innes.J.B., Long.A.J., Zong.Y., 1993, Late Devensian and Holocene Relative Sea-Level Changes at Rumach, Near Arisaig, Northwest Scotland, *Norsk Geologisk Tidsskrift*, Vol.73., pp.161-174.

412. Shotton.F.W., 1977, The Devensian Stage: Its Development, Limits and Sub stages, Philosophical *Transactions of The Royal Society of London*, Vol.280., pp.107-118.

413. Siegenthaler.U., Monnin.E., Kawamura.K., Spahni.R., Schwander.J., Stuaffer.B., Stocker.T.F., Barnola.J.M., Fischer.H., 2005b, Supporting Evidence from the EPICA Dronning Maud Land Ice Core for Atmospheric CO_2 Changes During the Past Millenium, *Tellus,* Vol.57b., pp.51-57.

414. Siegenthaler.U., Stocker.T.F., Monnin.E., Lüthi.D., Schwander.J., Stauffer.B., Raynaud.D., Barnola.J.M., Fischer.H., Masson-Delmotte.V., Jouzel.J., 2005a, Stable Carbon Cycle-Climate Relationship During the Late Pleistocene, *Science,* Vol.310., pp.1313-1317.

415. Sikstrom.U., Jacobson.S., Pettersson.F., Weslien.J., 2011, Crown Transparency, Tree Mortality and Stem Growth of *Pinus sylvestris*, and Colonization of *Tomicus piniperda* after an outbreak of *Gremmeniella abietina, Forest Ecology Management,* Vol.262., pp.2108-2119.

416. Simmons.I.G., Dimbleby.G.W., Grigson.C., 1981, The Mesolithic, In: Simmons.I.G., Tooley.M.J., (eds) *The Environment in British Prehistory*, pp.82-124., Duckworth, London.

417. Sinclair.W.T., Morman.J.D., Ennos.R.A., 1998, Multi Origins for Scots Pine (*Pinus sylvestris L.*) in Scotland: Evidence from Mitochondrial Variation, *Heredity*, Vol.80., pp.233-240.

418. Sissons.J.B., 1967, Glacial Stages and Radiocarbon Dates in Scotland, *Scottish Journal of Geology,* Vol.3., pp.375-381.

419. Sissons.J.B., 1977a, Former Ice-Dammed Lakes of Glen Moriston, Inverness-Shire and their Significance in Upland Britain, *Transactions of the Institute of British Geographers*, Vol.2., pp.224-242.

420. Sissons.J.B., Dawson.A.G., 1981, Former Sea Levels and Ice Limits in Part of Wester Ross, Northwest Scotland, *Proceedings of the Geologists Association,* Vol.92., pp.115-124.

421. Smith.A.G., 1970, *Studies on the Vegetational History of The British Isles,* CUP Archive, Belfast.

422. Smith.A.G., 1970., *The Influence of Mesolithic and Neolithic man on British Vegetation; A Discussion, in: Studies in the Vegetational History of the British Isles,* Walker.D., and West.R.G., Cambridge University Press, pp.81-96.

423. Smith.M.A., 1998, Holocene Regional Vegetation History of the Lairg Area, In: *The Lairg Project 1988-1996: The Evolution of an Archaeological Landscape in Northern Scotland,* (eds) McCullagh.R.P.J., Tipping.R., STAR: Edinburgh, pp.177-199.

424. Smout.T.C., MacDonald.A.R., Watson.F., 2005, *A History of the Native Woodlands of Scotland, 1500-1920,* Edinburgh University Press, Edinburgh.

425. Snowball.I., Sandgren.P., 2002, Geomagnetic Field Variations in Northern Sweden during the Holocene quantified from Varved Lake Sediments and Their Implications for Cosmogenic Nuclide Productijn Rates, *The Holocene*, Vol.12., pp.517-530.

426. Soranzo.N., Alia.R., Provan.J., Powell.W., 2000, Patterns of Variation at a Mitochondrial Sequence-Tagged-Site Locus Provides New Insights Into the Postglacial History of European *Pinus sylvestris* Populations, *Molecular Ecology*, Vol.9., pp.1205-1211.

427. Spahni.R., Chappellaz.J., Stocker.T.F., Loulergue.L., Hausammann.G., Kawamura.K., Flückiger.J., Schwander.J., Raynaud.D., Masson-Delmotte.V., Jouzel.J., 2005, Atmospheric Methane and Nitrous Oxide of the Late Pleistocene from Antarctic Ice Cores, *Science*, Vol.310., pp.1317-1321.

