Oxygen isotopic evidence for high-magnitude, abrupt climatic events during the Lateglacial Interstadial in northwest Europe: Analysis of a lacustrine sequence from the site of Tirinie, Scottish Highlands

Candy, I., Abrook, A., Elliot, F., Lincoln, P., Matthews, I.P. and Palmer, A.

Department of Geography, Royal Holloway, University of London, Egham, Surrey, TW20 0EX, UK

Corresponding author: ian.candy@rhul.ac.uk, 01784 443649

Abstract

The Last Glacial to Interglacial Transition (LGIT) is a period of climatic instability. $\delta^{18}O$ records are ideal for investigating the LGIT as this proxy responds rapidly to even minor climatic oscillations. Lacustrine carbonates offer the opportunity to investigate spatial diversity in patterns of climatic change during the LGIT but this requires the generation of $\delta^{18}O$ records from a range of latitudinal and longitudinal settings. This study presents a coupled pollen and stable isotopic study of lacustrine carbonates spanning the Windermere Interstadial (the British equivalent of GI1, the Lateglacial Interstadial) from the site of Tirinie in the Scottish Highlands, a region where $\delta^{18}O$ records are currently absent. The Interstadial is characterised by three $\delta^{18}O$ peaks, warm intervals, and two $\delta^{18}O$ declines, cold episodes, the timing of which is constrained by the presence of crypto-tephra. The landscape at Tirinie was highly responsive to these climatic oscillations as the sedimentary and pollen record respond to each isotopic shift. The paper concludes by highlighting that, across the British Isles, lacustrine $\delta^{18}O$ records of the Interstadial have a consistent stratigraphy/structure, however, the magnitude of the isotopic shifts are regionally variable. Potential causes of this variability are discussed.

Keywords

$\delta^{18}O$, lacustrine carbonates, Lateglacial Interstadial, British Isles, Abrupt Change

1 Introduction

Oxygen isotopic ($\delta^{18}O$) analysis of continental archives has dramatically improved our understanding of abrupt climatic events in the Late Quaternary (Lotter et al., 1992; von Grafenstein et al., 1999; 2000; Marshall et al., 2002; Daley et al., 2011). Although the existence of abrupt events, particularly during the Last Glacial to Interglacial Transition (LGIT), has been identified through the application of palaeoecological proxies (Atkinson et al., 1987; Mayle et al., 1999; Walker et al., 1993, 2003; Lang et al., 2010; Brooks et al., 2012), $\delta^{18}O$-based studies offer two key advances. Firstly, $\delta^{18}O$ analysis can be carried out on very small samples allowing a resolution of reconstruction that is frequently impossible with many fossil based techniques (Lotter et al., 1992). Secondly, most $\delta^{18}O$ archives record the $\delta^{18}O$ of meteoric waters which is, through the observable relationship between the $\delta^{18}O$ of rainfall and prevailing air temperature (Dansgaard, 1964; Rozanski et al., 1992; 1993), highly responsive to short-lived temperature fluctuations (Marshall et al., 2002; 2007; Daley et al., 2011). Consequently, in archives such as the Greenland ice cores (GRIP, 1993; Rasmussen et al., 2006; 2008), the high-resolution analysis of $\delta^{18}O$ has allowed, not only the identification and characterisation of the interstadial/stadial oscillation that occurred during the LGIT (Greenland Interstadial 1 (GI-1)/ Greenland Stadial 1 (GS-1), known in continental Europe as the Bølling/Allerød and Younger Dryas periods respectively) but also the identification of shorter-term climatic oscillations during the Interstadial (GI-1e, c and a, which are ameliorations, and GI-1d and b, which are deteriorations).

