A Chironomid Based Palaeoecological Investigation of Late Glacial Climate Change in North Wales

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Abstract

There is a need to improve our understanding of past climate change as past changes offer considerable insight into the functioning of the climate system. Although extensive, reconstructions of the last deglaciation in north-west Europe (10-17ka yrs. B.P.) have not fully resolved the nature of the short fluctuations that characterised this period. Non-biting midges (Chironomids) have been shown to be effective in recreating summer temperatures, and so a Chironomid based temperature reconstruction for a late-glacial lacustrine sequence from North Wales is presented and compared with existing reconstructions for the British Isles. An existing temperature transfer function has been applied to the results and they are evaluated against existing evidence of late-glacial climate. The climate throughout the late glacial is unstable, and at least one cooling episode during the Windermere Interstadial can be identified. The results from this study are used to reconstruct precipitation at a nearby site once supporting a Younger Dryas cirque glacier, which is in general agreement with existing reconstructions, reinforcing the findings of others who argue that the Younger Dryas in Wales was characterised by high annual rainfall, comparable with modern precipitation. The study concludes that late-glacial climate change in the British Isles is likely to be considerably more complicated than previously found, and it seems likely that there are close parallels with the Greenland ice core and continental terrestrial records, and that seasonality is likely to have played an important role in the nature of this change.

Acknowledgments

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Chapter 1: Introduction to The Last Glaciation in north-west Europe

1.1. Glacial/interglacial cycles

Throughout the Quaternary (2.6-1.8m yrs. B.P. - present) there have been regular switches between periods with extensive global glacier ice and cooler temperatures, termed "glacials", and periods with significantly less ice cover and higher global surface temperatures, termed "interglacials" (Benn & Evans, 1998). In the British Isles the last glacial period, known as the Devensian (Mitchell *et al.*, 1973) lasted from 110-12k yrs. B.P., with the maximum extent of ice growth between 27-20k ^{14}C yrs. B.P. (Fretwell *et al.,* 2007*)* when there was extensive ice cover over much of Northern Europe and North America, referred to collectively as the Palaeoartic. The main ice sheet over the British Isles flowed southward from the north of Ireland, south Scotland and northern England across the British Midlands and much of Wales, covering the Arctic Sea and the British Isles as far south as London (Benn & Evans, 1998). An ice dome centred over the higher reliefs of North Wales varied between a state of conjunction and independence from the main ice flow, although there is considerable debate regarding the size, centre and direction of flow from this ice dome (McCarrol & Ballantyne, 2000).

This dissertation is concerned with the termination of the Devensian glaciation and the transition to the current interglacial period of the present day, which started around 10.2k yrs. B.P. (Taylor *et al.*, 1997). As temperatures increased, driven by long term variations in the Earths orbit, the ice melted between 17-13k yrs. B.P. (Woodcock & Stracken, 2000), and England and Wales were completely ice free by 13k yrs. B.P., although Bradwell *et al.*, (2008) argue for the existence of glaciers in Scotland at this time. This period of glacier retreat across the Palaeoartic, or "Late-Glacial period" was not linear and was characterised by unstable and rapidly fluctuating temperatures, as described in figure 1.1 and in detail in the following sections.

1.2 The last deglaciation

Immediately following deglaciation in the British Isles warmer conditions allowed the development of open scrubland, followed by open woodland characterised by *Juniperus (*Juniper) and *Salix* (Willow). In the British Isles this warm period is termed the Windermere Interstadial. Although warm these conditions were variable, and in parts of Europe at least one cold fluctuation termed the Older Dryas has been identified (Lowe & Walker, 1997). Johnsen *et al.* (1992) delineate no less than two large deteriorations in climate during the interstadial in ice cores from Greenland. This warm period ended abruptly with the onset of a sharp deterioration of climate that caused the reappearance of valley glaciation in

Figure 1.1. A schematic of the temperature changes through the Late Glacial. Modified from Walker, 1995 & Johnsen *et al.*, 1993. The blue lines denote the timescale of this study. Note the difference between radiocarbon and calendar age dates.
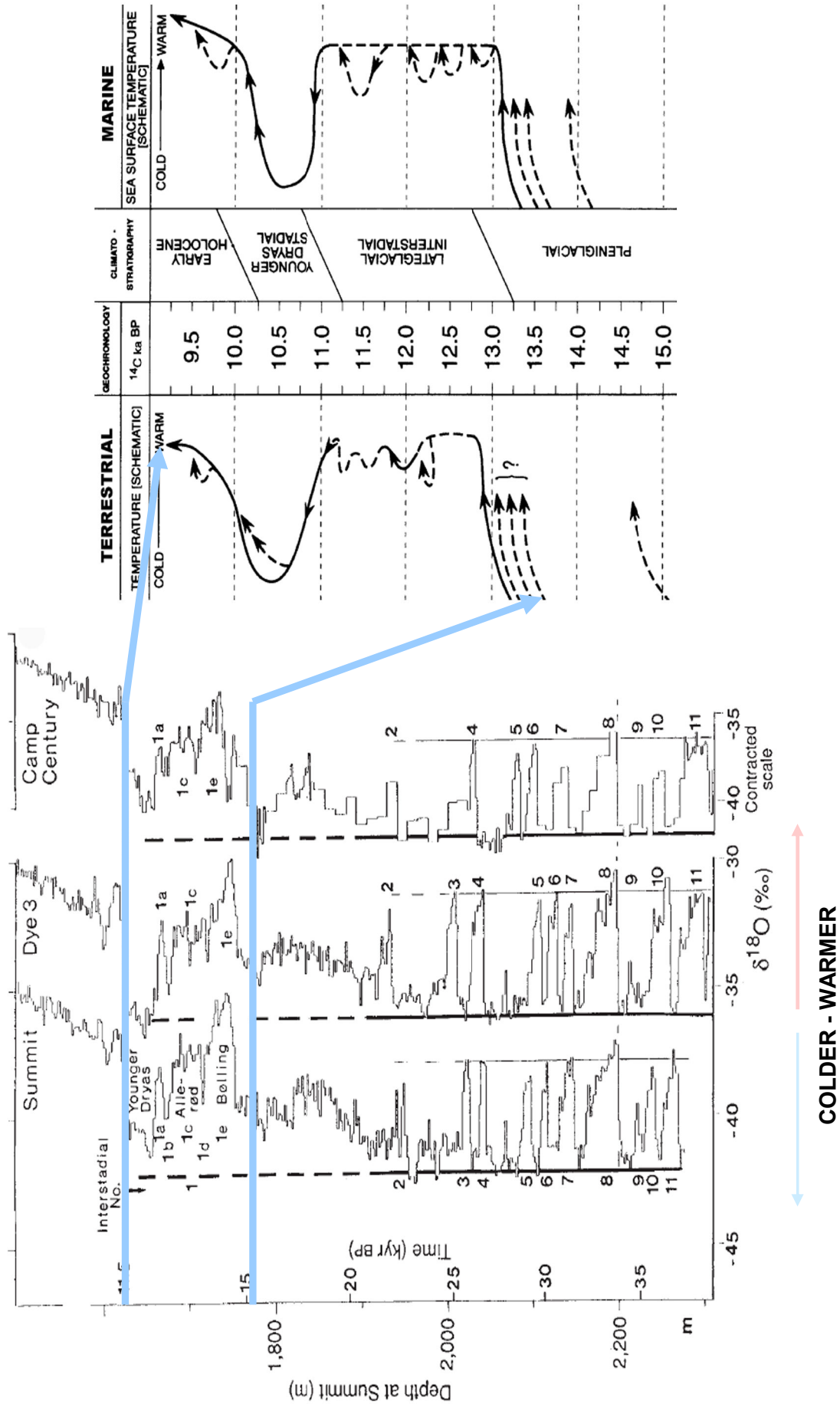
Stratigraphic Unit					Date (Illustrative)
General term	British Isles	Ireland	NW Europe	Greenland	
Early Holocene	Early Holocene	Early Holocene	Early Holocene	GI-0	10k-present
Younger Dryas stadial	Loch Lomond stadial	Woodgrange Interstadial	Younger Dryas	GS-1	11-10k
Late Glacial inter-stadial	Windermere Interstadial	Nahanagan stadial	Bolling oscillation	GI-1a GI-1b GI-1c GI-1d GI-1e	13-11k
			Older Dryas		
			Allerod oscillation		
Pleniglacial	Devensian termination	Devenisan termination	Wiechelian Termination/ pleniglacial	GS-2	17-13k

Figure 1.2. Stratigraphic units for the last glacial-interglacial transition. Note the abbreviations GS for "Greenland stadial number" and GI for "Greenland interstadial number". After Lowe & Walker (1997) &

the British Isles and parts of north-west Europe for around 1k yrs. (Brauer *et al.*, 2000); this deterioration is termed the Loch Lomond Re-advance in Britain (Benn & Evans, 1998). The climate amelioration that followed the Loch Lomond Re-advance marked the start of the Holocene, the interstadial that runs to the present day (Roberts, 1998). Regional nomenclature differs and a brief guide is presented in figure 1.2.

Some of the best evidence of this period of deglaciation comes from oxygen isotope and other geo-chemical evidence from the Greenland ice cores, which clearly show the characteristic course of initially warm temperatures immediately after deglaciation, followed by a slow decline in temperatures, culminating in a prolonged period of cold temperatures (e.g. Johnsen *et al.*, 1992; see figure 1.3). This sequence has been recognised from studies in Greenland, the British Isles, Western Europe and North America, and there is evidence for a similar oscillations in climate in a number of other locations throughout the Northern Hemisphere and possibly the globe, although the latter two are the subjects of considerable debate (see Roberts, 1998).

Björk *et al.* (1998a) have advocated the use of the GRIP ice core record as the type-sequence for the Late Glacial in north-west Europe, but although most north-west European records agree on the key events present in the GRIP record there is significant regional variation and ambiguities in European and the British terrestrial records when they are compared, particularly with regard to the exact timing

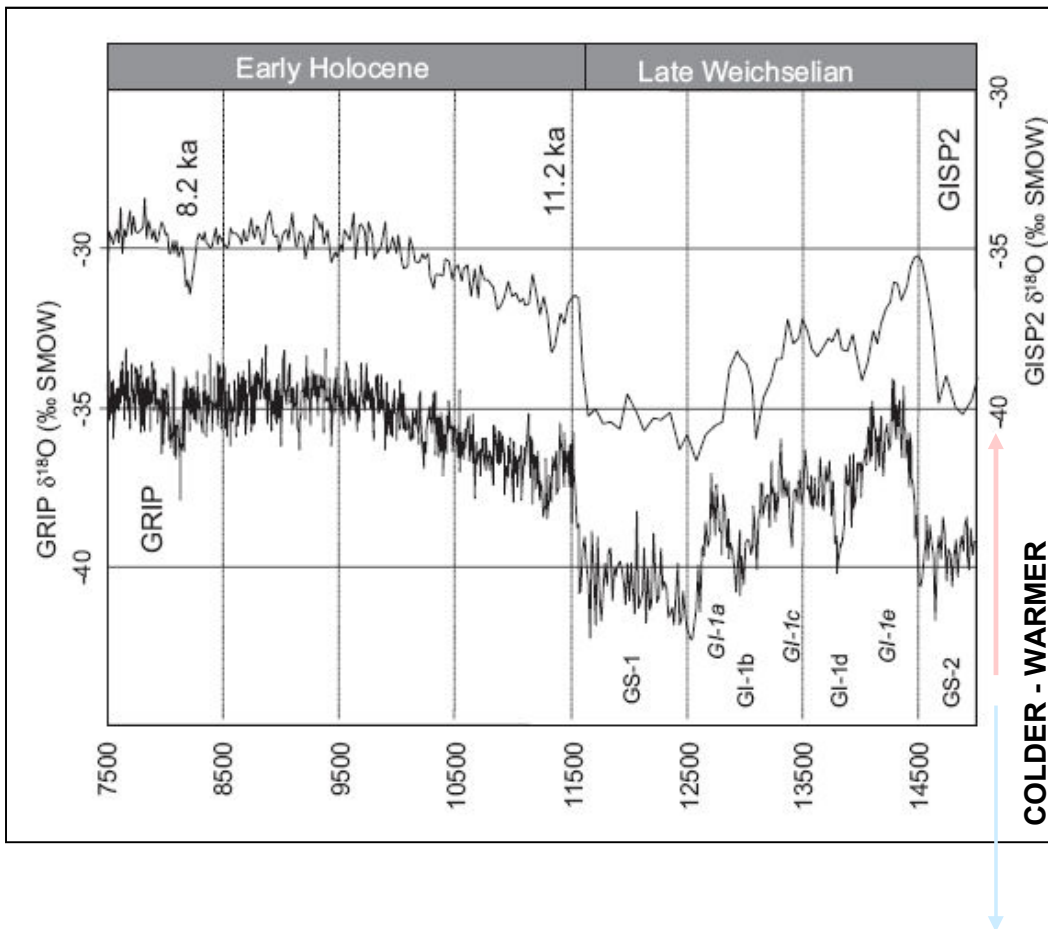*Fig. 1.4: Movements of the polar front through the Late-Glacial. From Tolan-Smith (1998).*

Fig. 1.3: Oxygen isotope records for the Greenland GISP2 core. Shifts to the left indicate colder conditions & vice versa. The stadials and inter-stadials are labelled. From Hoek & Bos (2007).
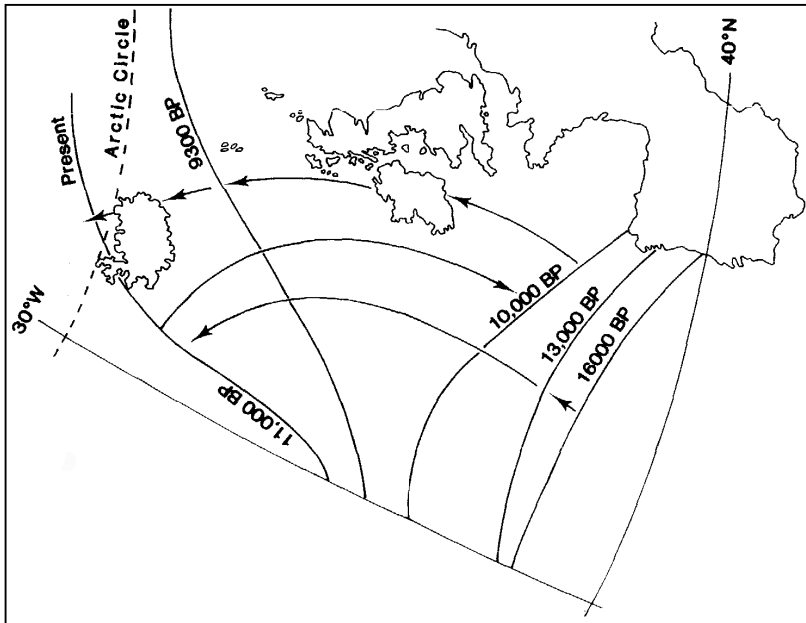
and extent of smaller oscillations in climate present throughout the GRIP record.

There are also regional variations in the response of vegetation and other biota to changes during the Late Glacial (Walker *et al.,* 1994). As a result of these differences the specific details of conditions during the Windermere Interstadial and Loch Lomond Re-advance are the subject of considerable research, debate and disagreement. The following sections outline some of the differences and disagreements between sources and identify areas of the literature that would benefit from further study. It concludes by introducing the importance of understanding how climate changed throughout the Late Glacial, and introduces recent techniques that have opened up new possibilities to improve the record.

1.3 The Late Glacial Interstadial

Across north-west Europe there is clear evidence of a return to warm conditions following the retreat of glaciation to more northerly latitudes in the region. These changes were driven by the northerly shift of the Polar Front as ice receded (see figure 1.4). Across Great Britain the initial an open-scrub habitat was succeeded by a *Juniperus*, *Salix* and *Pinus* open woodland habitat through the Windermere Interstadial (Sparks & West, 1972). In north-west Europe the flora changed from a sparse steppe habitat to a forest of *Betula* and *Salix*, followed by *Pinus*, although the period was punctuated by a short lived return to an open steppe habitat (Brauer *et al.,* 2000) at least once, and possibly twice. The expression of these short returns to cold conditions (termed the Older and Oldest Dryas in Europe) is debated in records from the British Isles, where some workers find evidence for a single fluctuation (e.g. Burrows, 1974; Crabtree, 1972) whilst others argue that no ecotonal boundaries were crossed during the interstadial (e.g. Seddon, 1962; Simpkins, 1974). However much of this work was based on pollen studies and vegetation response is known to lag behind climate change; this may explain the suppressed expression of climate changes in these records. Some workers have used assemblages of Coleoptera (beetles), which are highly responsive to temperature change, to reconstruct past temperatures, but these have not fully resolved the events during the interstadial, with different sites yielding different numbers of events of varying magnitude through the interstadial (e.g. Atkinson *et al.*, 1987).

More recent work with Chironomid (non-biting midge) based proxies has suggested that the interstadial in the British Isles is even more complex, with some studies showing two main cooling events (e.g. Walker *et al.*, 2003; Bedford *et al.*, 2004) and others up to three (e.g. Brooks & Birks, 2000). None have been positively correlated with the north-west European stratigraphy, although some are superficially similar. Some reconstructions do show some affinity with the Greenland ice core stratigraphy

(Mayle *et al.*, 1999) they are far from analogous, and there is no clear indication that changes in the British Isles follow the Greenland or European stratigraphy closely.

The Greenland ice cores suggest an early peak in temperature, followed by a stepped decline punctuated by three short, rapid cooling events (Johnsen *et al.*, 1992). Although the Greenland ice core record provides arguably the most detailed record of climate change through the Late Glacial it lacks the scope to infer the nature of change elsewhere, which must be resolved with further use of in-situ terrestrial proxy records. Given that the European Older Dryas has been correlated with a cooling event in the Greenland ice cores (Johnsen *et al.*, 1992) it would seem likely that this cooling also occurred in the British Isles, but as yet no conclusive timeline of events during the Interstadial in the British Isles can be agreed (Mayle *et al.*, 1999).

These disagreements could represent a regional difference in the expression of climate changes through the interstadial, or it could be the case that the changes were very similar but that our interpretations are limited by the quality of our terrestrial records of change. Further research is needed to clarify the timing, magnitude and geographical extent of these changes in order to understand the nature of change during the interstadial. This is because all proxy records have a certain amount of noise and it is only by combining, comparing and expanding the literature that any synchronous, wide spread climatic events can be identified.

1.4 The Late Glacial Stadial in Europe

The Younger Dryas was a ~1,000yr period of rapid cooling that followed the warming of the Windermere Interstadial. The existence of the Younger Dryas in Europe "has been known for most of the century" (Anderson, 1997 p230), and the expansion and diversification of palaeoecological records has revealed it may well be hemispheric or global in nature (Peteet, 1995), although the wealth of evidence comes from the north Atlantic region. The cooling was large enough to cause glacial advances over much of northern Europe and North America, and there is evidence of the effects of cooling on a hemispheric and possibly global scale (Roberts, 1998). The period is important because it represents an interruption of deglaciation, and does not fit with the Milankovitch cycles that are dominant in controlling the glacial-interglacial cycle (discussed fully in chapter 2). Rather, it seems to represent instability in the climate system at this time, although the causes of the perturbation are heavily debated.

Although a number of climatic deteriorations have been identified during the Late Glacial, the perturbation is unique within the late quaternary in both its magnitude, with mean annual temperature reductions of ~7-8°c and its rapid onset, in the order of 30-200 cal. yrs. (Alley *et al.*, 1993). This has led to it being described as the "canonical example of abrupt climate change" (Carlson *et al.*, 2007, p. 6556), and the focus of much research within the community.