428. Spence.D.H.N., 1964, The Macrophyte Vegetation of Freshwater Lochs, Swamps and Associated Fens, In: *The Vegetation of Scotland*, (eds) Burnet.J.H. pp.306-425, Oliver and Boyd, Edinburgh.

429. Speranza.A.O.M., Van Geel.B., Van Der Plicht.J., 2000, Improving the Time Control of the Subboreal/Subatlantic Transition in a Czech Peat Sequence by ^{14}C Wiggle-Matching, *Quaternary Science Reviews,* Vol.19., pp.1589-1604.

430. Speranza.A.O.M., Van Geel.B., Van Der Plicht.J., 2002, Evidence for Solar Forcing of Climate Change at ca. 850 cal. BC from a Czech Peat Sequence, *Global and Planetary Change,* Vol.35., pp.51-65.

431. Steven.H.M., Carlisle.A., 1959, *The Native Pinewoods of Scotland*, Edinburgh, Oliver and Boyd.

432. Stocklin.J., Korner.C., 1999, Recruitment and Mortality of *Pinus sylvestris* near the Arctic Tree Line: The Role of Climatic Change and Herbivory, *Ecology Bulletin*, Vol.47., pp.168-177.

433. Stuiver..M., Braziunas.T.F., Becker.B., Kromer.B., 1991, Climatic, Solar, Oceanic, and Geomagnetic Influences on Lateglacial and Holocene Atmospheric ^{14}C/^{12}C Change, *Quaternary Research*, Vol.35., pp.1-24.

434. Stuiver.M., and Braziunas.T.F., 1993, *Sun, Ocean, Climate and Atmospheric 14CO2: An Evaluation of Causal and Spectral Relationships, The Holocene,* Vol.3., pp.289-305.

435. Stuiver.M., Reimer.P.J., Bard.E., Beck..J.W., Burr.G.S., Hughen.K.A., Kromer.B., McCormac.F.G., Van Der Plicht.J., Spurk.M., 1998, INTCAL98 Radiocarbon Age Calibration, 4,000-0 cal. BP, *Radiocarbon,* Vol.40., pp.1041-1083.

436. Sugita.S., Macdonald.G.M., Larsen.C.P.S., 1997, Reconstrcution of Fire Disturbance and Forest Succession from Fossil Pollen in Lake Sediments: Potential and Limitations, In: *Sediment Records of Biomass Burning and Global Change,* (eds) Clark.J.S., Cachier.H., Goldammer.J.G., Stock.B.J., Springer-Verlag, Berlin, pp.387-412.

437. Tallantire.P.A., 2002, The Early Holocene Spread of Hazel (*Corylus avellana L.*) in Europe North and West of the Alps: An Ecological Hypothesis, *The Holocene*, Vol.12., pp.81-96.

438. Tansley.A.G., 1949, *The British Islands and their Vegetation*, 2nd Edition, Cambridge.

439. Taylor.K.C., Lamorey.G.W., Doyle.G.A., Alley.R.B., Grootes.P.M., Mayewski.P.A., White.J.W.C., Barlow.L.K., 1993, The 'Flickering Switch' of the Late Pleistocene Climate Change, *Nature*, Vol.361., pp.432-436.

440. Terral.J.F., Mengual.X., 1999, Reconstruction of Holocene Climate in Southern France and Eastern Spain Using Quantitative Anatomy of Olive Wood and Archaeological Charcoal, *Palaeogeography, Palaeoclimatology, Palaeoecology,* Vol.153., pp.71-92.

441. Thomas.J., 1991, *Rethinking the Neolithic,* Cambridge, Cambridge University Press.

442. Thorarinsson.S., 1980, Distant Transport of Tephra in Three Katla Eruptions and One Grimsvötn Eruption, *Jökull,* Vol.30., pp.65-73.

443. Thorarinsson.S., 1981, Greetings from Iceland: Ash-falls and Volcanic Aerosols in Scandinavia, *Geografiska Annaler*, Vol.63., pp.106-118.