In western and central Europe palaeoecological techniques (Atkinson et al., 1987) have been the prime basis for the reconstruction of the climatic stratigraphy of the LGIT, however, recent research on the $\delta^{18}O$ analysis of lacustrine carbonates has greatly enhanced this work (Lotter et al., 1992; Whittington et al., 1996; von Grafenstein et al., 1999; 2000; Marshall et al., 2002; Van Asch et al., 2012). In the British Isles, for example, high-precision $\delta^{18}O$ analysis of three LGIT marl sequences in western Britain and Ireland have shown a climatic stratigraphy, during the Lateglacial Interstadial (equivalent to the Windermere Interstadial of Britain and used in this paper to refer to such sequences in Britain and Ireland), comparable to that of the Greenland ice cores (Marshall et al., 2002; Diefendorf et al., 2006; Van Asch et al., 2012). These records show that the Lateglacial Interstadial comprises three warm phases separated by two periods of abrupt cooling (Van Asch et al., 2012). These have been tentatively related to GI-1e, 1c and 1a and G1-Id and 1b, respectively, although the chronology of these lacustrine sequences is not currently sufficient to discuss the a/synchroneity of these events (Marshall et al., 2002; Diefendorf et al., 2006; Van Asch et al., 2012). These three records show a consistent pattern in their $\delta^{18}O$ stratigraphy, however, this is perhaps unsurprising as they are all found in similar geographic/topographic/climatic settings. All three are: 1) lowland sites, less than 20 metres above sea level, 2) at a very similar latitude (53-54ºN) and 3) in close proximity to the coast. The likely drivers of abrupt change during the LGIT, such as ocean circulation (Broecker et al., 1989), may, however, produce strong regional gradients in the magnitude of climatic response, both latitudinally and longitudinally. In order to understand regional variations in environmental response to abrupt climate change there is, therefore, a need to generate further LGIT δ^{18}O records from sites in a diverse range of locations from across western Europe.

In this study we present the litho-, tephro-, pollen- and oxygen isotope stratigraphy of a lacustrine carbonate sequence spanning the Lateglacial Interstadial from the site of Tirinie in the southeast Grampian highlands, Scotland (Lowe and Walker, 1977). This site represents a key locality in understanding regional variations in the expression of LGIT climate change in that; 1) it is one of the most northerly δ^{18}O records of Lateglacial climate change anywhere in Europe (see Whittington et al., 2015), and 2) it is the only relatively high-altitude (320 metres a.s.l.) δ^{18}O record in western Europe. The pollen stratigraphy of this sequence confirms its attribution to the Lateglacial Interstadial whilst the occurrence of two discrete crypto-tephra layers of known age anchor the timing of the deposition of this unit to key points within the Lateglacial. The δ^{18}O record derived from the Tirinie sequence appears to be unaffected by detrital contamination and is interpreted as reflecting climate, primarily temperature, variability across the Interstadial. This locality would appear to be highly sensitive to climate change during the LGIT as both the sedimentology and the pollen record respond in step with the δ^{18}O signal. The data presented here allows two key observations to be made. Firstly, that the δ^{18}O record of the Lateglacial Interstadial constructed from the Tirinie sequence shows a stratigraphy that is consistent with that recorded in other lacustrine δ^{18}O records from Britain and from the δ^{18}O signal of the Greenland ice cores. Secondly, that the magnitude of the δ^{18}O oscillations that occur in the Tirinie sequence are significantly greater than those seen in any other British δ^{18}O record of the Lateglacial Interstadial and are, in some cases, comparable with the isotopic shifts observed within the Greenland ice cores. The paper concludes by highlighting the fact that whilst there appears to be good regional consistency in the δ^{18}O stratigraphy of the Lateglacial Interstadial, the magnitude of the climatic events that occurred during this interval are likely to have varied significantly across western Europe.

2 Background

The site of Tirinie (Figure 1) in Glen Fender, Perthshire is in the south-east Grampian Highlands (NN/889768; 56º46’23.09”N, 3º48’21/04”W) approximately 10 miles to the north-west of Pitlochry (Lowe and Walker, 1977). The sediment sequence presented here is found in a small south-west/north-east trending elongate palaeo-lake basin which is approximately 4 metres in depth (Figure 1). The topographic low, which is currently the physical expression of the basin, is an active mire, fed by a spring to the northeast and drained by a stream to the southwest. The surface area of the mire, which represents the surface area of the palaeo-lake, is ca 3,500 m^2, however, the basin is fed by a relatively extensive catchment, most of which occurs in the hills to the north, with an area of ca 454,500 m^2, some 120 times greater than the area of the lake basin (Figure 1). The basin itself is a former kettle hole which formed during the wastage and stagnation of the Late Devensian Ice Sheet (DIS) (Rose and Smith, 2008), the basin lies beyond the limits of the Loch Lomond Readvance ice sheets that grew during the Loch Lomond Stadial (the British correlative of the Younger Dryas and Greenland Stadial 1). The bedrock that underlies the site is Schiehallon Quartzite Formation, whilst the peaks of the neighbouring hills and the flanks of the surrounding slopes are comprised of Blair Atholl Dark Limestone and Dark Schist Formation (BGS Geoindex, 2013). It is not known whether the local Quaternary glacial deposits contain a significant carbonate content.