1.4.1 Records of the Stadial

The onset of the Younger Dryas was very rapid, with the reversal completed in the order of ~50 yrs. (Alley *et al.*, 1993), although other proxy records tend to show a slower response in the order of 50-200 years. Although the Younger Dryas is conventionally dated to between 11-10k ^{14}C yrs. B.P. (Bell & Walker, 2005), the identification of ^{14}C dating plateaux, or systemic bias around these dates means they are difficult to calibrate to better than a few hundred calendar years (Lowell & Kelly, 2008; Reimer, 2004). Precisely dating the Younger Dryas has important implications if geographically separate records are found not to be synchronous; it could indicate lags in the chosen proxies response, the change in climate in an area or even question theories that attempt to explain the Younger Dryas, which are critically reliant on accurate dating (e.g. Knutti *et al.*, 2004). The exceptionally well constrained dates from the NGRIP ice core (Rasmussen *et al.*, 2006) give what is probably the best estimation to date and in northern Europe the timing of the onset and termination of the event is remarkably synchronous between records, as shown in figure 1.5.

Record	Technique	Onset	Termination	Reference
GRIP ice core (Greenland)	annual lamination counting w/ geochemical checks	12,940±320 cal. yrs. B.P.	11,640±250 cal. yrs. B.P.	Alley *et al.*, (1993)
GISP2 ice core (Greenland)		12,700±100 cal. yrs. B.P.	11,550±70 cal. yrs. B.P.	Johnsen *et al.*, (1992)
NGRIP ice core (Greenland)		12,846±138 cal. yrs. B.P.	11,653±99 cal. yrs. B.P.	Rasmussen *et al.*, (2006)
[14]C calibrated ages	German tree rings	n/a	11,530±20 yrs. B.P.	Spurk *et al.*, (1998)
Kråkenes Lake, Norway	varve counting	n/a	11,530$^{+40}_{-60}$ cal. yrs. B.P.	Gulliksen *et al.*, 1998
Lake Meerfelder Maar, Germany	varve counting	12,680 cal. yrs. B.P.	11,590 cal. yrs. B.P.	Brauer *et al.*, 2000

Figure 1.5. Dates for the Younger Dryas Period (dates given as years before 1950AD)

In the British Isles there is geomorphological evidence of a re-advance of glacier ice during the Loch Lomond Stadial in Scotland, the English Lake District and North Wales, and other evidence of a widespread increase in the extent and severity of periglacial processes in the British Isles during this time

(Harrison *et al.,* 2008). The pollen record generally agrees that the habitat changed to open-steppe tundra, with discontinuous vegetation cover dominated by *Artemisia* and *Rumex* (Pennington, 1977). Pollen and macrofossil evidence suggest mean summer temperatures in the British Isles were around 10-12°c during the stadial (Isarin & Bohncke, 1999; see figure 1.7), which is in line with geomorphological evidence (e.g. Harrison *et al.*, 2008) and glacier reconstructions (e.g. Hughes, 2002). The terrestrial biological record is particularly useful in tracing the rapid onset of the Loch Lomond Stadial and in reconstructing the exact temperatures; Coleoptera based proxies have indicated mean summer and winter temperatures of ~+10°c and ~-20°c respectively (Atkinson *et al.*, 1987; see figure 1.6). More recent data from Chironomid based proxies in Scotland indicate a similar mean summer temperature (Brooks & Birks, 2000).

One key disagreement in the literature is over the nature of precipitation during the Loch Lomond Stadial; interpretation of past vegetation often suggest low levels of precipitation during the period (Isarin *et al.,* 1998), and this position is supported by recent evidence from relic rock glaciers (Harrison *et al.,* 2008) and some climate models (Renssen *et al.*, 2001). However, evidence from glacier reconstructions suggests that the period must have had a rainfall comparable to modern climate (Grey, 1982; Hughes, 2002). Although the theory of longer winters and correspondingly shorter summers (see Denton *et al.*, 2005) could go some way to settling these differences there is a considerable need for clarification of this important aspect of climate through the stadial. There is evidence of a southerly shift in the course of winter storm tracks (Isarin *et al.*, 1998; Brauer *et al.*, 2008), but it is ambiguous as to whose case, if any, this supports.

Key to the periglacial and glacial inferences of precipitation are a prior knowledge of temperatures, because their models are based on relationships between altitude, temperature and precipitation; if temperature can be constrained, precipitation can be predicted (Sutherland, 1984; Hughes & Braithwaite, 2008). In the British Isles the Coleoptera based record is widely used to constrain palaeo-temperature, although this method may have a fairly low precision because of the wide temperature tolerances of many species. It produces only a possible range of temperatures, and the actual summer and winter values within that range are estimated by comparison with a limited number of modern sites. Coleoptera based reconstructions remain largely untested by comparison with independent alternative proxies, limiting the certainty of any precipitation reconstruction in the British Isles and beyond.

Figure 1.6: Coleoptera-*inferred summer temperatures (T$_{max}$) for the Younger Dryas.* Temperatures are in degrees celcius, and the thick black lines indicate the limits of major glaciation. From Coope *et al.* (1998).
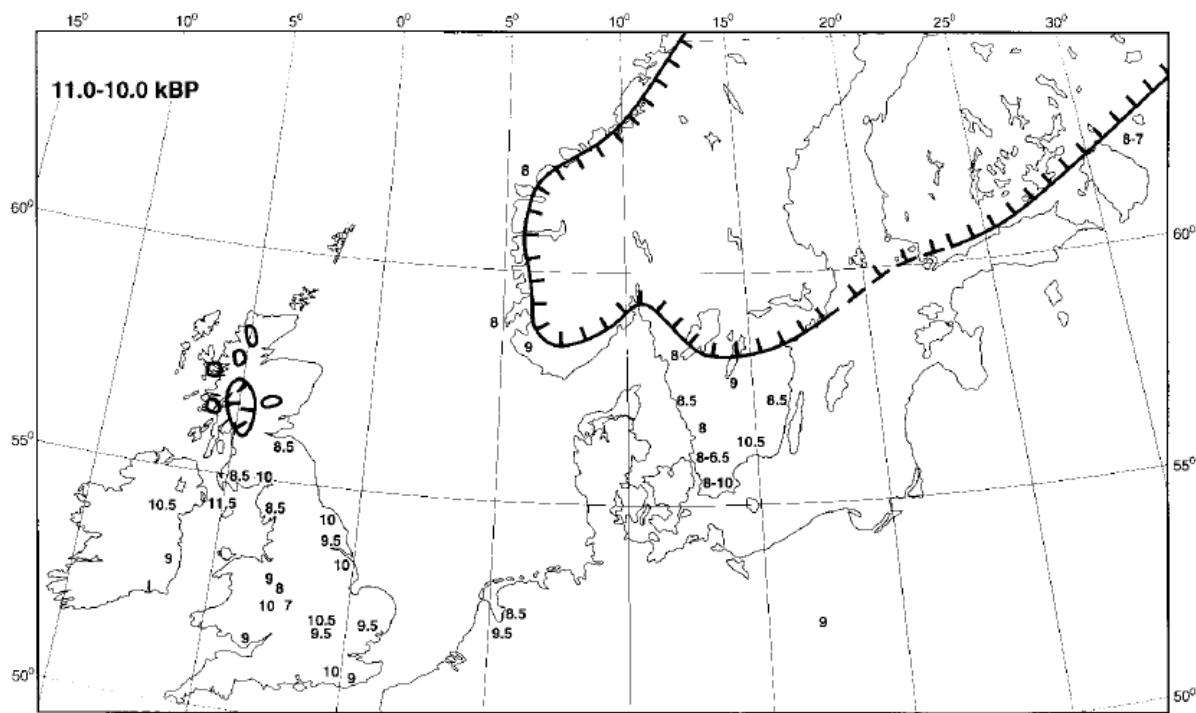

Figure 1.7a: Minimum mean temperature of the warmest month during the Younger Dryas inferred from flora assemblages. From Isarin *et al.* (1998).
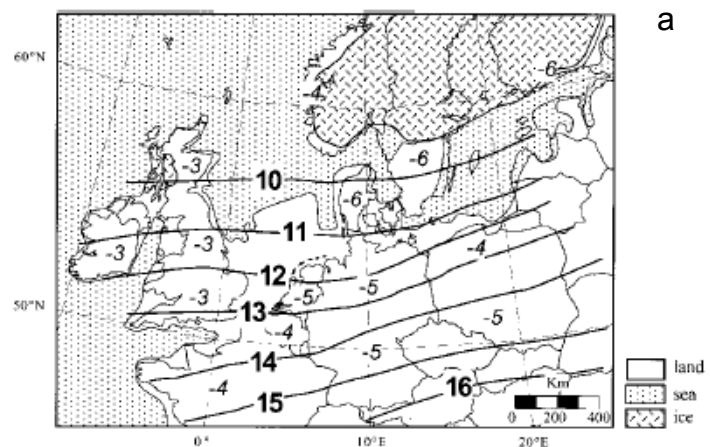
Figure 1.7b: Annual temperature ranges for the Younger Dryas (bold) and difference to the present day (italic). From Isarin *et al.* (1998).
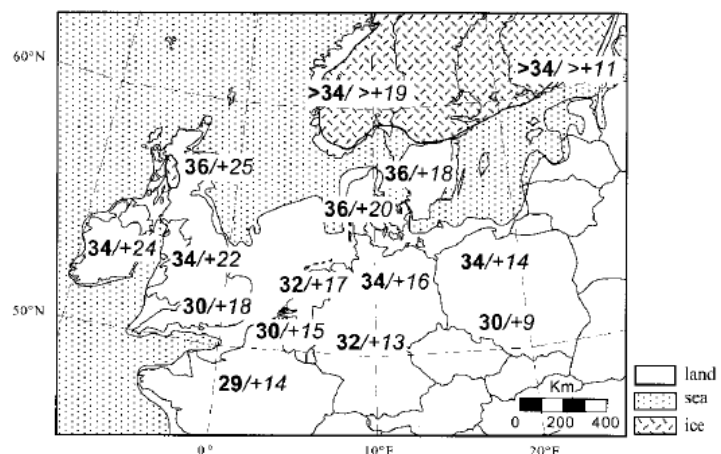
Taking a broader perspective, a similar response is evident across much of Northern Europe, where light demanding, cold tolerant species dominated (Isarin & Bohncke, 1999) to form a treeless habitat, with shrub and herb communities characterised by *Juniper*, *Salix*, *Cyperaceae* (sedges), *Chenopodi-aceae* and *Gramineae* (grasses) in much of west and central Europe (Goslar *et al.*, 1993; Isarin & Bohncke, 1999; Litt & Stebich, 1999), although there may have been more extensive tree cover then previously thought (Kullman, 2002). There were glacial advances in the Alps and other mountainous regions of Europe (e.g. Kerschner *et al.*, 2000), as well as re-advances of the Fennoscandian ice sheet (Bell & Walker, 2005).

There is some evidence for hemispheric response, including evidence of renewed glacier activity in North America that is coeval with the European Younger Dryas (Bell & Walker, 2005), and small cirque glaciers have been found to have developed in Greece (Hughes *et al.*, 2006). Other studies find the effects may have been global, where advances in Patagonian glaciers (Ackert *et al.*, 2008) and high resolution pollen sequences from China showing a reduction in forest cover have been well dated to the same period (Stebich *et al.*, in press). The record of changes in many southern hemisphere regions are either absent or a subject of debate (e.g. New Zealand, see Barrows *et al.*, 2007; Applegate *et al.*, 2008), and suggests that if anything, a change in weather patterns, rather than a temperature depression affected these regions during the stadial.

Alley (2006) has summarises the Younger Dryas as "a cold event across much of the Northern Hemisphere, but especially centered on the northeastern Atlantic, with anomalously dry conditions in broad monsoonal regions of Africa and Asia, tropical atmospheric circulation shifted southward especially in the Americas, and southern warmth." (p247). However the explanations of the cause of such a large perturbation of the climate system are controversial and "investigation of the forcing of such an extreme environmental state offers a new view of climate change" (Mayewski *et al.*, 1993, p197).

1.4.2 Causes of the Stadial

There has recently been a re-energising of debate surrounding the mechanisms that caused the Younger Dryas, and although some workers have proposed some radical new theories the sum conclusions of the new research have been to strengthen the weight of much of the previous literature (Colman, 2007). Berger (1990) outlines three possible broad mechanisms for the Younger Dryas:
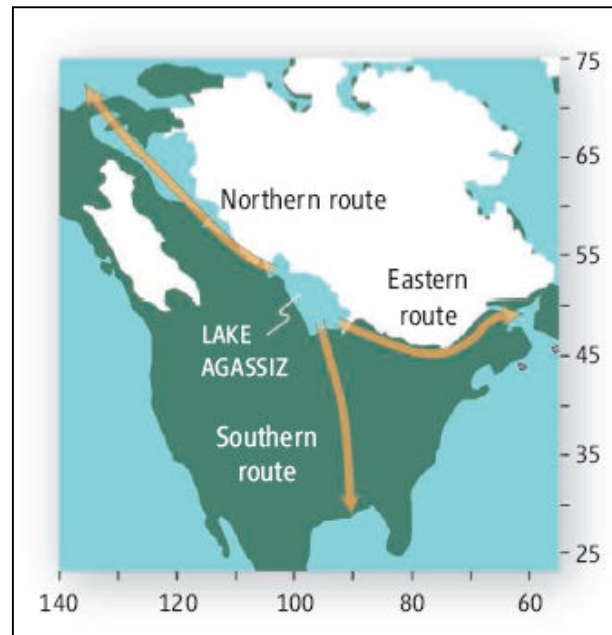
Fig. 1.8: Proposed drainage routes of palaeo-lake Agassiz. From Broeker (2006).

- Positive feedbacks within the climate system (ocean circulation etc.)

- An internal threshold feedback (e.g. collapse of the ice sheets)

- External forcing of the climate system (e.g. volcanism, solar radiation)

The generally accepted theory is that some freshwater perturbation caused a weakening or cessation of the process of North Atlantic Deep Water Formation (NADW), causing a corresponding weakening of the Meridonal Overturning Circulation (MOC, also known as the thermohaline circulation). This led to a southerly shift of the Polar Front because of a weakened receipt of warm North Atlantic Drift (NAD) water from the equatorial regions and caused terrestrial temperature shifts consistent with a reduced ocean surface temperature. This mode of forcing has been implicated in other events, such as the "Finse/8.2ka event" (Alley *et al.*, 1997; Barber *et al.*, 1999). However, there is still no consensus on the origin of the freshwater flux needed to disrupt the NADW, and so a summary of the main proposed theories follows, along with a brief overview of the proposals that do not implicate changes in MOC directly.

The theory that the freshwater source was the palaeo-lake Agassiz in North America has been given much attention by workers as a credible explanation for the disruption to the MOC, and is arguably still the generally accepted theory, if such a theory exists. The lake was glacially-bound (Upham, 1895) and early workers suggested that breaching of the Northern and Eastern margins led to a catastrophic diversion of drainage of the lake via the Hudson Strait into the North Atlantic. Recent geochemical evidence from the Hudson Bay area show a large increase in freshwater drainage to the area at the start

of the Younger Dryas that can be traced using geochemical "fingerprinting" to the proposed Eastern (Hudson Bay) drainage route of Lake Agassiz (Carlson *et al.*, 2007); this provides some of the strongest proof yet for the Agassiz/Eastern route hypothesis (see fig. 1.8).

However there is no geomorphological evidence for any catastrophic drainage from Lake Agassiz at this time (Broecker, 2006), although this could be explained by theory that the drainage occurred within the ice, leaving no traces following deglaciation. The lack of geomorphic evidence for such a large discharge has given rise to alternative theories for the origin of the freshwater pulse. Some workers have suggested alternative drainage routes to the North from palaeo-lake Agassiz into the Arctic (Broecker, 2006), or the source of freshwater was from another glacially bounded lake, for example in the Russian Highlands (Spielhagen *et al.*, 2005). Another potential source of freshwater is from the mobilisation of a large sea of floating ice in the Arctic Ocean (Bradley & England, 2008). Although these propositions are plausible a lack of sedimentary evidence means that these theories remains poorly supported. The Northern drainage hypothesis requires further investigation – although there is baymetric and geomorphological evidence of large scale drainage from Russian highlands (Spielhagen *et al.*, 2005) the nature of this drainage is not understood and the linkages to the North Atlantic remain unproven.

Renssen *et al.* (2000) have suggested that there was a period of reduced solar insolation at the start of the Younger Dryas, and speculate that this caused cooling and subsequent destabilisation of the ice sheets, leading to increased iceberg discharges and changes in the NADW formation, although it is only through an amplifying mechanism like that of the proposed change in NADW formation that the changes in solar insolation are great enough to explain the magnitude of cooling. Renssen *et al.* (2000) also suggest that an increase in cloud cover cooled the earth through increased reflectance, although a sceptic might argue that this does not account for the magnitude of cooling during the Younger Dryas.

Recently Firestone *et al.* (2007) proposed that a previously unknown extraterrestrial impact across North America caused the destabilisation of part of the Laurentide ice sheet which lead to the freshwater perturbation of the North Atlantic, although this has proved controversial and strong arguments against the theory have subsequently been published (e.g. Buchanan *et al.*, 2008).

Although the case for disruption of the NADW/MOC during the Younger Dryas is theoretically sound and well supported by climate modelling and by evidence from the North Atlantic region, the lack of evidence for the origin and/or passage of the massive freshwater flux at best weakens the theory. The stability of the system is crucial to our understanding of its operation, and some workers have suggested it is an inherently unstable system (e.g. Broecker *et al.*, 1985; Praetorius *et al.*, 2008).

Chapter 2: Climate Change Past & Present

2.1 The relevance of the last glacial

The driver of the glacial-interglacial cycles are well explained by orbital forcing, or "Milankovitch Theory" (Imbrie & Imbrie, 1979; Hays *et al.*, 1976) but short, rapid climate change events ranging from centuries to millennia cannot be explained by orbital forcing theory alone (Walker & Lowe, 2007). The Late Glacial is the most geologically recent example of the unstable nature of global climate, and as such these rapid and often short lived climate oscillations provides a unique insight into the stability of the climate system. "The Lateglacial therefore constitutes a 'virtual laboratory' in which the complex forcing mechanisms that drive climatic change can be investigated" (Walker *et al.*, 2003, p475).

By studying climate change during the Late Glacial we can understand the complex interactions between different parts of the climate system, both in space and time. This is essential to our understanding of future climate change; if the climate does not respond in a linear fashion to climate forcing we need to understand the thresholds that can cause switches to different modes of operation, as the effects of a switch would be potentially catastrophic for society. Alternatively, if these sub-Milankovitch climate variations are part of some as yet unexplained regular climate oscillation (e.g. Renssen *et al.*, 2000) this will have wide ranging implications for our understanding of the earth's climate (Benn & Evans, 1998).

There is an increasing concern about rapid variations in future climate and the effect of anthropogenic activity on the climate system (Houghton, 2004), although our predictions regarding the nature of future change require considerable improvement (IPCC, 2007). Our understanding of Earth processes is crucial in mitigating the effects of future climate change and much of our current knowledge of Earth processes is derived from the study of past climates.

Many of the predictions made regarding anthropogenic perturbation of the climate system have been derived through climate modelling, where our knowledge of all aspects of the climate system are mathematically systemised and computer models are used to predict change on global and regional scales for a number of different scenarios (Houghton, 2004; Giorgi & Mearns, 1991; e.g. Randall *et al.*, 2007). Palaeoecological studies not only enhance our understanding of climate processes, but provide historical data needed to test models to assess the validity of their predictions; if a model cannot accurately predict historical climate change its use in simulating future climates may be limited (Stute *et al.*,

2001; Cronin, 1999).

The instrumental record is temporally and spatially poor, extending to ~150-200 years before present and mainly limited to old cities in Europe and North America, with further problems in the accuracy and homogeneity of the data (Parker *et al.*, 2000). There is then a need to increase the geographical and temporal breadth of the climatic record in order to improve the climate models (Cronin, 1999; Jansen *et al.*, 2007); this is particularly important in understanding rapid climate change, which modellers can have difficulty resolving within GCM's (Jansen *et al.*, 2007). Improved reconstructed temperatures, rainfall data *etc.* derived from palaeoclimatic research are therefore of use in improving climate change predictions.