444. Thorpe.I.J., 1996, *The Origins of Agriculture in Europe*, London, Routledge.

445. Tipping.R., 1994, The Form and Fate of Scottish Woodlands, *Proc Soc Antiq Scot,* Vol.124., pp.1-54.

446. Tipping.R., 1996, Microscopic Charcaoal Records, Inferred Human Activity and Climate Change in the Mesolithic of Northernmost Scotland, In: *The Early Prehistory of Scotland*, (eds) Pollard.A., Morrison.A., Edinburgh, Edinburgh University Press, pp.39-61.

447. Tipping.R., 1997, Vegetational history of southern Scotland, *Botanical Journal of Scotland,* Vol.2., pp.151-162

448. Tipping.R., 2002, Climatic Variability and 'Marginal' Settlement in Upland British Landscapes: a Re-Evaluation, *Landscapes*, Vol.3., pp.10-28.

449. Tipping.R., 2008, Blanket Peat in the Scottish Highlands: Timing, Cause, Spread and the Myth of Environmental Determinism, *Biodiversity and Conservation,* Vol.17., pp.2097-2113.

450. Tipping.R., Ashmore.P., Davies.A.L., Haggart.A.B., Moir.A., Newton.A., Sands.R., Skinner.T., Tisdall.E., 2008, Prehistoric *Pinus* Woodland Dynamics in an Upland Landscape in Northern Scotland: The Roles of Climate Change and Human Impact, *Vegetation History Archaeobotany,* Vol.17., pp.251-267.

451. Tipping.R., Bunting.M.J., Davies.A.L., Murray.H., Fraser.S., McCulloch.R., 2009. Modellinh Land Use Around and Early Neolithic Timber 'Hall' in Northeast Scotland from High Spatial Resolution Pollen Analyses, *Journal of Archaeological Science,* Vol.36., pp.140-149.

452. Tipping.R., Davies.A., McCulloch.R., Tisdall.E., 2008, Response to Late Bronze Age Climate Change of Farming Communities in North East Scotland, *Journal of Archaeological Science*, Vol.35., pp.2379-2386.

453. Tipping.R., McCulloch.R., 2003, *Vegetation History and Human Impact at Reidchalmai, Little Rogart, Near Golspie, Sutherland,* Archive Report, Historic Scotland

454. Tipping.R., Tisdall.E., 2004, Continuity, Crisis and Climate Change in the Neolithic and Early Bronze Age Periods of Northwest Europe, In: Shepherd, I.A.G., Barclay, G. (eds.), *Scotland in their European Context*, Society of Antiquaries of Scotland, Scotland, pp.71-82.

455. Tipping.R., Tisdall.E., Davies.A., Wilson.C., 2007, Living with Peat in the Flow Country: Prehistoric Farming Communities and Blanket Peat Spread at Oliclett, Caithness, Northern Scotland, in: Barber.J., et al (eds) *Archaeology from the Wetlands: Recent Perspectives*, Society of Antiquaries of Scotland: Edinburgh, pp.165-174.

456. Tipping.R.M., 2003, *The Quaternary of Glen Affric and Kintail, Field Guide*, Quaternary Research Association, London.

457. Tisdall.E.W., 2000, *Holocene Climate Change in Glen Affric, Northern Scotland: A Multiproxy Approach,* Unpublished Ph.D. thesis, University of Sterling.

458. Tisdall.E.W., 2003, Loch Coulavie: Stratigraphic Data on Holocene Lake-Level and Proxy Precipitation Change. In: *The Quaternary of Glen Affric and Kintail, Field Guide*, (eds) Tipping, R. Quaternary Research Association, London, pp.29-40.

459. Tolan-Smith.C., 1998, Radiocarbon Chronology and the Lateglacial and Early Postglacial Resettlement of the British Isles, *Quaternary International*, Vol.49-50., pp.21-27.

460. Troels-Smith.J., 1955, Characterisation of Unconsolidated Sediments, *Danmarks Geologiske Undersogelse IV Series,* Vol.3., pp.1-73.

461. UK Committee on Climate Change Adaptation Sub-Committee, 2011, *How Well is Scotland Preparing for Climate Change? pp.1-58*

462. UK National Statistic, 2011, Mineral Extraction in Great Britain.

463. USEPA. 1999, *Innovative Technology Verification Report: Sediment Sampling Technology, Aquatic Research Instruments Russian Peat Borer*, EPA/600/R-01/010, Office of Research and Development.