Lowe and Walker (1977) conducted the only previous study of the Tirinie record. The lithostratigraphy comprises a tripartite sequence, typical of many lateglacial sequences that have accumulated within sedimentary basins that formed during deglaciation of the LDIS but occur beyond the limits of the main Loch Lomond Readvance ice-caps (Lowe and Walker, 1977). The succession, which is underlain by glacial gravels, comprises two units of organic-rich lake marls separated by a minerogenic unit of silts and clays (Lowe and Walker, 1977). The whole sequence is capped by the accumulation of wood peat. This sequence was interpreted as reflecting a warm-cold-warm oscillation with marl accumulation, reflecting high biological productivity in a lake basin, occurring under warm conditions punctuated by the accumulation of minerogenic sediments under cold conditions, reflecting low productivity in the lake and reduced vegetation cover generating increased soil erosion and allogenic input.

This interpretation is supported by the pollen assemblage from this sequence with the lowermost marl unit being characterised by a *Betula*, *Juniperus* and *Empetrum* pollen assemblage, the minerogenic clays and silts being characterised by an increase in *Artemisa* and *Rumex* at the expense of *Betula* and *Juniperus* and the uppermost marl being characterised by *Betula*, *Juniperus*, *Empertrum* (see Lowe and Walker, 1977 for details). The first appearance of *Corylus and Ulmus* occurs in the final stages of the accumulation of the uppermost marl. The correlation of these phases of vegetation history with the regional pollen assemblage zones would suggested that: 1) the lowermost marl unit was deposited during the Windermere Interstadial, 2) the unit of minerogenic clays and silts was deposited during the Loch Lomond Stadial, and 3) the uppermost unit of marls, and the overlying peats, were deposited in the earliest Flandrian/Holocene (see Lowe and Walker, 1977; Lowe et al., 1994; 2008; Mayle et al., 1999). While Tirinie is only one of a large number of LGIT pollen records from Scotland it is one of only a small number that are from carbonate-rich sediments and permits a comparison between vegetation response to climatic forcing through paired isotopic and palynological samples.

3 Methodology

3.1 Core recovery and sedimentology

An auger survey of the Tirinie basin was carried out in April 2012 to map the bathymetry of the basin. The most stratigraphically expanded tri-partite sequence encountered was then sampled using a 1 metre long, 50 mm diameter Russian core. Penetration of the underlying glacial gravel was not possible and the length of the core head, 80 mm, meant that the lowermost part of the succession was not sampled. The recovered sediments were described using the Troels-Smith classification scheme (Troels-Smith, 1955). Carbonate content was measured using a Bascomb Calcimeter which calculates the percentage composition of carbonate in the sample through measuring the amount of CO_2 liberated when the sample reacts with HCl. The mineralogy of the carbonate rich levels was established using whole rock XRD analysis.

3.2 Pollen extraction and preparation

Pollen samples were prepared at 1 cm resolution using a 1 cm^3 volumetric sampler. Following sub-sampling, samples were deflocculated in sodium pyrophosphate ($Na_4P_2O_7$) with the addition of exotic *Lycopodium* tablets to allow for the estimation of pollen concentration. The deflocculated samples were then sieved under a 125µm sieve and a 10µm mesh. Additional sample pre-treatment involved the addition of 10% Hydrochloric acid (HCL) to a maximum of 5ml and Erdtman's acetolysis using a ratio of 9:1 Acetic anhydride ($[CH_3CO]_2O$) and Sulphuric acid (H_2SO_4). These two procedures remove any calcium carbonate and biological cellulose respectively. Prior to the acetolysis stage, the samples were floated with 5ml of Sodium polytungstate (SPT) at a specific gravity of 2.0g/cm^3 to separate the pollen from the sediment. The resultant palynomorph materials were mounted onto slides with the addition of glycerol-jelly.

In keeping with standard practices for Lateglacial pollen counting (Walker, 1975; Lowe and Walker, 1977) a minimum counting sum of 300 total land pollen (TLP), excluding aquatics and spores, was achieved in all but 8 levels (347-354cm). Within these stratigraphically lower levels a sum of 100 TLP was attained due to low pollen concentrations. An assessment of pollen concentration required the counting of exotic spores and the division of the number of fossil pollen grains counted by the number of exotic pollen grains counter multiplied by the number of exotic pollen grains added. Pollen identification was undertaken using an Olympus CX41 binocular microscope at 400x magnification.