Palaeoclimatic data can also be used to understand mechanisms that explain predicted climate change by identifying and detailing former climatic states that are not analogous to any observed phenomena today. "The use of palaeoclimate data – model comparisons are becoming standard fare in the investigations of climate change" (Cronin, 1999, p7), where they assist in explaining past rapid climate oscillations (Isarin & Renssen 1999).

However, palaeoecological records are limited in both geographical extent, the nature of the information they give and the accuracy of that information. There is a need to improve records of past climate change so that they might be used to inform our understanding of climate change and climate systems. The concern over the depth of our understanding is such that there has been a considerable research investment in recent years, particularly to develop new palaeoecological proxy techniques and absolute dating methods (see Oliver *et al.*, 1999).

2.2 Problems with proxy records

The development of pollen environmental reconstructions underpinned much of the initial research into the last glacial, and these methods are still essential to many modern reconstructions. However, there are considerable problems given the relatively low local production of pollen at the time, leading to a considerable confounding input of foreign material for many sites. The response of vegetation is also known to lag considerably behind environmental change, show be sensitive to other factors, e.g. precipitation.

The Coleoptera based models are useful but limited in their precision, using the mutual climatic range method to predict only the possible range of temperatures and some modern analogues from a small number of sites to predict the most likely value given the range. The very high resolution geochemical records from the Greenland ice have demonstrated the degree of variability in climate through the late glacial, but their use is limited by their remote geographical location, in many ways removed from the climate systems that affect areas like the British Isles and Europe.

Given the development of Chironomid based transfer functions that are both complementary and in some ways superior to the Coleoptera-based and other previously used methods there is a need to expand the use of Chironomid based studies to improve the literature on seasonal temperature through the late glacial interstadial.

This need is particularly important given the continued need for such records in climate modelling and in precipitation reconstructions from glacial and periglacial evidence. Through enhanced understanding of seasonal temperatures in the Late Glacial workers will be able to better understand precipitation responses to climate change, vegetation response to seasonality and the way in which regional and global climate functions through climate modelling.

2.3 Advantages/limits of Chironomid proxies

The non-biting midges (Chironomidae) are a family of insects of the order true flies (Insecta: Diptera); they are found in a diverse range of environments and some species can tolerate very low temperatures, extremes of acidity and are present in lakes of all trophic statuses (Cranston, 1995a; Armitage *et al.*, 1995); "there are species that thrive in almost every conceivable freshwater habitat" (Pinder, 1995 p107). They are frequently encountered in lentic, lotic and even some terrestrial environments and their chitinous larval head capsules are well preserved in lake sediments (Porinchu & MacDonald, 2003), where they are often the most numerous biological remains (Walker, 1987). The morphology of the larval head capsules is distinctive between species allowing the identification of fossil specimens to an often high taxonomic resolution (Brooks *et al.,* 2007).

Many of the thousands of species fill specific ecological niches and so Chironomid population assemblages are sensitive indicators of climate and ecology, leading to their use in limnology. The earliest use was in lake classification (Porinchu & MacDonald, 2003), and early work based on Chironomid communities have been referred to as the "beginning stages…of limnology as a modern sci-

ence" (Saether, 1979 p65). Characteristic species were used as indicators of trophic status (e.g. Wiederholm, 1979; Wiederholm & Eriksson, 1978) and in pollution monitoring (McGill *et al.*, 1979; Kansamen, 1985). These earlier studies were always qualitative and limited by the ecological information and statistical techniques available for performing analysis.

More recent studies have seen a sharp increase in Chironomid based methods in palaeolimnology, using computer based statistical techniques to infer absolute values for past lake conditions. These include salinity (Mousavi, 2002), trophic status (Langdon *et al.*, 2004), acidification (Brodin & Gransberg, 1993) and oxygen supply (Clerk *et al.*, 2000). Currently, one of the most widespread uses of Chironomids in palaeolimnology is to reconstruct past temperatures, as Chironomids are particularly sensitive to mean seasonal temperatures and mean annual temperature (Lotter *et al.*, 1997; Boggero *et al.*, 2006). Walker *et al.* (1991) found that mean summer surface water temperature best explained variation in Chironomid populations. The criticisms of Hann *et al.* (1992) who argued that the influence of depth, transparency and substrate were more important in explaining variation have largely been refuted by subsequent studies, the wealth of which have concluded mean summer temperature or mean July temperature explains the variation of modern Chironomid populations well (e.g. Larocque *et al.*, 2001; Brooks *et al.*, 2001; Luoto, 2008). Chironomid populations are far less affected by winter temperature, although seasonality may well have an influence on the length of the breeding season for those species that breed more than once a year (see Tokeshi, 1995).

The first temperature transfer function (Walker *et al.*, 1991) was criticized for using single measurements to represent mean summer temperatures (Hann *et al.,* 1992), and more recent studies have used local weather stations corrected for altitude to provide a more realistic estimate, along with a broad range of measurements of other geochemical and environmental variables, although the accuracy of the original work remains good (Walker *et al.,* 1992). All Chironomid transfer functions use weighted average partial least squares regression (WA-PLS), a statistical technique that uses modern populations to calculate the optimum value and tolerance range of particular taxa in relation to a specific environmental variable (e.g. summer temperature; see ter Braak & Juggins, 1993). The sensitivity of the proxies response to environmental variables is usually tested using multivariate ordination techniques, and the predictive power is calculated using "jackknifing" and "bootstrapping". Similar methods have proved extremely valuable in diatom based palaeolimnology for several years (see Battarbee, 1994). A three component WA-PLS model has been shown for Chironomid based reconstructions to be the most accurate at inferring temperature compared to WA, PLS or the modern analogue tech-

nique (Lotter *et al.*, 1999). It also does not suffer from the limitations of other techniques like "key indicator species" or "mutual climatic range (MCR)" used in Coleoptera and pollen reconstructions.

Although the statistical techniques used have been available for several years (ter Braak & Juggins, 1993), Chironomid based studies have been limited by the poor taxonomic resolution of the family for sub-fossil larval specimens. Chironomid taxonomy has been far from harmonious between workers (Murray, 1980; Porinchu & MacDonald, 2003), and until the publication of a more complete guide by Brooks *et al.* (2007) sub-fossil Chironomid taxonomy for species encountered in the palaeoarctic was limited mainly to the works of Wiederholm (1983) and Oliver & Roussel (1983); both are limited in the number of species described and the taxonomic detail. The only other more recent guide available prior to Brooks *et al.* (2007) focused primarily on North American species (Walker, 2007) and being web-based was constantly updated, rendering it unsuitable for use in constructing training sets. The only guides written specifically for the British Isles (e.g. Cranston, 1982) are based on diagnostic features rarely present in sub-fossil specimens and do not include several species encountered in Late Glacial samples that are no longer present in the British Isles.

Brooks *et al.* (2007) has been considered the first comprehensive taxonomic guide focussed on sub-fossil specimens encountered in the palaeoartic (Brodersen, 2008) and is the compatible with the taxonomy used in the training set in this study (S. J. Brooks, pers. comm.). It is imperative that the same taxonomy is used in both the modern training set and samples under investigation if the transfer function is to be applied correctly (Brodersen, 2008). The publication widens the possibilities for new workers to use Chironomid based reconstructions in palaeolimnology and palaeoclimatology.

There are a rapidly growing number of Chironomid based temperature transfer functions that have been developed for use in the European Palaearctic (e.g. Olander *et al.*, 1999; Brooks & Birks, 2001; Luoto, 2008). A wider temperature gradient, better taxonomic resolution and improved temperature data make the Norwegian dataset (Brooks and Birks, 2001) perform better than the Fennoscandian model (Olander *et al.*, 1999). Although some studies (e.g. Luoto, 2008) present lower errors, the Brooks & Birks (2001) model has the widest temperature range, is geographically relevant to the British Isles and has been proven to be effective in reconstructing temperatures in the British Isles in sediments dating to the late glacial (Brooks & Birks, 2001). A comparison of the models is given in table 5.4.

Chironomid evidence of late glacial climate has been published from only two sites in the British Isles, namely Whitrig Bog, SE Scotland (Brooks & Birks, 2000) and Hawes Water, NW England (Beford *et al.*, 2004). The Whitrig Bog study identified at least two climate oscillations during the late glacial inter-stadial, and the Hawes Water study identified four from the same period. Both broadly agreed on the temperature ranges through the Late Glacial, which were in keeping with previous research. Although attempts have been made to correlate these sequences with the Greenland ice core stratigraphy, any definite conclusions are hampered by need for more studies to elucidate the regional geography of Late Glacial climate change.

Chironomid proxies can improve on summer temperature records inferred from other biological evidence, and add a new dimension to records of mean annual temperature from geomorphological or pollen evidence in the British Isles. The improved resolution of Chironomid proxies can assist in resolving the degree of variability of the late glacial interstadial in the British Isles, as well as providing local reconstructions of temperature, which can inform research on glacial and periglacial landforms.

Chapter 3: Aims of Research

Chapter 1 identified some of the gaps in our understanding of the late glacial, and discussed the need for palaeoecologists to be able to better understand past environmental change. Chapter 2 outlined the increasingly pressing questions of anthropogenic influences on future climate and has shown that although Chironomid analysis has been shown to have the potential to contribute to resolving these problems, "Widespread use of Chironomid analysis is relatively recent" (Porinchu & MacDonald, 2003 p379), so there is considerable research potential in the British Isles.

Some questions raised by the literature are:

- What is the nature of change during the Windermere Interstadial in the British Isles?
- What were the changes in temperature, seasonality and precipitation during the Loch Lomond stadial in the British Isles?
- How can Chironomid proxies add to our understanding of palaeoecology?

There is considerable potential for clarifying the palaeoecological record in the British Isles using a Chironomid based proxy, particularly in helping to resolve the particulars of climate during the Younger Dryas, and to identify the timing and magnitude of change during the Windermere interstadial.

In addition, during the past five years there have been a number of studies using Chironomidae proxies to reconstruct past air temperatures, and although some have compared them to other local proxy records few explicitly evaluate their use. Further palaeoecological studies using modern techniques are desirable simply because they allow comparison with the existing literature, testing the methods and adding to the continual improvement of research methods.

This dissertation aims to:

- Inform the debate on the late glacial climate oscillations in the British Isles.

How?: By applying a modern quantitative Chironomid based temperature inference model to a Late Glacial lake sediment stratigraphy.

- Evaluate the Chironomid based temperature reconstructions against other relevant studies that have inferred temperatures for the same region.

How?: By comparing the reconstruction with the existing literature.

- Inform the debate on the level of precipitation during the Younger Dryas.

How?: By using Chironomid inferred temperatures as part of a glacier mass balance model.

It will conclude with a brief assessment of the methods used and the future use of Chironomid based temperature reconstructions in the British Isles.

Chapter 4: Site Description

4.1 Location, Vegetation & Topography

The chosen site for coring is a ~400 m^2 area of mire called Ffridd-y-Fawnog, located approximately 8km west-north-west of Y Bala, Wales in Great Britain (figs. 4.1 & 4.2). It is level at +370m m.s.l and currently covered by mire vegetation (as defined by Rodwell, 1991), with *Sphagnum* spp. and *Eriophorum* spp. dominating the flora and correspondingly high water table (pers. obs.). The surroundings are mountainous, with the Arenig range surrounding the site and the highlands of Snowdonia National Park to the North. The site is common land used for low density sheep grazing.

4.2 Previous Onsite Research

The site is known to have formerly been a small lake basin (Hughes, 2000), draining a small catchment of approximately 1.5 km^2. The bedrock is made up of volcanic rocks and there is a drift of diamicton of Devensian origin (BGS, 2008). There is a four part stratigraphic sequence of sedimentation of the palaeo-lake, which has been shown to represent the Late Glacial climate oscillation at this site (Hughes, 2000) and is analogous to other sites in the British Isles that are know to date to the Late Glacial (e.g. Walker, 1980; Ince, 1983; Hill *et al.* 2008). The sequence is as follows:

Holocene peats

Holocene organic lake sediments

Younger Dryas grey clays

Windermere interstadial organic lake sediment
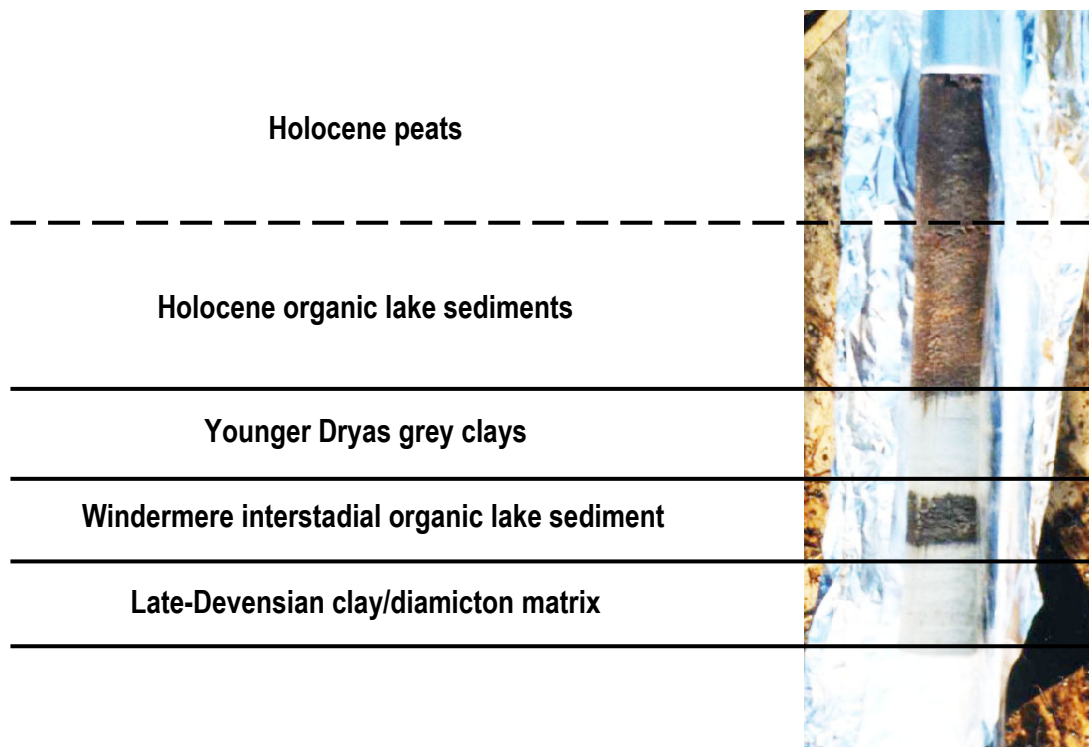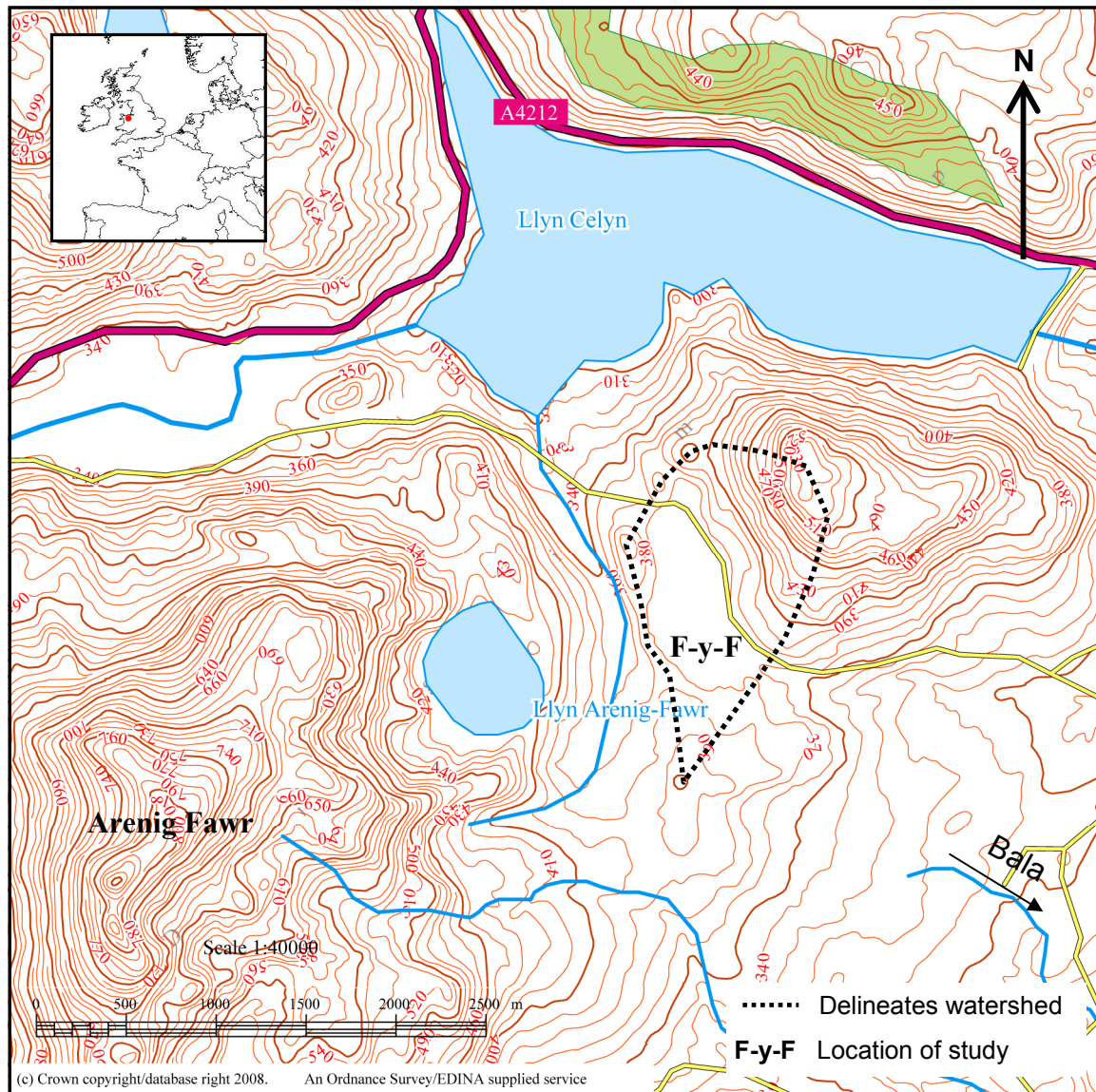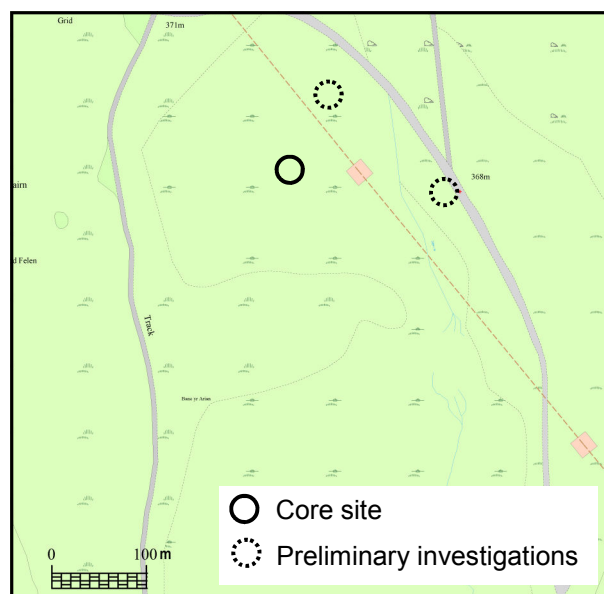
Late-Devensian clay/diamicton matrix

Figure 4.3: Photograph of the 2002 core from Fridd-y-Fawnog. Annotations represent the findings of Hughes (2000). Courtesy of P. Hughes.

Above: *Figure 4.1. Topographical map of Ffridd-y-Fawnog showing the possible catchment.* Contours are drawn every 10m.

Right: *Figure 4.2. Detailed map of the core site, showing the locations of coring.*

The pollen record for the bottom Fridd-y-Fawnog sediment profile is presented by Hughes (2000), and is a classic late glacial vegetation sequence synonymous with a large number of sites in the British Isles that date to the late glacial (Walker *et al.*, 2003; Isarin & Bohncke 1999; Walker, 1980; Ince, 1983); it is described fully in Chapter 7. The topmost peat sequence has been studied and found to represent a sequence of lake sediment and peats transition from lake to terrestrial peat sediment that date to the Holocene (L. Richardson, pers. comm.).