464. Valet.J.P., Tric.E., Herrero-Bervera.E., Meynadier.L., Lockwood.J.P., 1998, Absolute Paleointensity from Hawaiian Lavas Younger than 35ka, *Earth and Planetary Science Letters,* Vol.161., pp.19-32.

465. Van Der MoleN.P.C., 1992, *Hummock-Hollow Complexes on Irish Raised Bogs, A Palaeo/Actuo Ecological Approach of Environmental and Climatic Change,* PhD Thesis, University of Amsterdam, pp.213.

466. Van Geel.B., Buurman.J., Waterbolk.H.T., 1996, Archaeological and Palaeoecological Indications of an Abrupt Climate Change in The Netherlands and Evidence for Climatological Teleconnections around 2650BP, *Journal of Quaternary Science*, Vol.11., pp.451-460.

467. Van Geel.B., Buurman.J., Waterbolk.H.T., 1996, Archaeological and Palaeoecological Indications of an Abrupt Climate Change in The Netherlands, and Evidence for Climatological Teleconnections Around 2650BP, *Journal of Quaternary Science*, Vol.11., pp.451-460.

468. Van Geel.B., Raspopov.O.M., Van Der Plicht.J., Renssen.H., 1998b, Solar Forcing of Abrupt Climate Change around 850 Calendar Years BP. In: *Natural Catastrophes during the Bronze Age Civilizations,* (eds) Peiser.B.J., Palmer.T., Bailey.M.E., British Archaeological Reports, Oxford, PP.162-168.

469. Van Geel.B., Van Der Plicht.J., Kilian.M.R., Klaver.E.R., Kouwenberg.J.H.M., Renssen.H., Reynaud-Farrera.I., Waterbolk.H.T., 1998, The Sharp Rise of ^{14}C ca. 800 cal BC: Possible Causes, Related Climatic Teleconnections and the Impact on Human Environments, *Radiocarbon*, Vol.40., pp.535-550.

470. Van Geel.B., Van Der Plicht.J., Kilian.M.R., Klaver.E.R., Kouwenberg.J.H.M., Renssen.H., Reynaud-Ferrara.I., Waterbolk.H.T., 1998a, the Sharp Rise in Atmospheric ^{14}C ca.800 cal.BC: Possible Causes, Related Climatic Teleconnections and the Impact On Human Environments, *Radiocarbon,* Vol.40., pp.535-550.

471. Vasari.Y., 1977, Radiocarbon Dating of Lateglacial and Early Flandrian Vegetational Successions in the Scottish Highlands and Isle of Skye, In: *Studies in the Scottish Lateglacial Environment*, (eds) Gray.M., Lowe.J.J., Pergamon Press, Oxford, pp.143-162.

472. Vasari.Y., Vasari.A., 1968, Late and Postglacial Macrophytic Vegetation in the Lochs of Northern Scotland, *Acta Botanica Fennica*, Vol.80., pp.1-20.

473. Vega Leinert.A.C., Jones.D.H., Wells.J., Smith.D.E., 2000, Mid-Holocene Environmental Changes in the Bay of Kaill, Mainland Orkney, Scotland: An Integrated Geomorphological, Sedimentological and Stratigraphical Study, *Journal of Quaternary Science*, Vol.15., pp.509-528.

474. Vidakovic.M., 1991, *Conifers – Morphology and Variation*, Graficki Zavod Hrvatske, Zagreb.

475. Von Post.L., 1946, The Prospect for Pollen Analysis in the Study of the Earths Climatic History*, New Phytologist,* Vol.45., pp.193-217.

476. Walker.M., 1973, *Archaeological Excavation of a Microlithic Assemblage at* Shieldaig*, Wester Ross, Scotland,* Preliminary Report, Unpublished Report, Edinburgh.

477. Walker.M.J.C., 1984, Pollen Analysis and Quaternary Research in Scotland, *Quaternary Science Reviews,* Vol.3., pp.369-404.

478. Walker.M.J.C., and Lowe.J.J., 1990, Reconstructing the Environmental History of the Last Glacial-Interglacial Transition: Evidence from the Isle of Skye, Inner Hebrides, Scotland*, Quaternary Science Reviews,* Vol.9., pp.15-49.

479. Walker.M.J.C., Ballantyne.C.K., Lowe.J.J., Sutherland.D.G., 1988, A Reinterpretation of the Lateglacial Environmental History of the Isle of Skye, Inner Hebrides, Scotland, *Journal of Quaternary Science*, Vol.3., pp.135-146.