For assistance in pollen identification the Royal Holloway reference collection was consulted along with Moore et al., (1991), Reille (1992) and Punt et al., (2007). All pollen diagrams were constructed using the C2 palaeoenvironmental programme (Juggins, 2007).

3.3 Tephra analysis

Tephra extraction followed the stepped flotation procedure adopted by Turney et al. (1997a); Turney (1998) with the modifications of Blockley et al. (2005). Contiguous 5cm scan samples throughout the sequence were ashed at 550°C for two hours, and immersed in 10% HCl to remove organics and carbonates respectively. The samples were then wet sieved between 80-15µm and floated in sodium polytungstate between densities of 2.0-2.5g cm^{-3} to remove the majority of biogenic and minerogenic detritus. The floated samples were mounted onto slides using Canada balsam heated to ca.85°C and examined using an Olympus CX41 binocular microscope at 100x-400x magnification. In samples where tephra shards were identified, the sequence was contiguously resampled at 1cm intervals to more accurately refine the stratigraphic shard profile and precisely locate peaks in shard abundances. Tephra shard peaks were subsampled at 1cm resolution and prepared for single grain geochemical analysis. Individual shards were picked onto an epoxy resin stub using micro-manipulation techniques, covered, and then polished to expose shard surfaces. The samples were analysed on a Cameca SX-100 wavelength-dispersive electron microprobe (WDS-EPMA) using calibrated primary and secondary standards at the University of Edinburgh's Tephra Analytical Unit. Analyses were performed using a voltage of 15 keV, an operational beam diameter of 5µm and a beam current of 2 nA for Na, Al, Si, Fe, K, Ca and Mg and 80 nA for F, Mn, Cl, P, S, and Ti, standard data associated with these analyses are presented in supplementary data table 1.

3.4 Stable isotope analysis

Material for $\delta^{18}O$ and $\delta^{13}C$ analysis were sampled at 5 mm intervals by cutting out 5mm x 5mm x 5mm cubes of sediment. These bulk samples were disaggregated using sodium hexametaphopshate and then sieved over a 63µm mesh with the <63µm fraction being collected and treated with hydrogen peroxide to remove organic material (see supplementary information 2). Samples for isotopic analysis were weighed using a Cahn C-31 Microbalance (400 – 600µg). The stable $\delta^{18}O$ and $\delta^{13}C$ values were measured by analysing CO_2 liberated from sample reaction with phosphoric acid at 90°C. Internal (RHBNC-PRISM) and external (NBS-19, LSVEC) standards were run every 10 samples. The carbonate stable isotopes were analysed using a VG PRISM series 2 mass spectrometer. All stable isotopic values are quoted with reference to VPDB. Internal precision produces analytical uncertainties of +/-0.07 ($\delta^{18}O$) and +/-0.04 ($\delta^{13}C$).

4 Results

4.1 Lithostratigraphy

The recovered sequence closely matches that of Lowe and Walker (1977), characterised, as it is, by two units of marl separated by a unit of minerogenic silts and clays (Figure 2a). The detailed stratigraphy of the lowermost marl unit (Figure 2b) is relatively complex with numerous oscillations in sedimentology occurring, primarily reflecting variations in the relative concentration of carbonate versus minerogenic material. Ten lithological sub-units (TIR-L*n*) can be identified through this unit with sub-units TIR-L2, 4, 6 and 9 being characterised by high $CaCO_3$ concentrations (typically between 60 and 90%) and units TIR-L1, 3, 5, 7 and 10 being characterised by relatively low values (typically between 0 and 20%). The marls consist entirely of calcite, which is fine-grained with apparent but weakly developed laminations. During sieving and binocular microscopic analysis it became apparent that the marl was also texturally homogeneous. The sieved material contained no ostracod valves (J. Holmes pers comm.), calcified stems or chara oogonia. Mollusc shell fragments do occur but are rare and, consequently, the material is, in the main, a fine-grain, pure calcite mud. Even in units TIR-L 1, 3, 5, 7 and 10, where % carbonate is low, the calcite appears to be a uniform fine-grained mud.

4.2 Pollen succession

Six pollen zones were visually identified within the lowermost marl unit (Figure 3) and these are described in detail in Table 1. The pollen assemblage is dominated by *Poaceae*, *Cyperaceae, Betula, Juniperus*, *Empetrum* and *Rumex* although the relative proportion of these vary significantly throughout the marl. In general the boundaries of the six pollen zones correlate with the main shifts in lithostratigraphy that are observable in the record. The pollen-succession is primarily characterised by the expansion/contraction of trees and shrubs relative to herbs. Pollen zones TIR-P1a, TIR-P2/3 and TIR-P5 are characterised by a combined % of trees/shrubs >40%, whilst in pollen zones TIR-P1b, TIR-P4 and TIR-P6 herbs expand to account for >70% of the spectrum. The pollen zones characterised by high % tree/shrub pollen occur within the carbonate rich units, whilst the pollen zones characterised by high % herb pollen occur within the minerogenic sediments.