4.3 Previous Local Research

In areas where a high number of pollen studies have been conducted and dated using radiocarbon dating it is sometimes possible to coarsely date sediment using the pollen profile alone (see Walker, 2005). The pollen zones in the British Isles are distinctive and well constrained by radiocarbon (^{14}C) dating of material in a number of local studies (e.g. Walker, 1980; Ince, 1983; Isarin & Bohncke 1999; Walker *et al*, 2003), and although ^{14}C dating would allow a more reliable framework for dating (see Walker, 2005) is not available to this study. The pollen based chronology in Hughes (2002) can robustly date the sediment to the level of accuracy needed for this project (Davis, 1984).

There are a number of studies that have attempted to reconstruct Late Glacial temperatures in the region, including evidence from Coleoptera remains (e.g. Coope & Brophy, 1972) pollen and macrofossils (e.g. Walker *et al.,* 2003) and glacial geomorphology (e.g. Hughes, 2002). These reconstructions tend to give temperatures roughly analogous to the present day during the Bøller/Allerød, and summer temperatures in the region of 10°c for the Younger Dryas. There are other regional studies using Coleoptera (Atkinson *et al.*, 1987; Coope *et al.*, 1998), giving a good basis for comparison with other independent records. Figure 4.4 is a summary of these reconstructions and demonstrates the similarity and variation between them.

Proxy	Extent	T_{max} – B/A	T_{max} – Y.D	Reference
Coleoptera	British Isles	14-18°c	9-15°c	Atkinson *et al.,* 1987
Coleoptera	England & Wales	15-17°c	9-11°c	Coope *et al.* 1998
Glacial geomorphology	Arenigs, Gwynedd	n/a	~8.4°c (yr. av.)	Hughes, 2002
Coleoptera	North Wales	~15-17°c	~10 °c	Coope & Brophy, 1972
Pollen/plant macrofossils	Western Europe	n/a	11-12°c	Isarin & Bohncke, 1999

Fig 4.4. Summary of summer temperature reconstructions applicable to Fridd-y-Fawnog, Gwynedd. (B/A stands for Bøller/Allerød interstadial, Y.D for the Younger Dryas stadial)

Chapter 5: Methods

5.1 Core extraction and sub-sampling

The coring site was chosen as the location most likely to be the deepest part of the palaeo-lake based on the modern surface topography and test cores made at strategic locations thought to represent to represent the palaeo-lake shoreline. Sampling should be from a deeper point of a lake as the Chironomid fauna varies by lake depth; the sediments from the deepest parts of a shallow lake provide a good representative sample of the dominant fauna of the whole lake (Heiri, 2004). This is also likely the area of highest sedimentation, so producing the most material per unit time for analysis (Glew *et al*., 2001) and hence better resolution. A high voltage electricity pylon and its fittings were accommodated and the site was well away from the pylon, power lines and the area between the pylon and the road, where heavy machinery may have been operated to build both the road and the pylon and disturbed the sediments.

A sequence of 500mm cores were extracted using a Russian-type corer for the soft lake and peat sediments, and a gauge auger was used to extract the more consolidated clays. The total depth was 710cm below surface level. Additional cores were taken at obvious stratigraphic boundaries to provide extra material for analysis/preliminary tests if necessary.

The cores were extruded on site and transported horizontally in 4 inch gauge semicircular PVC pipe covered by LDPE plastic wrap. The cores were sub-sampled at ~30mm intervals, the actual length depending on the volume of material retrieved. All samples were stored in glass receptacles at +4ºc and sealed in LDPE film.

5.2 Loss on Ignition Measurements

Loss on ignition measurements were performed at approx. 10cm intervals, depending on the availability of material. The samples were dried at +105ºc for 16 hours and weighed to four decimal places, then ignited for 2 hours at 550ºc and reweighed. The percentage loss between the two is expressed as the loss-on-ignition; this is the standard method described by Bengtsson & Enell (1986). The loss-on-ignition measurements help correlate the stratigraphy of the core with the pollen chronology of Hughes (2002) as well as expressing any relative changes in the organic content of the samples.

5.3 Chironomid sorting & slide mounting

The methodology for processing Chironomid remains generally follows a standard procedure of heating the sediment in Potassium Hydroxide (10% KOH) to disaggregate organic material bound by humic acids, assisted by gentle agitation. The sediment is then separated by sieving to allow hand picking of the Chironomid remains (as per Walker & Paterson, 1985), although some workers have used ultrasonic bath processing (e.g. Lang *et al.*, 2003) and kerosene flotation (Rolland & Larocque, 2007).

The chosen method is often dependent on the type of sediment being prepared, so a preliminary test was conducted on a redundant section of the retrieved clays. There was no significant difference in the total number or relative abundance of Chironomid remains between ultrasonically processed sediment or sediment prepared using the standard KOH method, contrary to the conclusions of Lang *et al.* (2003), so a method minimising the possibly of damaging the capsules or introducing preferential preservation was used. The kerosene flotation method was not appropriate for the departmental lab environment, and so was not evaluated.

The method described by Brooks *et al.* (2007) was followed, although the clay sediments were subjected to a longer KOH bath as used by others (e.g. Olander *et al.*, 1999). The exact method used is described in fig. 5.1.

The two size fractions were sorted at x20+ magnification under a stereomicroscope with fine forceps and needles, and the Chironomid head capsules from each sample were separated from other matter and stored in a shallow petri dish in deionised water (as per Walker, 2001). The Chironomid head capsules were then carefully air dried and mounted on glass slides in Euparal© mounting fluid with standard 22x22mm glass cover slips (as per Porinchu & MacDonald, 2003).

Instruction	Notes/reason
1. Place subsample in 100m beaker	Keep airtight in cold store
2. Decant 50ml 10% KOH into beakers	To deflocculate the sediment
3. Place on hotplate at ~70°c	
4. Occasionally gently agitate and remove after 20 minutes	(30m for minerogenic sediments)
5. Sieve to 125 & 250 micron fractions	To exclude fine fractions
6. Decant fractions to petri dish in solution of deionised H_2O	To allow analysis under sterioscope
7. Save contents of wash bucket & condense w/ centrifuge	For possible pollen analysis

Fig. 5.1. Procedure used to prepare samples for chironomid analysis

5.4 Chironomid taxonomy

As previously discussed, Chironomid taxonomy is far from harmonious between workers. The taxonomic guide employed (Brooks *et al.*, 2007) is compatible with the taxonomy employed by Brooks & Birks (2001; S.J. Brooks, pers. comm.), the transfer function used in this study. This guide is considered to be the most comprehensive and simplest to use, and in any case it is imperative that the same taxonomy is used in both the modern training set and samples under investigation if the transfer function is to be applied correctly. Examples of some of the specimens encountered are given in figure 5.2.

Prepared Chironomid slides were viewed under x40-100 magnification under a transmitted light microscope. Interference optics were occasionally used to elucidate detail in the specimens. Each specimen was identified to the best possible level of taxonomic resolution with notes taken on the size and condition of each capsule encountered (see appendix a). Micrographs were employed to compare specimens where appropriate.

Identification is based mainly on the morphology of the mentum, ventromental plates and the number and condition of teeth on the manibles (see figure 5.3). A number of *Tanypodinae* can also be differentiated on the basis of the arrangement of the cephalic setae (Rieradevall & Brooks, 2001). Species were identified to the highest possible taxonomic resolution, and where there was uncertainty the lower resolution was always recorded.
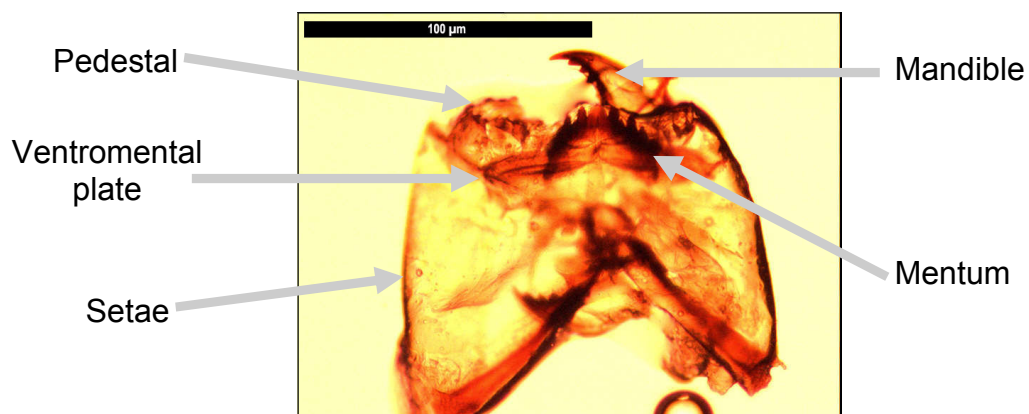
Figure 5.3. A micrograph of a Pseudochironomus *sub-fossil larval head capsule, with the parts of the anatomy important to identification labelled.*

Figure 5.2a-f: Examples of the chironomid taxonomy used in this study
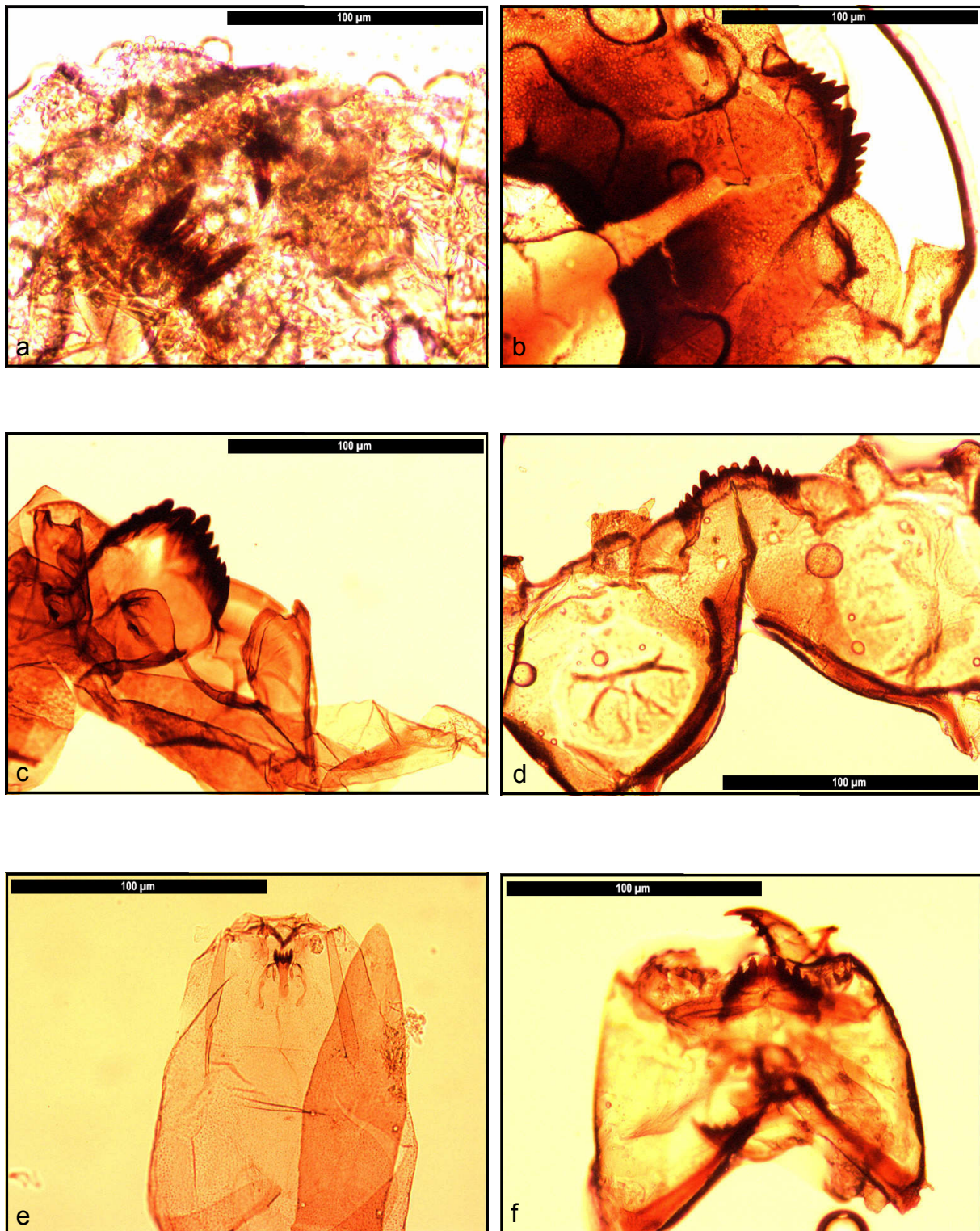
Transmitted light micrographs of a) *Tanypodinae*, b) *Einfeldia pagana*, c) *Glyptoten-dipes pallens*, d) *Microtendipes pedellus*, e) *Thienemannimyia*-type, f) *Psuedochi-ronomus*

5.5 Temperature transfer function

The transfer function used is a three component, weighted average, partial least squares (WA-PLS) regression model constructed from data collected from 109 Norwegian lakes covering a July temperature range (T_j) of 3.5-15.6°c (Brooks & Birks, 2001); further details are given in figure 5.4, along with the details of other transfer functions. It can be seen from these figures this model is best suited to the study. Dr. Steve Brooks of The Natural History Museum, London kindly agreed to run the temperature transfer functions from unpublished data described in Brooks & Birks (2001). This model has been previously used in the British Isles (Brooks & Birks, 2000; Bedford et al., 2004) and Norway (Brooks & Birks, 2001).

It is generally accepted that a minimum of around 45 head capsule counts per sample is generally required, because there is evidence that suggests under this number there is considerable error in the ability to gain a realistic sample of the population (Heiri & Lotter 2001). Samples can be bulked together if they fall under this, but this obviously leads to a decrease in temporal resolution.

	Brooks & Birks, 2001	Walker, 1997	Olander et al., 1999	Larocque et al., 2001	Luoto, 2008
Area	Norway	Canada	Finland	N. Swedish Lapland	Finland
No. of lakes	109	39	53	100	77
No. of taxa	119	34	38	85	110
Range	0.3-23.0	6-27	6.1-15.4	7-14.7	17-11.3
RMSEP	2.13	2.26	1.53	1.13	0.721
Max. bias	2.84	2.4	3.88	2.1	0.794
WA-PLS components	3	2	1	2	1

Figure 5.4. Table showing statistics relating to chironomid based summer temperature transfer functions. The transfer function used in this study is marked in bold.

Chapter 6: Data Presentation & Description

6.1 Visual stratigraphy and loss on ignition results

Figure 6.1 shows the loss on ignition results and visual stratigraphy, showing the same tripartite sequence that characterised earlier work on the site (Hughes, 2000). The top of the core is made up of terrestrial peats, including several fragments of woody debris. The LOI results have a wide range from just 3% to over 87%, and there are rapid changes between some samples, the greatest change being 52% over 188mm. The size of the variation is too large to be attributed to the confounding effects of different sediment types (Santisteban *et al.*, 2004), although these effects may preclude the use of this dataset to quantitatively calculate organic content. The results still provide a robust qualitative measure. The major stratigraphic boundaries correlate well with the main Chironomid faunal zones.
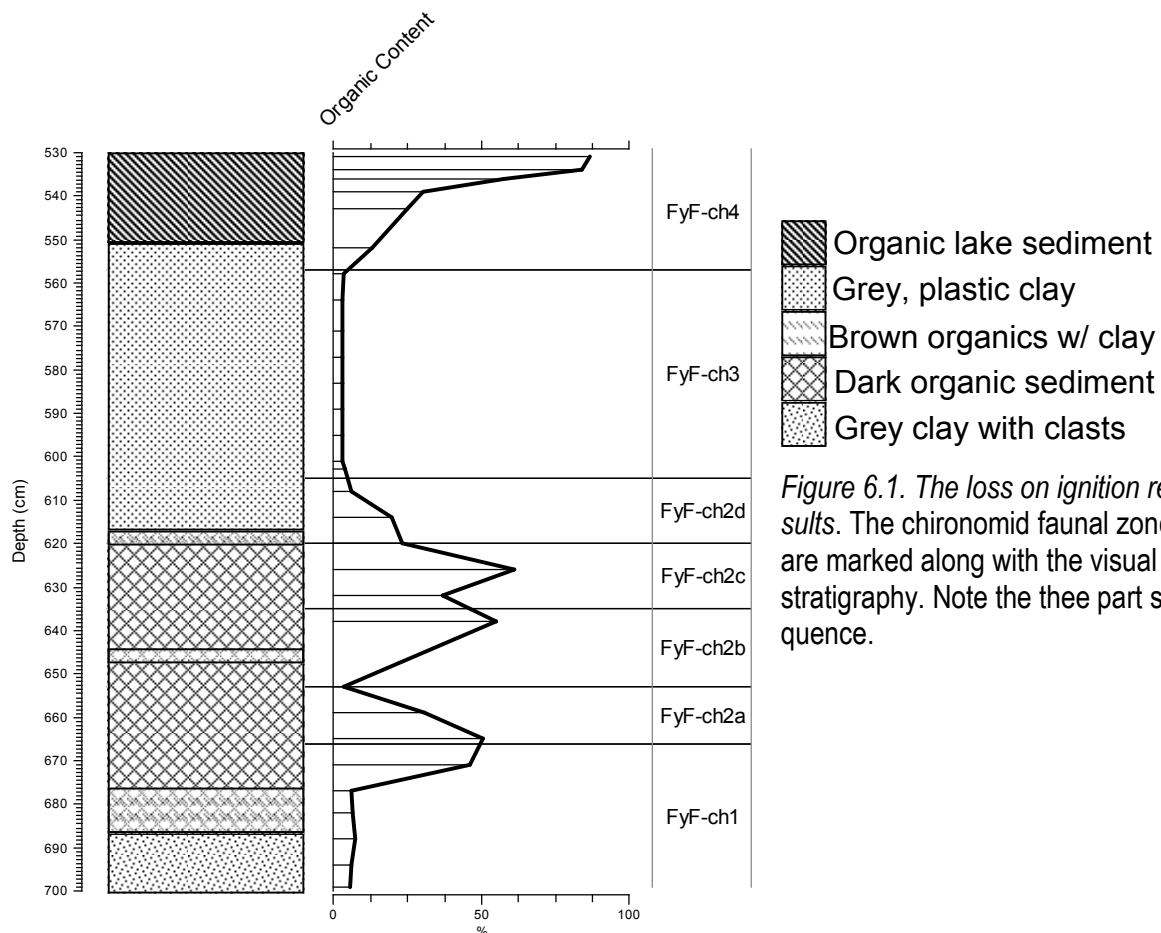

Figure 6.1. The loss on ignition results. The chironomid faunal zones are marked along with the visual stratigraphy. Note the thee part sequence.

6.2 Chironomid counts & stratigraphy

The bottom section of core which was examined yielded a total of 698 individual head capsules, 590 (85%) of which could be identified to at least tribe level. The number of capsules varied greatly between samples, with a standard deviation of 17.26. The highest head counts (n=64.5) were encountered in the in organic material, and the lowest densities (n=0) in the minerogenic sediments, although the necessary use of a gauge auger to retrieve the clays limited the total volume of sediment, and hence the number of remains retrieved from the lower parts of the sequence.

There were 51 different taxonomic groups, most of which were species level groupings (although some were at genus/tribe level, a full list is given in appendix c). However, 32 of these taxa had less than 5% abundance and less than 5 occurrences. Although these taxa provide important information about species richness and diversity, individually they are of limited use in the reconstructions, so where possible they have been bulked into larger taxonomic groups, otherwise they have been grouped together. There are no strict rules regarding cut off values for reconstructions, so all taxa were included in the transfer function, but in stratigraphic diagrams rare taxa, as defined above were bulked together.