480. Walker.M.J.C., Lowe.J.J., 1981, Postglacial Environmental History of Rannoch Moor, Scotland, III, Early to Mid-Flandrian Pollen Stratigraphic Data from Sites on Western Rannoch Moor and Near Fort William, *Journal of Biogeography,* Vol.8., pp.475-491.

481. Walker.M.J.C., Lowe.J.J., 1987, Flandrian Environmental History of the Isle of Mull, Scotland III, A High Resolution Pollen Profile from Gribun, Western Mull, *New Phytologist*, Vol.106., pp.333-347.

482. Wanner.H., Solomina.O., Grosjean.M., Ritz.S.P., Jetel.M., 2011, Structure and Origin of Holocene Cold Events, *Quaternary Science Reviews*, Vol.30., pp.3109-3123.

483. Watts.W.A., 1961, Post Atlantic Forests in Ireland, *Proceedings of the Linnean Society,* Vol.172., pp.33-38.

484. Wentworth.R., 1989, *Survey, Discovery and Excavation in Scotland,* Council for Scottish Archaeology, pp.33-34.

485. Wheeler.B.D., and Proctor.M.C.F., 2000, Ecological Gradients, Subdivisions and Terminology of Northwest European Mires, *Journal of Ecology,* Vol.88., pp.187-203.

486. Whittington.G., Edwards.K.J., Cundill.P.R., 1990, Palaeoenvironmental Investigations at Black Loch in the Ochil Hills of Fife, Scotland, O'Dell Memorial Monograph, Vol.22, pp.64-69.

487. Whittington.G., Edwards.K.J., Cundill.P.R., 1991, Late and Post Glacial Vegetational Change at Black Loch, Fife, Eastern Scotland – A Multiple Core Approach, *New Phytologist*, Vol.118., pp.47-166.

488. Whittle.A., 1999, *The Neolithic Period, c. 4000-2500/2200BC,* In: Hunter.J., Ralston.I., (eds) The Archaeology of Britain, pp.58-7.

489. Wickham-Jones.C.R., 1990, *Rhum: Mesolithic and Later Sites at Kinloch, Excavations 1984-86,* Edinburgh, Society of Antiquities of Scotland.

490. Wickham-Jones.C.R., Firth.C.R., 2000, Mesolithic Settlement of Northern Scotland: First Results of Fieldwork in Caithness and Orkney, In: *Mesolithic Lifeways: Current Research from Britain and Ireland,* (eds) Young.R., Leichester, Leicester University Archaeology Monographs Vol.7., pp.119-132.

491. Wicks.K., Mithen.S., 2014, The Impact of the Abrupt 8.2ka Cold Event on the Mesolithic Population of Western Scotland: A Bayesian Chronological Analysis using 'Activity Events' as a Population Proxy, *Journal of Archaeological Science*, Vol.45., pp.240-269.

492. Wijmstra.T.A., Hoekstra.S., De Vries.B.J., Van Der Hammen.T., 1984, Preliminary Study of Periodicities in Percentage Curves Dated by Pollen Density, *Acta Botanica Neerlandica*, Vol.33., pp.547-557.

493. Williams.E., 1989, Dating the Introduction of Food Production into Britain and Ireland, *Antiquity*, Vol.63., pp.510-521.

494. Willis.K.J., Bennett.K.D., 1994, The Neolithic Transition – Fact or Fiction? Palaeoecological Evidence from the Balkans, *The Holocene*, Vol.4., pp.326-330.

495. Wilson.B.F., 1975, Distribution of Secondary Thickening in Tree Root Systems, In: *The Development and Function of Roots*, (eds) Torrey.J.G., Clarkson.D.T., Academic Press, London, pp.197-219.

496. Wisniewski.J., 1982, The Potential Acidity Associated with Dews, Frosts and Fogs., *Water, Air, and Soil Pollution,* Vol.17., pp.361-377.

497. Woodbridge.J., Fyfe.R.M., Roberts.N., 2012, The Impact of the Neolithic Agricultural Transition in Britian: A Comparison of Pollen-Based Land-Cover and Archaeological ^{14}C Date-Inferred Population Change, *Journal of Archaeological Science*.

498. Yang.S., Odah.H., Shaw.J., 2000, Variations in the Geomagnetic Dipole Moment Over the Last 12,000 Years, *Geophysical Journal International*, Vol.140., pp.158-162.