4.3 Oxygen and Carbon isotopic record

The $\delta^{18}O$ and $\delta^{13}C$ values (Figure 4) show very different trends to each other with no evidence for co-variance between the two datatsets ($R^2 = 0.116$). The $\delta^{13}C$ values are high at the base of the record (2.39‰ at 354.5 cm depth) and decline throughout the marl unit (-3.70‰ at 327.5cm depth). In contrast the $\delta^{18}O$ values oscillate around an average (-7.55‰, 1σ = 0.88). The $\delta^{18}O$ record comprises 3 major peaks in values (TIR-Oe = 352.5 cm, -7.39‰; TIR-Oc = 342.5cm, -6.12‰; TIR-Oa = 328, -

6.82‰) separated by 2 major lows in values (TIR-Od = 351.5cm, -9.03‰; TIR-Ob = 334.5cm, -9.35‰). A decline in $\delta^{18}O$ values occurs during the middle of peak TIR-Oc. Exceptionally low % $CaCO_3$ content between 348.5cm and 350.5cm meant that no meaningful isotopic analysis could be undertaken, producing a gap within the record. The absence of isotopic data between 348.5cm and 351.5cm, therefore, means that the absolute magnitude of the low in $\delta^{18}O$ values that occurs at this level cannot be quantified and the lowest value, -9.03‰, does not necessarily reflect the isotopic minimum of this interval. The shifts between these peaks and lows are relatively large (TIR-Oe to TIR-Od = -2.32‰, TIR-Od to TIR-Oc = +2.98‰, TIR-Oc to TIR-Ob = -3.30‰, TIR-Ob to TIR-Oa = +2.53‰).

4.4 Tephrostratigraphy

Two tephra peaks were located within the Tirinie sequence. One of these, at a depth of 348 cm below surface, occurs within the lower marl in subunit TIR-L3 directly before the transition to subunit TIR-L4. The second occurs at a depth of 305 cm, within the minerogenic-rich unit that lies directly above the lower marl. Reported depths correspond to the level of maximum shard abundance (ca 700 shards/g) but in both cases the peaks are dispersed across 10-20 cm of sediment. The two tephra layers found at Tirinie can be distinguished on the basis of geochemistry and clearly represent the product of two distinct eruptions rather than a single tephra deposit that has been reworked (Figure 2 and 5).

5 Interpretation

5.1 Stratigraphy of the Tirinie sequence

The data presented here confirms the interpretation of Lowe and Walker (1977) that the lowermost marl unit is of Lateglacial Interstadial age. This is based upon three lines of evidence. Firstly, that the marl occurs as part of a characteristic tripartite Lateglacial sequence, separated from deposits of the current interglacial by minerogenic sediments (Lowe and Walker, 1977; Lowe et al., 1994; Mayle et al., 1999). Secondly, that the pollen record of this unit, dominated by *Poaceae, Cyperaceae, Betula, Juniperus* and *Empetrum*, is characteristic of the “Windermere Interstadial” pollen assemblage zone in Scotland and northern England (Lowe and Walker, 1977; Lowe et al., 1994; Mayle et al., 1999). Finally, the two tephra horizons within this sequence allow stratigraphic correlation with the Lateglacial Interstadial (Figure 5). Correlation can be made with known tephra horizons on the basis of their chemistry, which suggests that the lowermost tephra (348 cm below surface) most likely corresponds to the Penifiler tephra and the uppermost tephra (with 305 cm below surface) with the Vedde Ash. These two tephra markers are key chronological indicators because: 1) they occur within distinct positions within the climatic stratigraphy of the Lateglacial in Britain and the North Atlantic region (Matthews et al., 2011), and 2) they both have quantified age estimates associated with them.