6.2.1 Zone Descriptions

The grouping was based initially on a number of different constrained cluster analyses (shown in fig. 6.2), adjusted slightly after manual analysis. The zones are prefixed FyF-ch so they are differentiated from the pollen stratigraphic zones for the site in Hughes (2000). A stratigraphic diagram of all major taxa is shown in figure 6.4 to which the following zone descriptions refer to.

The diversity is shown in figure 6.3. The diagram shows a gradual increase in diversity, punctuated by small extinction events, although the fauna remains relatively diverse throughout the section. The species diversity is discussed more fully in chapter 7.

Zone FyF-ch1 (700-666cm)

This zone could be divided into two sections on the basis of a distinct faunal change at 687cm, but because of very low count sums for the oldest section there can be no guarantee that this is not a purely statistical artefact.

For the first section of the zone there are very low count sums which add caution to any descriptions of the fauna derived from these samples. However, the lake appears to be species poor, the fauna being
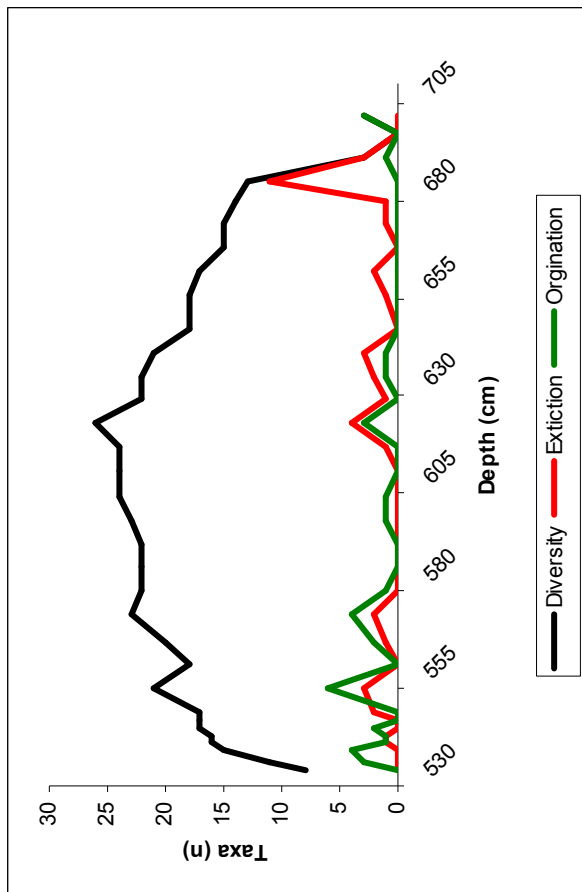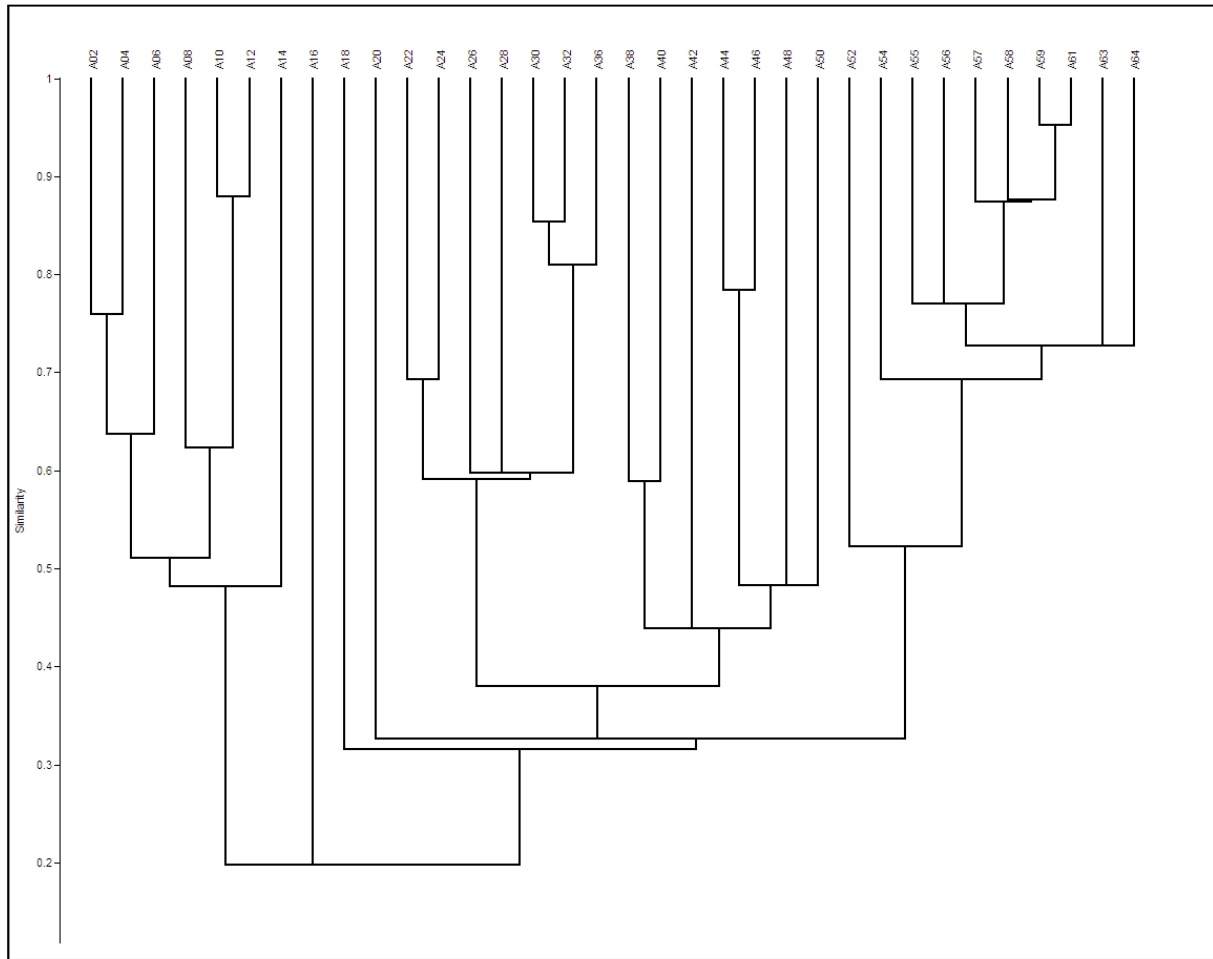
Above: *Figure 6.3. Species diversity plotted against depth.*

Right: *Figure 6.2. Constrained cluster analysis of chironomid stratigraphy. The model used is a paired group algorithm using the "Horn" measure of similarity. The analysis was performed using "PAST" (Ryan et al. 1995)*

*Figure 6.4. Stratigraphic diagram of major taxa. The diagram is arranged alphabetically by taxonomic group (**tribes** are marked in bold).*
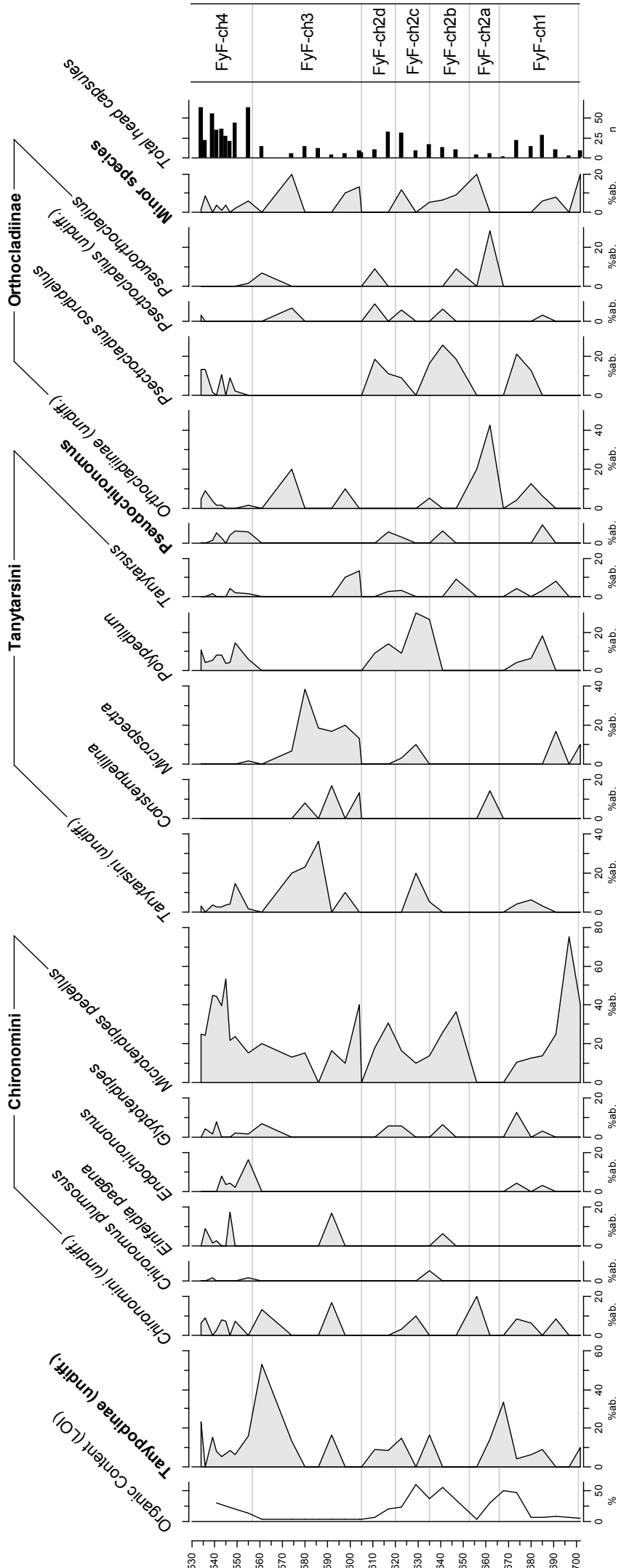
dominated by *Microtendipes pedellus* and *Microspectra* and a number of poorly preserved remains.

At 687cm there is a distinct change in the fauna; *Microtendipes pedellus* makes a significant but drastically reduced contribution, with the first appearances of a large number of predominantly thermophilic taxa contributing to the large increase in species diversity and total head capsules. There are significant populations of *Psectrocladius sordidellus* and *Polypedilum*, whilst the presence of *Glyptotendipes* and *Polypedilum* are also recorded. The state of preservation remains poor with many remains only identifiable to a coarse taxonomic resolution.

Zone FyF-ch2 (666-605cm)

This zone is characterised by unstable populations and a changing fauna, the initial appearance of several species and a gradual increase in the total volume of head capsules. Defining the dominant faunal zones is difficult given the unstable populations, as revealed by the variations revealed by the constrained cluster analysis, and the following subzones represent the most significant changes.

FyF-ch2a (653-666cm)

The start of this zone is marked by a decline in total head capsules. The fauna is dominated by poorly preserved undifferentiated *Orthocladiinae* and *Tanypodinae*, and a smaller but significant contribution from undifferentiated Chironomidae. Of the species that could be identified *Pseudorthocladius*, *Constempellina* and *Cricotopus intersectus* were common.

FyF-ch2b (635-653cm)

There is a gradual increase in the total number of head capsules throughout this zone. This section contains a rise and fall of the *Microtendipes pedellus* population, with an optima of 35% at 645cm, followed closely by *Psectrocladius sordidellus*, with an optima of 30% at 637cm and a number of minor taxa following a similar pattern. The latter part of the zone shows a small rise in *Einfeldia pagana* and *Polypedilum sordens*.

FyF-ch2c (620-635cm)

There is a general trend of increasing numbers of head capsules punctuated by a short decline at 626cm. There is a marked decline and following rise in populations of *Psectrocladius sordidellus* and *Microtendipes pedellus*, and an inverse proliferation and subsequent waning of *Polypedilum* and later and to a lesser extent *Tanytarsini* and *Microspectra*. Some taxa are only present at the

beginning and end of the zone including *Pseudochironomus* and *Glyptotendipes*. There is the first occurrence of *Chironomus plumosus* in this zone.

FyF-ch2d (605-620cm)

There is a pronounced increase and stabilisation of head capsule numbers, followed by a decline. This stage is marked by the complete disappearance of the *Tanytarsini* and of *Microspectra*. There are significant contributions from *Microtendipes pedellus*, *Psectrocladius sordidellus* and *Polypedilum* although the large numbers of species present demonstrate a diverse population. This diverse population declines rapidly towards the end of the zone although the exact speed is not known because it is seemingly completed within a few centimetres.

Zone FyF-ch3 (605-557cm)

Head densities declined considerably in this zone to ~5 per sample. The zone is marked by the appearance of *Microspectra* and *Constempellina*. The *Tanytarsini* increase considerably. These taxa decline significantly at the onset of FyF-ch4. Many of the taxa seen in previous zones disappear at the start of the zone, and some others, for example the species *Microtendipes pedellus* persists throughout most of the zone at drastically reduced levels, although even this population disappears at two points (585cm & 564cm). At the same time as these collapses there are peaks in the number of undifferentiated *Chiromini*; the first of these could be tentatively correlated to a peak in *Microspectra*, and the second correlates well to a peak in *Microspectra* and weakly to the appearance of some other occurrences of associated taxa.

Zone FyF-ch4 (557-520cm)

The onset of FyF-ch4 is marked by a rapid and large change in the total number of head capsules, along with the return of a number of thermophilic taxa seen in FyF-ch2 and the disappearance of the cold water taxa that characterised FyF-ch3. The beginning of the zone is characterised by the proliferation of *Endochironomus albipennis* and *Pseudochironomus*. The upper section shows two obvious short lived changes in the assemblages at 545cm & 535cm; the dominant *Endochironomus*, *Pseudochironomus* & *Polypedilum* fauna declines and *Einfeldia pagana* increases briefly. The first oscillation at 545cm is the most obvious but both are clearly defined.

6.3: Temperature transfer function

The transfer function for 11 bulked samples was performed by S. J. Brooks using WAPLS v.1.5 (Juggins & ter Braak, 2002), as described in chapter 5; the results are shown in figure 6.5. The mean sample error was 1.25°c (RANGE=0.37 °c). The need to bulk the samples has led to a reduced resolution (the minimum sample size is 45 full head capsules), but the sample specific errors are low. The resolution is good in the early part of the Holocene and lowest during the period of inorganic sedimentation throughout zone FyF-ch3. The results will be fully discussed in chapter 7.
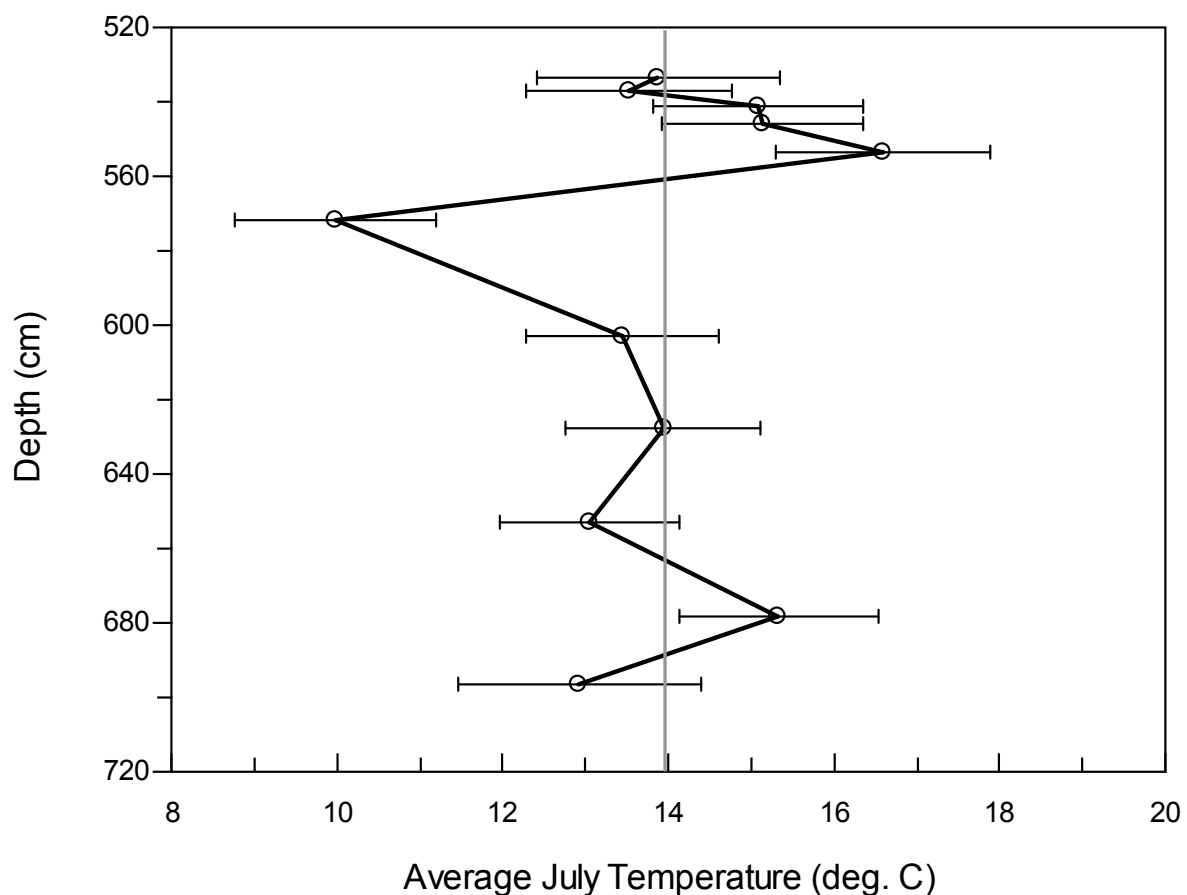

Figure 6.5. Reconstructed mean July temperatures for Fridd-y-Fawnog. Error bars represent sample specific errors as calculated using "WA-PLS" (Juggins & ter Braak, 2002). The grey line represents the modern July average at the site (Met Office, 2008a).

Chapter 7: Data Analysis & Discussion

7.1 Environmental interpretation of stratigraphy & LOI results

The visual stratigraphy and loss on ignition showed the same tripartite sequence frequently encountered in Late Glacial lake sediments, and that characterised earlier work on the site (Hughes, 2000; shown in figure 7.1). These studies have shown the bottom-most clay/diamicton matrix date to the termination of the Devensian ice sheet - the organic layer deposited on top of these clays date to the Windermere Interstadial. The grey clays found on top of these date the Loch Lomond re-advance, and the highly organic period the follows these represents the Holocene. The top of the core is made up of Holocene age terrestrial peats, including several fragments of woody debris, where the basin has filled and the terrestrial mire has accumulated.

The core extracted for use in this study can be correlated with the core extracted from the same site by Hughes (2000) on the basis of visual stratigraphy and LOI measurements. The section is longer than that extracted by Hughes (2000), suggesting that it represents a zone of higher sedimentation; this seems likely given the likelihood that the location was further towards the centre of the palaeo-lake than the core presented by Hughes (P. Hughes, pers. comm.). This is backed up by the extraction of a superficially identical but compacted sequence from the edge of the bog (site marked in figure 4.2).

The only inconsistency is the presence of a period of inorganic deposition from 653-656cm. This is based on only one measurement, but as adjoining measurements suggest a downward trend and as there is no reason to doubt the measurement it has been included, but should be approached with caution as there are many explanations for an anomalously low result. Alternatively the measurement could represent an episode of inorganic deposition that was not recorded in the shorter cores because of their lower resolution; further investigation of the composition of the sediments using a wider variety of techniques and better resolution would resolve the uncertainty over this particular result, and a thorough investigation into the basin morphology would help in interpreting differences in sedimentation.

From the LOI/visual stratigraphy and Hughes (2000) sections of the core can be dated with reasonable confidence to the key climatic episodes during the Late Glacial; this is shown in figure 7.1. It is on this basis that the sediment has been dated.
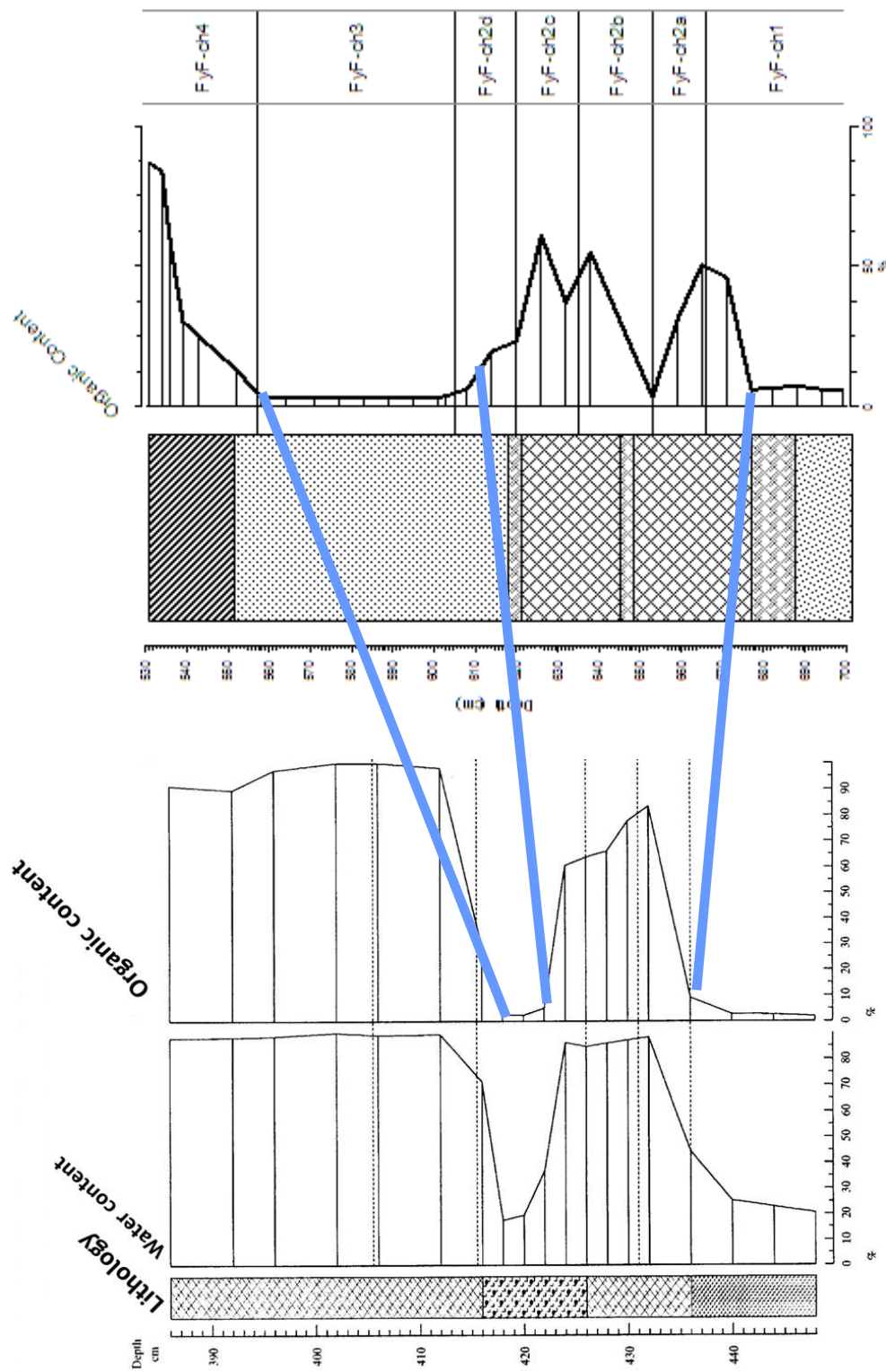
Figure 7.1. Comparison of the LOI stratigraphy (right) with Hughes (2000; left). The blue lines represent the points where changes can be cor-related.

7.2 Environmental interpretation of Chironomid stratigraphy

If the changes in the Chironomid fauna are to be interpreted as changes in the climate there should be reasonable confidence that the changes do not represent other variables, like a change in trophic status. Principle component analysis (PCA) is a technique whereby variations within a population is explained by a number of hypothetical variables (Harper, 1999). PCA of the core concluded that 64% of the variance in the Chironomid stratigraphy is explained by the first hypothetical variable; the second hypothetical variable explained 10% (see appendix d). This suggests that the populations observed are responding predominantly to a single factor: figure 7.2 shows the first and second axis eigenvectors for individual taxa, demonstrating the relationship of many taxa with this first hypothetical variable (component 1). If the variance was well explained by several factors it would confound the ability to make inferences from the data regarding just one environmental variable. Although there is no way of knowing for certain what the hypothetical variable this dataset is responding to, as discussed in chapter 2, studies of modern analogues have shown populations to be strongly responsive to temperature (e.g. Larocque *et al.*, 2001 & 2009), and so it is suggested that component 1 represents temperature.
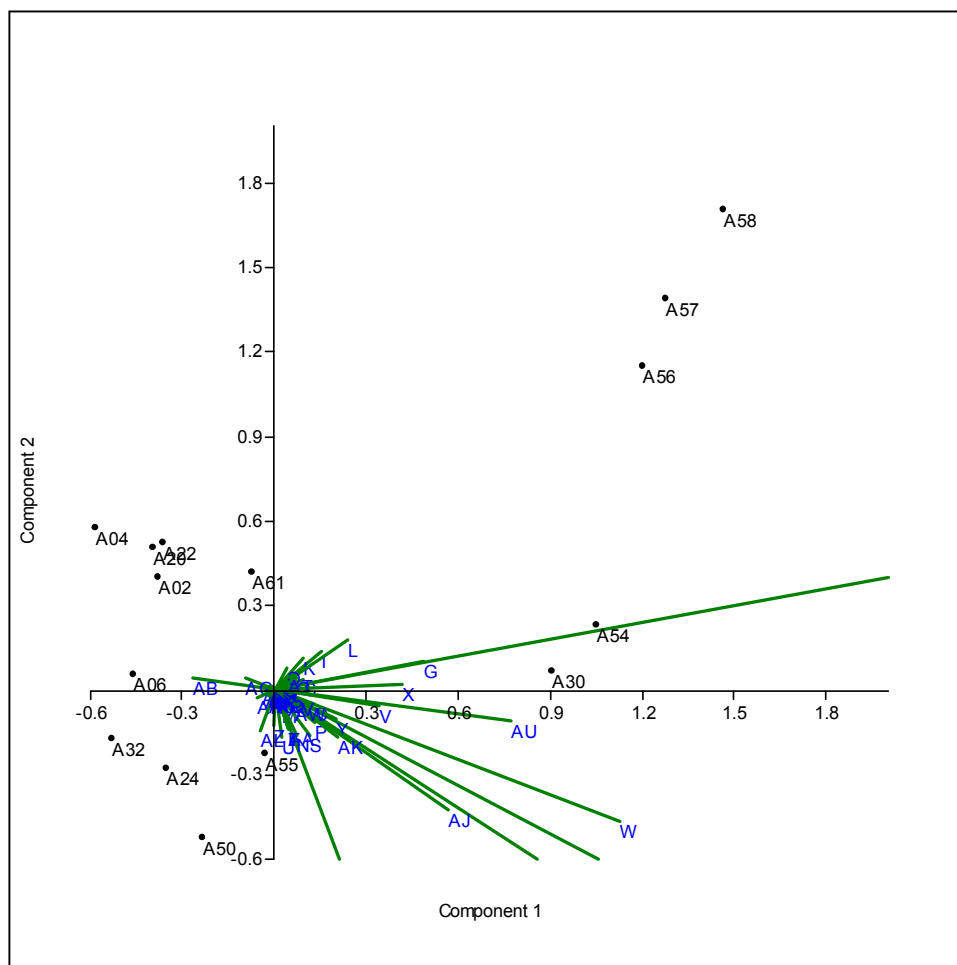
Fig 7.2. Principle component analysis of the chironomid count sums.
Components 1 and 2 account for 64.2% and 10.6% of the total variance respectively. Black dots represent samples, green lines taxa. See appendix d for further details.
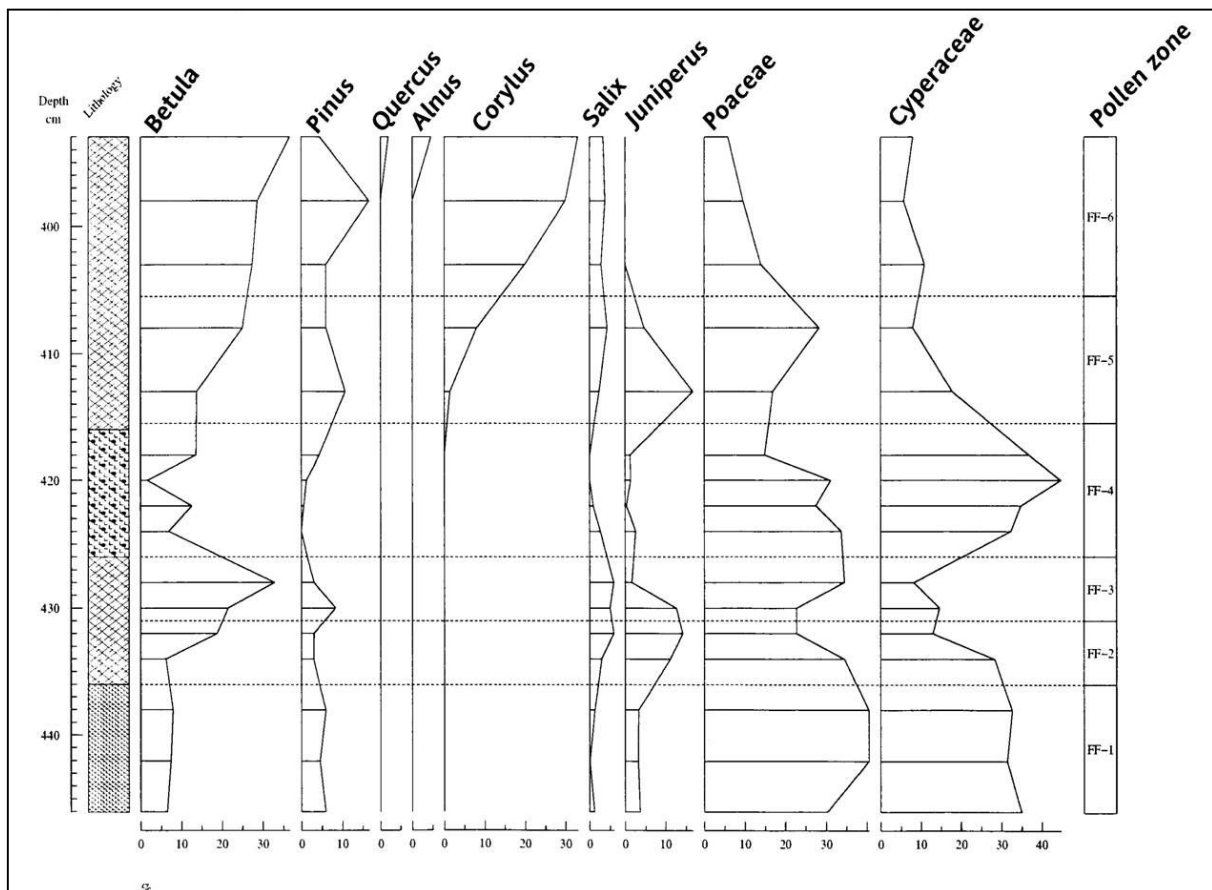
Fig 7.3. Pollen stratigraphy for Fridd-y-Fawnog. From Hughes (2000)

Given that the changes in the fauna are likely to be largely the result of temperature, a diagram showing the key temperature indicator species is given in figure 7.4. Qualitative ecological data for the taxa were collected from a number of sources (see figure 7.5 for a full list), and no major differences were found between interpretations of individual taxa response to temperature in the literature, although for ecological information for some species was unobtainable or poor. The pollen stratigraphy by Hughes (figure 7.3) has been used to contextualise the results following correlation using loss on ignition results.

7.2.1 The post-Devensian

The pollen stratigraphy for this zone is dominated by *Poaceae* and *Cyperaceae* with low *Betula* and *Pinus* counts, suggesting an open-steppe habitat and some pioneering tree populations. Exact correlation with this study should be tentative because of differences in the sedimentation rate at the two core sites, but given its stratigraphic position, the presence of diamicton clays and the pollen evidence FyF-ch1 is likely to represent the post-Devensian transition.
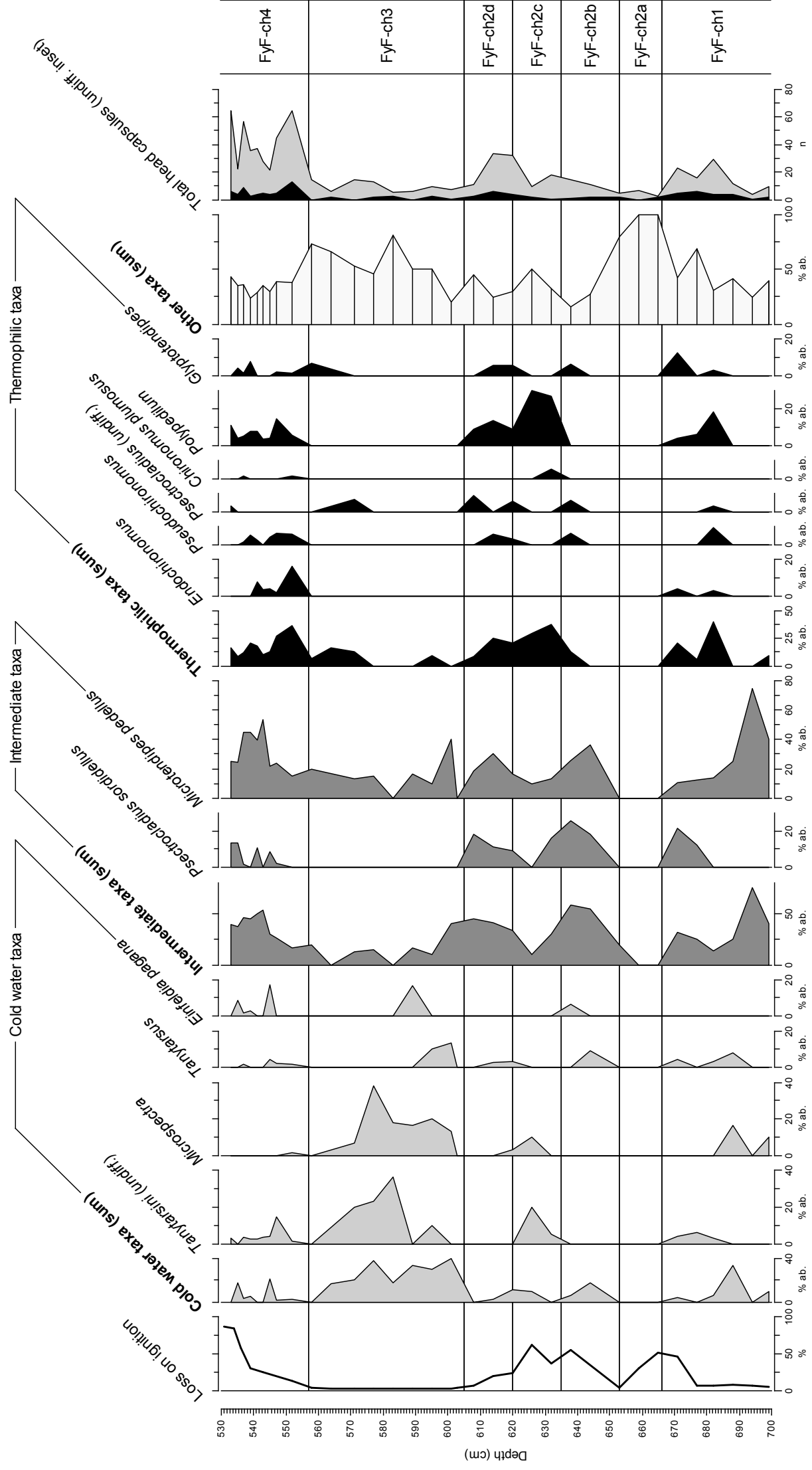
Fig 7.4. Chironomid stratigraphy from Fridd-y-Fawnog showing indicator species for temperature.

Figure 7.5. Table showing the temperature preferences of the taxa in this study. (a) Brooks *et al.,*
2007. (b) Bedford *et al.,* 2004. (c) Larocque *et al.,* 2001. (d) Cranston, 1995.

Taxa	Indicator
Ablabesmyia	Thermophilic[a]/eurytopic[b]
Chironomus anthracinus-type	thermophilic[abc]
Chironomus plumosus-type	thermophilic[abc]
Constempellina	?
Corynoneura artica-type	cold stenotherm[ab]
Cricotopus (isocladius) intersectus-type	intermediate stenotherm[ab]
Cricotopus bicinctus-type	intermediate stenotherm[ab]
Dicrotendipes undiff.	thermophilic[ab]
Dicrotendipes nervosus-type	thermophilic[abc]
Einfeldia pagana-type	cold stenotherm[a]
Endochironomus albipennis-type	thermophilic[b]
Endochironomus impar-type	thermophilic[b]
Eukiefferiella fittkaui-type	eurythermic[a]
Glptotendipes pallens-type	thermophilic[ab]
Glyptotendipes barbipes-type	thermophilic[ab]
Glyptotendipes severini-type	thermophilic[ab]
Heterotrissocladius maeaeri-type	cold stenotherm[abc]
Microspectra undiff.	?
Microspecta insignilobus-type	cold stenotherm[ac]
Microspectra pallidula-type	thermophilic[a]
Microspectra radialis-type	cold stenotherm[acd]
Microtendipes pedellus-type	intermediate stenotherm[abd]
Omisus	thermophilic[a]
Paracricotopus	?
Paramerina	eurythermic/thermophilic[a]
Paratanytarsus	cold stenotherm[c]
Paratendipes nudisquama-type	?
Polypedilum nubeculosum-type	thermophilic[ab]
Polypedilum sordens-type	thermophilic[ab]
Procladius	intermediate/eurytopic[ab]
Propsilocerus	thermophilic[a]
Psectrocladius undiff.	thermophilic[a]
Psectrocladius barbatipes-type	thermophilic[abcd]
Psectrocladius calcaratus-type	intermediate[ab]/thermophilic[cd]
Psectrocladius sordidellus-type	intermediate[ab]/thermophilic[cd]
Pseudochironomus	thermophilic[a]
Pseudodiamesa	cold stenotherm[abc]
Pseudorthocladius	?
Pseudosmittia	thermophilic[b]
Sergentia	cold stenotherm[bd]
Tanytarsus undiff.	warm[a]/cold stenotherm[c]
Tanytarsus lactescens-type	temperate[a]/cold stenotherm[c]
Tanytarsus lugens-type	cold stenotherm[acd]
Tanytarsus pallidicornis-type	warm[a]/cold stenotherm[c]
Thienemannimyia-type	warm[b]/cold stenotherm[a]
Chironomidae undiff.	Generally thermophilic[e]
Chironomini undiff.	warm stenotherm[a]
Diamesinae undiff.	cold stenotherm[e]
Oorthocladiinae undiff.	cold stenotherm[e]
Tanypodinae undiff.	?
Tanytarsini undiff.	?

The first half of the Chironomid zone initially has very low count sums, so any inferences made from these assemblages should be approached with caution with regard to their validity. Initially the intermediate taxa dominate the fauna, although they subsequently decline as there is an increase in the cold water taxa *Microspectra* and *Tanytarsus* centred around 688cm. This could represent a short cold oscillation during the post-Devensian warming, but given the low count sums this should be approached with caution, as most studies indicate rapid and constant warming immediately following deglaciation.

This possible short cooling is followed by a sharp increase in the thermophilic and intermediate taxa together with a rise in the total number of head capsules. This probably represents the transition to the full interstadial conditions of the Windermere Interstadial given the rise in productivity and temperature.