The Vedde Ash (12171 ±114 GICC05 yr b2k; Rasmussen et al., 2006; 2008) occurs midway through the Loch Lomond Stadial (Lowe and Turney, 1999; Davies et al., 2004)) and, therefore, implies that the minerogenic unit that overlies the marl correlates to this interval. The Penifiler tephra (13,939 ± 66 μ ± σ cal BP; Bronk Ramsey et al., 2015) occurs in the mid-part of the interstadial, after its onset, i.e. post-Bølling/GI-1e). In terms of climatostratigraphy the Penifilier tephra is found during a climatic amelioration after a major cooling episode within the Interstadial, dated in Scotland to ca 14,000 years ago and thought to equate with the climatic amelioration after GI-1d in Greenland, the Aegelsee Oscillation in Switzerland and the Older Dryas in the north Atlantic region (Matthews et al., 2011). The occurrence of the Penifiler tephra towards the base of the Tirinie sequence, coupled with the pollen stratigraphy, implies that the earliest part of the Interstadial maybe absent from the recovered core, either because stagnant ice persisted in this basin after the onset of warm conditions or that the length of the coring head resulted in the lowermost 8 cm of sediment, which may have contained this evidence, not being recovered.

In this model the lower and upper marls are interpreted as being of Lateglacial Interstadial and early Holocene age respectively whilst the intervening minerogenic unit is interpreted as being of Lateglacial Stadial age. Besides being supported by the pollen and tephro-stratigraphy this model is also consistent with the assumed relationship between marl accumulation and climate. It is widely suggested that marl accumulation is indicative of warm climates as; 1) the precipitation of carbonate in freshwater bodies is biologically mediated and will be promoted under warmer climates (Verrechia, 2007; Palmer et al., 2015; Whittington et al., 2015), and 2) these conditions promote a relatively dense vegetation cover that reduces landscape erosion, decreasing the amount of allogenic material that is supplied to the basin and allowing the accumulation of relatively pure marl (Palmer et al., 2015; Whittington et al., 2015). It is important to note that whilst most lacustrine carbonate records of the LGIT in Britain show evidence for reduced marl production during the Lateglacial Stadial in many cases carbonate precipitation continues across this cold interlude (Marshall et al., 2002; Van Asch et al., 2013). This is not the case at Tirinie where sediments of the Lateglacial Stadial record a carbonate content of 0%, a situation that is seen in other LGIT records from sites in Scotland (Whittington et al., 1996; 2015).

It is likely that the "switching off" of marl production at Tirinie is a combination of changes in both climate and hydrology during the LGIT. That the cold conditions of the Stadial would reduce biological activity and, therefore, decrease rates of carbonate precipitation can be observed in multiple LGIT records (Marshall et al., 2002; Whittington et al., 2015; Van Asch et al., 2013), however, the bedrock geology at Tirinie, with carbonate rich rocks only occurring on the top and flanks of the neighbouring hills, means that groundwater recharge is probably essential in supplying sufficient Ca^{2+} to the lake system to permit calcite precipitation. In such a setting it is the percolation of vadose waters through the carbonate-rich strata that results in the dissolution and uptake of Ca^{2+} by the groundwater that eventually recharges the lake system. By this process the lake waters have high concentrations of dissolved Ca^{2+} making them susceptible to supersaturation and, therefore, calcite precipitation This process will be enhanced during warmer climates, when the soil systems are free-draining, but is likely to cease during the cold conditions of the Stadial when soils are likely to be permanently frozen and impermeable. Therefore, it is proposed that whilst climatic cooling during the Stadial retards the biological activity that promotes carbonate precipitation it is also likely to cause a switch to a surface runoff driven system that reduces the supply of dissolved Ca^{2+} to the lake basin. Both of these processes will reduce the probability of marl being formed. The dominance of overland flow during the Stadial is supported by the mineral-rich sediments deposited during this interval that implies accelerated soil and landscape erosion.

The calcite that has accumulated to make up the Lateglacial Interstadial marl is interpreted as endogenic carbonate, that is to say it reflects fine-grained carbonate that precipitates within the water column in response to biological/photosynthetic changes in water chemistry and consequently settles out of suspension on the floor of the lake basin. In many palustrine/lacustrine environments a range of calcite types can make up the bulk marl material, however, the absence of any macroscopic or microscopic evidence for calcified stems, chara oogonia or osctracods coupled with the fine-grained, homogeneous and weakly laminated nature of the marl would suggest that endogenic carbonate falling out of suspension is the primary origin of the precipitates that comprise this unit.