The temperature transfer function gives an average temperature of around 12.9ºc for this period, which is consistent with the rapid warming at the termination of the Devensian ice sheet, but the need to bulk samples is likely to have obscured some of the changes within this zone. However, there is an early peak in temperatures and subsequent decline, followed by a small amelioration. This suggests an early peak in temperatures and gradual cooling, punctuated by a cold event.

7.2.2 Windermere Interstadial

Through the zone of organic deposition in FyF-ch2 the macrofossils of fruits of *Najas* and other organic matter are found (*pers. obs.*), although the significant inorganic element in the LOI results suggests unstable tills and/or stream incision into Devensian tills within the catchment. The pollen zones FF-2/3 can be correlated to this zone, which show the rise of *Betula*, *Salix* & *Juniperus* and reductions in *Poaceae* & *Cyperaceae*, suggesting the development of an open woodland environment. It is on this basis this section has been identified as representing the entire Windermere Interstadial.

The zone starts with a possible cold fauna in FyF-ch2a, although this is difficult to assert fully. This is followed by intermediate, slowly rising temperatures in FyF-ch2b. These intermediate temperatures are continued in FyF-ch2c, followed by a brief warming and subsequent return to intermediate temperatures in FyF-ch2d, although there may have been a possible short cold period following the climatic optima in FyF-ch2c. The increased organic deposition probably reflects a combination of increased productivity, local soil development and slope stabilisation. There is considerable variation within the zone, and as such the sub-divisions are treated separately below.

FyF-ch2a

This zone has very low Chironomid counts and is dominated by the *Orthocladiinae*, and *Tanypodinae*. The dominant *Orthocladiinae* are known the feed on diatoms (Pinder, 1995), and could be considered to be cold stenotherms (Brooks *et al.*, 2007, Pinder, 1995). Some *Psuedorthocladius* spp. & other *Orthocladiinae* spp. inhabit terrestrial habitats (Pinder, 1995) opening the possibility they may be allocondronous, although this is impossible to prove without analysing further samples. The only known European species of *Constempellina*, *C. Brevicosta* (Brooks *et al.,* 2007) has flourished in cold climates (Grimås & Wiederholm, 1979). Accompanied by the reduction in organic deposition, suggesting periglacial weathering of tills during this zone it may represent a return to cold conditions. The *Orthocladiinae* are, however, a diverse and poorly understood genus (Cranston, 1995b) and so a cautious inference that the zone represents a pioneering, cold stenothermic fauna is hampered by the lack of information on the genus.

FyF-ch2b

There is a gradual rise of Chironomid abundance throughout this zone, indicating increased lake productivity. The zone is dominated by the intermediate taxa, whilst cold water taxa are present in the first half of the zone their populations reduce as the thermophilic taxa appear at low levels towards the end of the zone. This suggests intermediate, slowly rising temperatures throughout the zone.

FyF-ch2c

This zone encompasses the collapse of the dominant intermediate fauna and increases in the thermophilic taxa *Polypedilum*, suggesting a rise to warmer temperatures. However, there is also a smaller rise in *Tanytarsini* and *Microspectra* which are cold water taxa. The undifferentiated *Tanytarsini* and *Microspectra* might represent specific species within those sub-genii that are warm stenotherms (e.g. *Tanytarsus lactescens* or *Microspectra pallidula*), and the higher numbers of head capsules supports the case for a warmer, more productive lake through this period. It may also indicate a specific seasonal regime where such a diverse fauna could exist, but the exact emergence patterns of particular species are poorly understood.

Alternatively there could be a short lived cold period within this zone that is disguised by the resolution of the sampling or record. This zone would benefit from higher resolution sampling to establish the exact succession of the fauna, but there is certainly the possibility of a short lived cold stage dur-

ing this period. The termination of the zone is marked by a return of the intermediate species at the expense of the thermophilic species, suggesting cooling. This warm interlude centred on 625cm could be interpreted as the climatic optimum of the Windermere Interstadial, although the timing seems a little late compared to other records.

FyF-ch2d

This zone is dominated by intermediate and thermophilic taxa, suggesting warm conditions. The head capsule counts are significantly higher during this zone, suggesting it may represent the period of highest productivity during the Interstadial. The populations are also predominantly meso-eutrophic indicators (see figure 7.6) which reflect the richer vegetation both in and around the lake.

The July temperature from the transfer function suggest summer temperatures very similar, or slightly higher than the present day, for which the July average is 15.4°c (Met Office, 2008a); assuming an environmental lapse rate of 0.6°c/100m, this equates to 13.1°c at the site altitude.
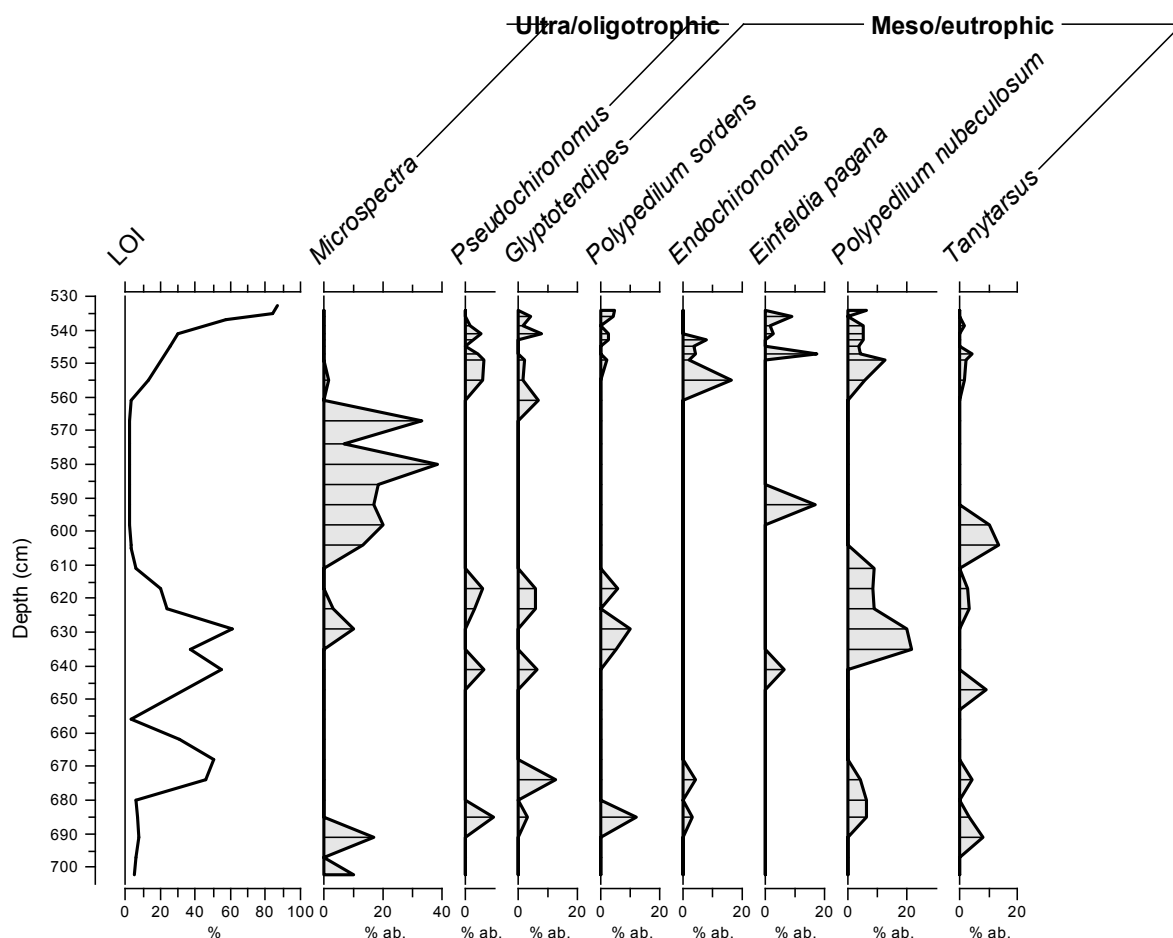
Figure 7.6. Chironomid stratigraphy from Fridd-y-Fawnog showing indicator species for trophic status

7.2.3 The Loch Lomond Re-advance

Hughes (2002) used the pollen sequence to attribute the grey clays at the site to the Loch Lomond Re-advance; the pollen record (FF-4) is dominated by *Poaceae* and *Cyperaceae*, with a *Betula* contribution. In the later part of the chronon *Pinus* increases in abundance. This is suggestive of an open-steppe grassland environment. There was a distinct lack of plant macrofossil material from this zone (*pers. obs.*), suggesting low lake productivity.

Although the dating framework and resolution of this study precludes a confident assessment of the rapidity of the onset of the Younger Dryas the shift in the proportion of cold adjusted taxa from 0% to 40% of the fauna over two samples is in keeping with the rapidity shown in other records. The transition is characterised by a collapse of the thermophilic fauna, and a contemporous rise and subsequent fall in the intermediate fauna and finally an expansion of cold water taxa. This rapid and complete change in the population is unlikely to be due to changes in substrate or food supply; for example, *Microtendipes pedellus* displays a preference for fine organics sediment (Brooks & Birks, 2001) so should thrive in the Windermere Interstadial, but its populations are suppressed during this zone, presumably because *Microtendipes* temperature optima is intermediate (Brooks & Birks, 2001; Brooks *et al.*, 2007).

The Chironomid fauna throughout the zone is dominated by cold water taxa. The intermediate taxa *Microtendipes pedellus*-type is present throughout most of the Loch Lomond Re-advance at low levels, and thermophilic taxa are absent from all but the onset and termination of the chronon. Peaks in the *Microspectra* population suggest there was some variation within the zone, although the general fauna remains comparatively stable throughout the chronon.

The return of thermophilic species in the second half of the zone could suggest warming or a change in the seasonality. The low number of Chironomid remains is usual in cold water lakes (Brooks *et al.*, 2007) and reflects the low productivity of cold water systems. An increased rate of sedimentation would have further reduced the concentration of remains; based on the GRIP/NGRIP ice core chronology (Rasmussen *et al.,* 2006) a rough estimate of 37.7mm per century could be speculated.

7.2.4 The Holocene (sub-Boreal)

This zone is made up of sediment of increasingly organic composition, and the pollen record (FF-5/6) shows a general decline of *Poaceae* and *Cyperaceae*, an increase in *Juniperus* followed by a decline

as *Betula*, *Pinus* & *Corylus* populations expand. There is the first appearance of *Quercus* & *Alnus* at the uppermost samples. This characteristic sequence represents the start of the Holocene in the British Isles (Sparks & West, 1972; Walker, 1982).

The Chironomid fauna changes completely – there is a rapid increase in the total number of head capsules and the reappearance of the thermophilic taxa prevalent in FyF-ch2; the cold water taxa prevalent in FyF-ch3 disappear. This suggests a warmer lake, and the reconstructed July temperatures for this section are correspondingly high, averaging 14.8°c.

There are two possible cold oscillations at 545 and 535cm, given the decline in the thermophilic taxa. Although this interpretation is limited by the lack of published data on the ecology of *Einfeldia pagana*, *Pseudochironomus* and *Polypedilum* are both thermophilic and there is a definite reduction in their populations. The decline in their populations could be tentatively interpreted as representing cooler conditions.

It is known that chironomids are responsive to trophic status, and this zone marks the reappearance of eutrophication, although the ecological niches for temperature and trophic status are intimately linked in Chironomid populations.

7.3 Comparison with other Late Glacial Climate Interpretations

The Chironomid record from Fridd-y-Fawnog is in general agreement with our current understanding of climate during the Late Glacial, although the very earliest sediments are difficult to interpret without absolute dating or further coring. A summary of the key changes in the Chironomid population is provided in figure 7.7.

The period immediately following termination of the Devensian ice sheet is often poorly recorded in most terrestrial proxies, and this record is no exception. There are Chironomid remains at the bottom of the core, albeit in very low quantities, and it is only by coring down to sterile clays or bedrock that the record can be fully understood. However, the possible temperature oscillation identified at the transition to the Windermere Interstadial could correlate with that identified by Bedford *et al.* (2004).

The Windermere Interstadial is notable in that the qualitative interpretation suggests optimal temperatures late in the zone. The Greenland records suggest early warming followed by progressive cooling

during the Windermere Interstadial period, and this is reflected in the Chironomid record of Brooks & Birks (2000). The record presented here has greater affinity with the Chironomid stratigraphy of Bedford *et al.* (2004), where initial warming is slower and interrupted by an early cooling event. If there are two cold oscillations they could represent the Oldest Dryas and Older Dryas, but if only one is present then it would seem more likely to represent the Older Dryas, suggesting the final warm period represents the Bølling. A cautious approach would only identify a single cold oscillation from this stratigraphy, and there is a need for absolute dating to confirm the timing of the events, but it seems likely that this section represents a full European style Allerød-Older Dryas-Bølling sequence, rather than a single "Windermere Interstadial" event.

The reconstructed temperatures from the interstadial are comparable with modern day temperatures at the site and represent full interstadial conditions. This is in keeping with reconstructions from Coleoptera and our understanding of the flora at the time. The bulked sample covering the onset of the period suggests peak temperatures at this time, in contrary to the qualitative interpretation of the data but more in line with the majority of reconstructions from Europe that suggest an early peak to interstadial temperatures. The gradual decline in temperatures fits with the majority of records from Europe and Greenland, and the slight deterioration could represent the Older Dryas.

The transition from the Windermere Interstadial to the Loch Lomond Re-advance has been shown to be particularly rapid in ice core records from Greenland (Johnsen *et al.*, 1992) and central Europe (Mangerud, 1987). The presence of an intermediate fauna in the period of decline at the beginning of the Younger Dryas suggests a slower onset than is present in these records. Given the sensitivity of the record at other periods (e.g. the pre-Boreal oscillations), this probably reflects a delayed temperature depression rather than a lag in the response of the population. This is in keeping with the conclusions of Bedford *et al.* (2004) that suggest the transition had a defined faunal zone. There are some suggestions in the literature of two distinct stages during the Younger Dryas, the initial very cold and dry climate, followed by a warmer period (Walker, 1995). This is reflected to some extent in this stratigraphy, where there is evidence for an amelioration of temperature in the second half of the stage.

The reconstructed temperature for the Younger Dryas agrees well with other British Isles temperature reconstructions, being slightly lower than that inferred by pollen but consistent with other proxies. Any detailed variation within the zone is obscured by the need to bulk samples, but the T_j of 9.98°c from the transfer function corresponds well with other Chironomid temperature reconstructions from the British

Isles (Brooks *et al.*, 2007; Bedford *et al.*, 2004) as well as coleopteran studies (Atkinson *et al.*, 1987) and reconstructions based on relic periglacial features (Isarin *et al.*, 1998). This is considerably higher than temperatures inferred from Scottish glacier reconstructions (Ballantyne *et al.*, 2007), although the southerly latitude of the site may account for some of the difference. Climate modelling of the period comes to similar conclusions regarding the temperature depression, assuming a shutdown of the MOC.

The early Holocene lacustrine sediments show two distinct cooling events which may well correlate with the early Holocene climate oscillations. These oscillations are recognised across much of the palaeoartic (Björk *et al.*, 1997b; Lotter *et al.* 1992) and date to 8.2ka and 11.2ka (Hoek & Bos, 2007). They have also been recognised in the Chironomid record presented by Bedford *et al.* (2004). Recent research has proposed two different mechanisms for their causes, namely a freshwater pulse for the 8.2ka event (Alley & Ágústsdóttir, 2005) and decreased solar forcing for the 11.2ka event (Björk *et al.*, 1998b).

The reconstructed temperatures for these events are 13.5 ºc and 15ºc for the 11.2ka and 8.2ka event respectively. The ~1ºc temperature depression during the 8.2ka event seems plausible when compared to evidence from France (Alley *et al.*, 1997). The depression lasted ~160 yrs. in the Greenland ice cores (Thomas *et al.*, 2007), but only 37 years in an exceptionally well dated speleothem from Ireland (Baldini *et al.,* 2002). To establish the length of the response in this record a very high resolution dating framework would need to be constructed for this core, given its short length. The agreement of this Chironomid stratigraphy with other records of the 8.2ka event directly contradicts Kurek *et al.* (2004), who conclude that midge inferred temperature reconstructions lack the sensitivity to pick up the 8.2ka event, although it should be noted that the change is within the sample specific errors of the reconstruction.

The high summer temperature (16.59ºc) at the start of the zone may be anomalously high given other Chironomid based reconstructions that use similar training sets (Brooks & Birks, 2000; Bedford *et al.*, 2004), although the temperature lies well within the range suggested from coleopteran proxies for mean summer temperatures (Atkinson *et al.*, 1987). The limits of the temperature range of the training set used in the transfer function must be considered when assessing the accuracy of this value because of the possibility of a "no analogue" situation, although the methods used do not preclude an accurate prediction beyond the sample range in the study.

Chironomid populations are known to be responsive to other variables, particularly trophic status, but this stratigraphy does not offer much in the way of information regarding the trophic status other than that the significant population of *Microspectra* throughout the Younger Dryas indicates oligotrophic conditions throughout the stadial. The response of *Einfeldia pagana,* a cold stenotherm and an indicator of meso/eutrophic conditions indicates that this particular species is responding mainly to trophic status. Figure 7.7 summarises the temperature changes described in this chapter.
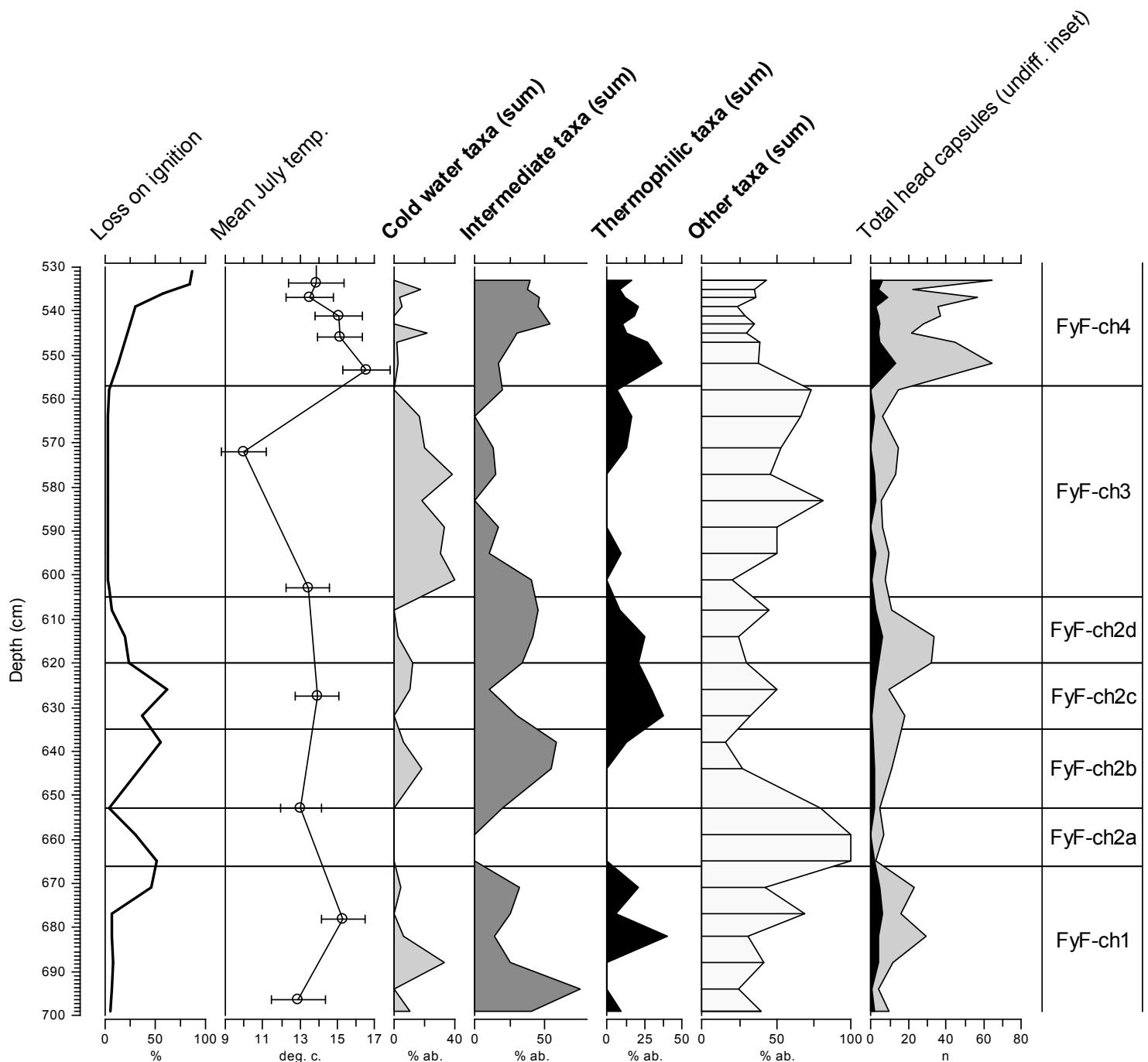
Figure 7.7. Stratigraphic diagram of the key changes in the chironomid population. The temperature transfer function results are included.

Chapter 8: Precipitation Reconstruction

8.1 How can temperatures help reconstruct precipitation?

One direct application of the Chironomid based temperature reconstruction for the Younger Dryas is to help enlighten discussions around the precipitation during the Younger Dryas (as discussed in chapter 2).

Using measurements from modern glaciers it has been shown that there is a relationship between precipitation and temperature at the equilibrium line altitude (ELA) on a glacier (Ohmura *et al.*, 1992). If either of these values is known, the other can be calculated. Because the study site is so close to the glacier the temperature at the ELA for the glacier can be calculated with good confidence, enabling the estimation of precipitation.

Hughes (2002) described the geomorphological and sedimentary evidence for a localised cirque glacier on the north-east side of Arenig Fawr, approximately 1km from the study site (but hydrologically separated by an igneous outcrop). The ELA for this glacier has been calculated using statistical relationships between form and ELA from modern analogues (see Ballantyne, 2007; Carr & Coleman, 2007). The method is a statistical regression between the temperature at the ELA and the shape and size of a glacier. The site of the glacier and the evidence of former glaciation is given in appendix b.

8.2 Application to Glacier Models

The July average temperature (T_j) from the Chironomid based reconstructions (9.98+/-1.21°c @ 364m) was converted to the June-July-August average (T_{max}) at the equilibrium line altitude (ELA, 511m) using an environmental lapse rate of 0.6°c/100m and a T_j to T_{max} conversion factor of 0.97, as per Ballantyne *et al.* (2007). The result is a reconstructed T_{max} at the ELA of 8.76°c. Using the Ohmura *et al.*(1992) regression the precipitation at the ELA is 3927^{+525}_{-561}mm/a. The more recent "degree-day model" (Hughes & Braithwaite, 2008) was also applied, assuming a sinusoidal temperature distribution and mean January temperature (T_{jan}) reconstructions from Renssen & Isarin (1998) cross checked with annual temperature range (T_{range}) inferences from Isarin & Renssen (1999). The result was runoff from the glacier of 4691mm/a water equivalent.

8.3 Discussion

Considering modern average precipitation for the region is between 2200-4700mm/a (Met Office

2008b), these findings suggest precipitation during the Younger Dryas was similar to modern values (see figure 8.1). It is important to note that Snowdonia is currently one of the wettest areas in Europe; for example, on Snowdon the yearly average is 5030mm/a (recorded at 762m; Lamb, 1972), suggesting the site currently lies at the higher end of this range. This reconstruction is higher than has been inferred from relic periglacial features (Harrisson *et al.,* 2008), climate modelling (e.g. Renssen *et al.,* 2001), and contradicts Dahl & Nesje (1992), who suggested precipitation would be lower because of colder temperatures. The high precipitation at the site may be due to the winter storm track running almost directly over North Wales during the Younger Dryas, depressed in latitude because of extensive winter sea ice cover (Isarin *et al.,* 1998; Brauer *et al.,* 2008). The findings of the Ohmura model could be countered by suggestions a difference between modern seasonality and that during the Younger Dryas (e.g. Denton *et al.,* 2005), thus a limited ablation season – this would cause the assumptions of a modern yearly distribution of temperatures assumed by the model to cause the model to overestimate precipitation. The "degree day" model incorporates temperature range reconstructions to account for any changes in seasonality, and produces a higher estimate. However it still assumes a sinusoidal yearly temperature distribution - if the temperature distribution during the Younger Dryas was skewed (e.g. Denton *et al.,* 2005), this estimate could be erroneous. It should also be bourn in mind that the two estimates are not directly comparable because the Ohmura regression estimates precipitation, whereas the "degree-day" model estimates runoff.
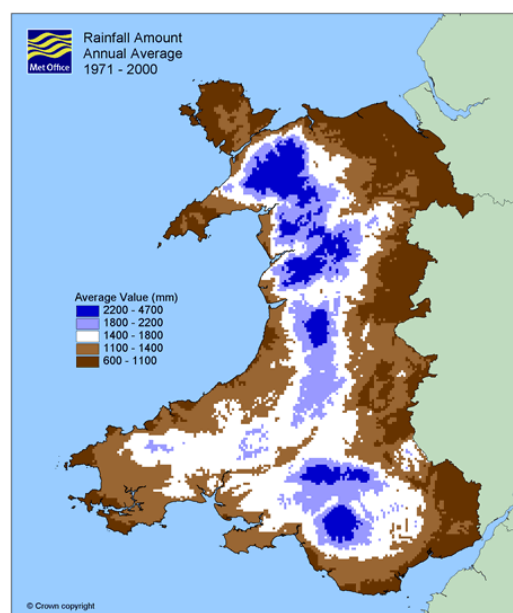

Figure 8.1. Modern rainfall in Wales. From Met Office 2008b.

Chapter 9: Conclusions and directions for future research

9.1 A Proposal for Expanding the Study

The study site has been shown to be sensitive to past climate change, but there are two main limitations to the study - the first is the limited dating framework. Whilst the main phases of the Late Glacial have been picked out, individual events duration cannot be quantified and changes in the rate of sedimentation cannot be assessed. Further studies at the site would benefit from a robust absolute dating framework such as ^{14}C dating; the local geology lends itself to traditional bulk organic matter methods, although recent research suggests that Chironomid remains themselves may provide the best material for nuclear isotope dating techniques (Langdon *et al.,* 2007). This could go some way to account for the burrowing activity of some species, the impact of which has not yet been assessed or accounted for in any palaeolimnological studies, although some species are known to burrow up to 30cm (Pinder, 1980). A robust dating framework would allow better comparisons with the existing literature. The search and identification of tephra horizons, particularly the "Vedde Ash" (Lowe & Turney, 1997) would also be beneficial in understanding the rate of sedimentation.

Secondly, the clays proved difficult to extract using a Russian type corer, so a narrower gauge auger was used providing less material for analysis. Although Chironomid based reconstructions use considerably less material than other insect based methods, the low Chironomid densities encountered limited the resolution of this study. In order to fully resolve the shorter Chironomid population changes throughout the core a larger volume of sediment needs to be extracted and sampled at a higher resolution than this study; this would necessitate the use of a piston type corer (e.g. Livingstone type), or ideally an excavation. This would increase the number of head capsules extracted and thus reduce the need to bulk samples for use with the transfer function.

The Chironomid record can only be used to recreate mean July temperatures, which whilst making it useful in particular studies would be of greater use if used in conjunction with other records. Further study would benefit from a multi-proxy approach starting with a high resolution pollen & macrofossil analysis from the same material as the Chironomid record. This would assist in comparing the core to other local studies, and because pollen could be extracted from the wash from the Chironomid preparation, allow direct comparison of the two. A comparison with coleopteran assemblages would allow a direct comparison of the two, along with an evaluation of seasonality at the site. Further investigation is likely to yield further insight into climate change throughout the late glacial and into much of the early Holocene.

9.2 What was the nature of change during the Windermere Interstadial in North Wales?

Conditions before the rapid onset of the interstadial are difficult to assess because of low count sums, but there is potential for a clearer record should more material be extracted from this site. Temperatures peaked early during the interstadial, and seem to have been roughly analogous with modern day temperatures, although at the peak of the interstadial they may have been slightly higher. The slow decline of temperatures seen in most other records is not obvious from the Chironomid stratigraphy but despite the coarse resolution the transfer function does reflect this slow decline.

The presence of a cold oscillation during the interstadial in this reconstruction seems likely from both the changes in the Chironomid fauna and the results of the temperature transfer function, both of which show a small downturn in temperatures. This would seem likely to correlate to the Older Dryas (Greenland interstadial GI2b), the largest change observed across the continent and in Greenland at this time. The multiple oscillations observed by other Chironomid workers (Bedford *et al.,* 2004; Brooks *et al.* 1997) are lent very limited support from these results, where there are unstable episodes where small changes may have been obscured by the sample resolution. There is considerable potential for further study of the site to assess the variability of the site through the Windermere Interstadial.

The temperature reconstructions fit with other records, and clearly demonstrate the lag of the pollen record when compared to the pollen record of Hughes (2000). They lie well within the ranges of the Coleoptera inferred temperature presented by Atkinson *et al.* (1987), although are slightly lower than the temperatures inferred by Coope *et al.* (1998), even when adjusted for altitude.

9.3 What were the changes in temperature, seasonality and precipitation during the Loch Lomond Stadial?

The rapid onset of the Younger Dryas is not unexpected given the considerable literature on the exceptional rapidity of the onset. The early return of some thermophilic taxa suggests that the return of milder conditions in the second half of the stadial suggested by other workers (e.g. Isarin *et al.* 1998) may hold some truth. The early Holocene warming is particularly obvious although some species seem largely unaffected, which is surprising given the widespread changes in the general ecology of the lake at the time. The temperature reconstruction for this period was hampered by low count sums, but the estimate obtained is very similar to the results from a wide variety of sources. The tempera-

tures are considerably colder than those proposed by Isarin & Bohncke (1999), who suggest minimum July temperatures of around 12ºc for the first part of the Younger Dryas. They are well within the range of the Coleoptera studies, and agree with the Coleoptera inferred temperatures in Coope *et al.* 1998.

Previous attempts to use glacier reconstruction models to calculate precipitation during the Younger Dryas were often hampered by the necessity to use temperature reconstructions from sites that were geographically removed enough as to leave doubt as to their relevance. The reinterpretation of Hughes (2002) through the use of the temperature reconstruction can leave little doubt as to the applicability of the temperature reconstruction given the two sites are less than 1km apart. This reinforces the original conclusions of Hughes (2002) that precipitation during the Younger Dryas was comparable to modern precipitation. This supports the view of many glaciologists, and counters that of other workers. The solution may not lie in who is right and who is wrong, but in how the differing estimates can be settled with each other – changes in seasonality during the Younger Dryas may hold the answer, with longer winters being the prime candidate.

9.4 Implications for current climate

A study of this nature is not directly concerned with the current climate, but even a small study of this nature has some implications for the current research on future climate change. The exact cause (or causes) of the Younger Dryas are still the subject of considerable debate, but the majority of theories relate in some way to a proposed instability of the MOC (as discussed in chapter 2). This has important implications for modern climate, as small changes in the strength of course of these waters have the potential to cause wide ranging temperature changes all around the Atlantic seaboard. It is also important to understand the threshold response of the system, as although theories of a future catastrophic switch are no longer well accepted, understanding the behaviour and effects of the system under extreme conditions allows workers (particularly climate modellers) an insight into the system. Even relatively small temperature variations can have far reaching impacts for ecology and society, and our ability to predict and mitigate these effects is reliant on a broad, detailed literature on past climate change (Jansen *et al.*, 2007).

9.5 How can Chironomid proxies add to our understanding of the Late-Glacial?

This study has increased the resolution of the temperature record for the late glacial changes in the British Isles, but also demonstrated the reliability of the method compared to other proxies. The possibility of the 8.2 and 11.2ka events in this stratigraphy could indicate an important avenue for future use, as the early Holocene would benefit from better proxy records (Hoek & Bos, 2007). The identification of these events using chironomids would contradict Kurek *et al.* (2004) who argue the proxy is not sensitive enough. Chironomids have been shown in this study, like others, to be sensitive indicators of climate change, and by comparison to other records particularly accurate. The precision of the transfer function is good, but a considerable number of capsules need to be retrieved and identified before accurate reconstructions can be made. Chironomid analysis is particularly time consuming, and there may be scope for developing new methods to reduce the time it takes to extract and prepare specimens.

Given the likely heightening of debate regarding possible changes in seasonality during the Younger Dryas, the ability to infer summer temperatures using chironomids will be of increasing importance. Their use in reconstructing Younger Dryas summer temperatures has widespread potential in the study of relic glacial and periglacial landforms because they can provide accurate local estimates of temperature if appropriate sites can be located. The use of subfossil chironomid remains in providing material for both nuclear isotope dating (Langdon et al., 2007) and in stable oxygen isotope analysis (Wang et al., 2008), allowing reconstructions of mean annual temperature (e.g. Wooler et al., 2004) are both interesting developments, and it seems likely that chironomid based methods will continue to play an increasing role in limnology and palaeoecological studies.

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Appendix a: A sample chironomid count sheet

Sample ID: A58 **Slide no.: 1 (left cover)**

Tribe	Species	Type	Condition	Size	Whole?	Notes
	Microtendipes	pedellus	g	m	y	
	Microtendipes	pedellus	m	m	h	
Tanypodinae						
	Einfeldia	pagana	m	s	y	
	Microtendipes	pedellus	m	s	h	
Chironomini			m	m	h	
	Glyptotendipes	pallens	p	m	y	
?			p	m	y	poss. Microtendipes?
	Microtendipes	pedellus	g	m	h	

Appendix b: The glacial evidence from the Llyn Arenig Fawr area.

a)   Diagram from Hughes, 2002.
b)   Digital 2½-D satellite image overlain on a DEM of the Llyn Arenig Fawr, from Google, 2008
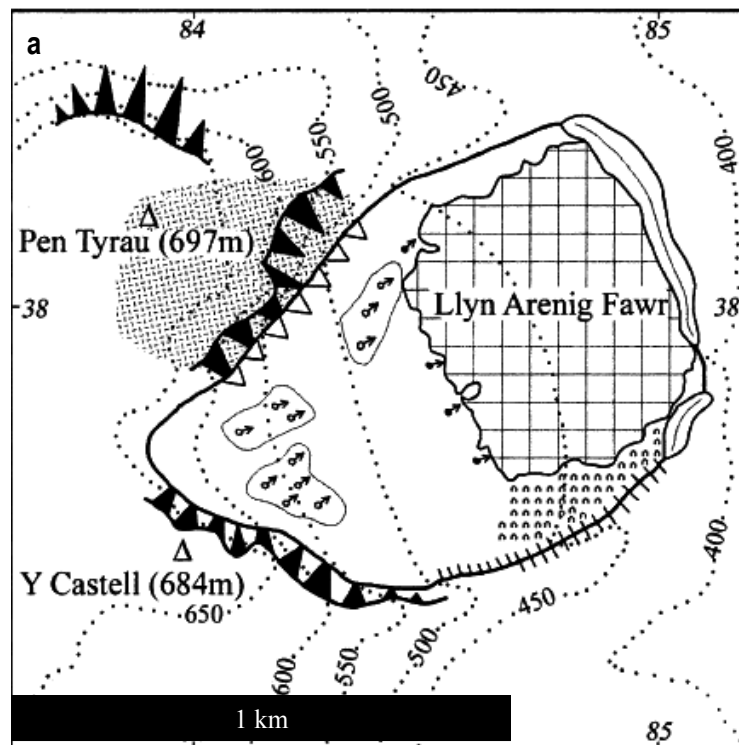
KEY			
/ Interpolated glacier limit	Relict periglacial features	Ice-smoothed bedrock (showing striae direction)	Lake
Drift limit	End or lateral moraine (with crests depicted)	Till fabric station (showing flow direction)	**P** Pollen site
Boulder limit	Hummocky moraine	Meltwater channel	50m land and former glacier surface contours
Periglacial trimline	Fluted moraine	Steep cliffs	Δ Mountain summit

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Appendix c. A full list of taxa encountered in the study

Including generic groups and statistics.

	Number of samples presence	Hill's n2 diversity index	Mean value	Standard Deviation
Undiff. *chironomid*	30	3.80763		17.9722
Undiff. *tanypodinae*	22	3.67517	4.29412	12.3349
Thienemannimyia	2	2	0.058824	0.235294
paramerina	2	2	0.117647	0.470588
Procladius	2	2	0.058824	0.235294
Ablabesmyia	2	2	0.058824	0.235294
Undiff. *chironomini*	15	3.6608	1.5	4.31822
Microtendipes pedellus	29	3.74627	10.7353	30.5073
Endochironomus impar	3	2.46154	0.235294	0.842225
Endochironomus albipennis	6	2.57143	0.882353	3.08473
Einfeldia pagana	7	3.22581	0.588235	1.81688
Glptotendipes pallens	4	2.77778	0.294118	0.986062
Glyptotendipes barbipes	6	3.16129	0.411765	1.28607
Glyptotendipes severini	6	3.33333	0.294118	0.892103
Sergentia	3	2.57143	0.176471	0.616946
Chironomus plumosus	4	3	0.176471	0.567274
Dicrotendipes	3	2.66667	0.117647	0.403274
Dicrotendipes nervosus	2	2	0.058824	0.235294
Omisus	3	2.66667	0.117647	0.403274
Paratendipes nudisquama	2	2	0.058824	0.235294
Chironomus anthracinus	4	3	0.176471	0.567274
Polypedilum sordens	10	3.46154	0.882353	2.62078
Polypedilum nubeculosum	17	3.69515	2.35294	6.73831
Undiff. *Tanytarsini*	19	3.66369	1.88235	5.41656
Tanytarsus	6	3.33333	0.294118	0.892103
Microspectra	4	2.90909	0.235294	0.769217
Tanytarsus lactescens	2	2	0.058824	0.235294
Microspectra radialis	10	3.44898	0.764706	2.27595
Microspectra pallidula	2	2	0.058824	0.235294
Microspecta insignilobus	2	2	0.058824	0.235294
Propsilocerus	2	2	0.058824	0.235294
Paratanytarsus	5	3.2	0.235294	0.729981
constempellina	5	3.2	0.235294	0.729981
Tanytarsus pallidicornis	7	3.42857	0.352941	1.05391
Tanytarsus lugens	2	2	0.058824	0.235294
Pseudochironomus	11	3.53922	1.11765	3.27885
Undiff. *Orthocladiinae*	16	3.68345	1.41176	4.05018
Pseudorthocladius	6	3.27273	0.352941	1.08146
Heterotrissocladius maeaeri	2	2	0.058824	0.235294
Psectrocladius	3	2.66667	0.117647	0.403274
Cricotopus intersectus	3	2.66667	0.117647	0.403274
Cricotopus bicinctus	2	2	0.058824	0.235294
Corynoneura artica	2	2	0.058824	0.235294
Paracricotopus	2	2	0.058824	0.235294
Eukiefferiella fittkaui	2	2	0.058824	0.235294
Undiff. *Diamesinae*	3	2.66667	0.058824	0.201637
Psectrocladius sordidellus	15	3.64939	2.61765	7.54892
Psectrocladius calcaratus	5	3.2	0.235294	0.729981
Psectrocladius barbatipes	3	2.66667	0.117647	0.403274
Pseudodiamesa	3	2.66667	0.117647	0.403274
Pseudosmittia	2	2	0.058824	0.235294

Appendix d: PCA analysis results and notes on the method.

All taxonomic groups were included in the PCA. PCA was performed using PAST (Hammer *et al.* 2001), although the inclusion of only group level taxonomic ranks does improve the scores.
The top figure shows the PCA for samples, showing that the populations in each sample are responding predominantly to the first axis. The bottom figure shows the drastic reduction in eigenvalue% for each subsequent axis after the first.

PCA Scores (samples)