5.2 Interpretation of the $\delta^{18}O$ and $\delta^{13}C$ signal

Before the palaeoenvironmental significance of the $\delta^{18}O$ and $\delta^{13}C$ signal from the Tirinie Lateglacial Interstadial marl can be discussed it is important to discuss three factors that may affect the way in which this signal is interpreted; 1) the possibility of detrital contamination, 2) the origin of the carbonate being analysed, and 3) the hydrology of the Tirinie palaeolake system. The inwashing of detrital carbonates, i.e. eroded limestone bedrock or carbonate rich Quaternary deposits, can contaminate any isotopic signature derived from the marl as it is impossible to separate allogenic and authigenic carbonate prior to isotopic analysis (Leng et al., 2010). It is proposed that contamination by detrital carbonate is not an issue at Tirinie for three reasons. Firstly, that a number of authors have suggested that isotopic datatsets derived from carbonate subsamples that are affected by varying degrees of detrital contamination typically show a strong degree of co-variance between $\delta^{18}O$ and $\delta^{13}C$ (Leng et al., 2010; Mangili et al., 2010), a pattern that is not seen in the Tirinie dataset ($r^2 = 0.126$, see supplementary information 1). Secondly, that the process of sieving removed any carbonate material >63µm, meaning that the analysed carbonate should be dominated by the fine-grained endogenic fraction. Finally, and probably most significantly, during times of intense allogenic input to the basin, i.e. the minerogenic deposits of the Lateglacial Stadial, the % carbonate content of the falls to zero. This would imply that the Tirinie catchment yields minimal amounts of detrital carbonate than can contaminate the marl isotopic signal. This is an argument that has been used to justify the purity of LGIT marl isotopic records form other LGIT records in Scotland (Whittington et al., 2015).

Establishing the origin of the carbonate that is being analysed is of importance because different freshwater carbonate types, i.e. mollusc shells, ostracod valves, chara stems and oogonia, precipitate at different times of the year with different vital offsets (Leng and Marshall, 2007). Consequently, subsamples of bulk marl sediment that comprise mixtures, in varying proportions, of some or all of these materials can produce isotopic values that are averages of these isotopically diverse carbonates which are, from the point of view of palaeoenvironmental reconstruction, meaningless. The nature of the calcite that comprises the Interstadial marl, homogeneous and fine-grained with no evidence for ostracods or chara remains, indicates that this material is made entirely of endogenic carbonate. Consequently, the resulting $\delta^{18}O$ and $\delta^{13}C$ values record the isotopic composition of carbonate precipitating in the water column and settling out of suspension. Even the presence of occasional mollusc shell fragments has negligible effect on the derived isotopic values as these are removed at the sieving stage.

The hydrology of the Tirinie basin is more difficult to assess as it is no longer an extant lake system and, therefore, rates of recharge and the relationship between the $\delta^{18}O$ of the lake water and that of the prevailing rainfall cannot be quantified. This is important because the use of the $\delta^{18}O$ value of lacustrine carbonates as palaeoclimatic proxies in temperate regions such as Britain is often based on the assumption that the $\delta^{18}O$ value of the lake water closely approximates the $\delta^{18}O$ value of rainfall which is strongly controlled by prevailing air temperature (Marshall et al., 2002; Van Asch et al., 2013). However, processes such as lake water evaporation and the inwash of major snow melt events can cause significant modification of the $\delta^{18}O$ of the lake water, resulting in the $\delta^{18}O$ of the precipitated carbonate recording intra-basin hydrology rather than prevailing climate (Tablot, 1990; Leng and Marshall, 2007). A lake basin such as Tirinie which was characterised by a relatively small volume, i.e. low surface area and shallow depth, has the potential to be particularly prone to such processes.

Despite these characteristics it is argued that such intra-basinal isotopic modification processes had minimal effect on the Tirinie $\delta^{18}O$ record. This is proposed because despite being of relatively small volume the Tirinie palaeolake was fed by a catchment of relatively large surface area (>120 times the surface are of the lake). As the volume of water that is available to recharge a lake basin is strongly related to the size of the catchment area that feeds the basin this is significant as it means that even if evaporation or snowmelt inwash is modifying the isotopic value of the lake water, the lake basin is constantly being recharged by waters that have an "unmodified" isotopic signal. In such a setting it is likely that intra-basin processes have a negligible impact on the $\delta^{18}O$ value of both lake water and the resulting lacustrine carbonates. This suggestion is supported by the fact that the isotopic signal that is derived from the Interstadial marls at Tirinie contains very little noise, that is to say that the shifts between the peaks and troughs in $\delta^{18}O$ are characterised by smooth transitions. A lake system that is effected by significant but highly seasonal variations in evaporation or snow-melt would be expected to show great noise within the data. Furthermore, most lake systems that are strongly effected by evaporation show a strong degree of co-variance between $\delta^{18}O$ and $\delta^{13}C$ values (Talbot, 1990), however, the isotopic dataset from Tirinie shows no such pattern (supplementary information 2). Finally, the possibility that lake levels may have changed during this interval making the system more susceptible to evaporation at different times cannot be discounted. This is highlighted because during the interval of the Stadial and the minerogenic rich levels within the Interstadial aquatics, particularly *Pediastrum*, peak which could reflect lowering of the lake level. Whether lake levels did vary or not the absence of isotopic modification by evaporitic processes is supported by; 1) the absence of co-variance (outlined above) and 2) the fact that during intervals where % aquatics increase, the associated $\delta^{18}O$ values are at their lowest, which is the reverse of the pattern that would be expected to be seen under periods of increased evaporation.

If it is assumed that the Interstadial carbonate marl at Tirinie contains minimal detrital contamination, consists of a single carbonate type (endogenic carbonate) and is precipitating in a basin that is not strongly effected by intra-basinal isotopic modification of the lake waters then the derived $\delta^{18}O$ and

$\delta^{13}C$ signal can be interpreted in the context of the following processes (Stuiver, 1970; Talbot, 1990; Leng and Marshall, 2004 and see Turney et al. 1997b for a discussion of these concepts in the context of the British Lateglacial). In lake systems the $\delta^{13}C$ of an authigenic carbonate is a reflection of the $\delta^{13}C$ of the dissolved inorganic carbon (DIC), which is in turn a function of the DIC of the ground- and surface-waters which recharge the lake basin, biological activity within the lake-basin and equilibration of CO_2 between the atmosphere and the lake water. (Talbot, 1990; Leng and Marshall, 2004). The $\delta^{13}C$ values of lacustrine carbonates precipitated in most open system lakes are typically between -3 and +3‰ (Talbot, 1990). The $\delta^{13}C$ values (+2.39 to -3.70‰) at Tirinie are, therefore, consistent with these values. The gradual decrease in $\delta^{13}C$ values that occurred during the accumulation of the Tirinie sequence could result from a number of factors, however, it is proposed that the most likely explanation is the impact of the progressive recolonization of the landscape after deglaciation on the $\delta^{13}C$ value of groundwater DIC. A gradual increase in vegetation cover and density would result in a higher contribution of plant respired CO_2 to the soil zone and, consequently, to vadose waters during aquifer recharge (Talbot, 1990; Andrews, 2006; Candy et al., 2012). As plant respired CO_2 is relatively depleted with respect to ^{13}C (Cerling and Quade, 1993) this would produce a progressive decline in the $\delta^{13}C$ value of groundwater DIC across the interval of the Interstadial.

The $\delta^{18}O$ of lacustrine carbonate is a function of the $\delta^{18}O$ of the lake water and the temperature at which carbonate precipitation occurs (Talbot, 1990; Leng and Marshall, 2004), the latter controlling the fractionation of oxygen isotopes during carbonate mineralisation, at ca -0.24‰/+1°C (Andersen and Arthur, 1983; Hays and Grossman, 1991; Kim and O'Neil, 1997). The $\delta^{18}O$ of lake water is controlled by a number of factors (see Talbot, 1990; Darling et al., 2003; Darling, 2004), primarily; 1) the $\delta^{18}O$ of precipitation, as this is the source of the moisture from which all surface water bodies recharge, and 2) any intra-basin fractionation processes, such as evaporation, that may cause the modification of the $\delta^{18}O$ value of the lake water. The latter processes are discounted based on the discussion above. The $\delta^{18}O$ of rainfall can be controlled by a number of factors (see Rozanski et al., 1992; 1993), including: 1) prevailing air temperature (which controls the fractionation of oxygen isotopes during condensation), 2) the amount effect, 3) the seasonality of rainfall and 4) the distance of movement/trajectory of the air masses that are the source of the rainfall. The accordance of $\delta^{18}O$ variations with other temperature proxy data derived from invertebrates demonstrate that, within isotopic records of the LGIT in Britain, air temperature plays a key role in controlling the $\delta^{18}O$ of carbonates ((Marshall et al., 2002; Lang et al., 2010; Van Asch et al., 2012). Air temperature exerts a major control on the $\delta^{18}O$ of rainfall, +0.58‰/+1°C for the modern mid-latitudes (Rozanski et al., 1992; 1993), which results in a direct relationship between air temperature and the $\delta^{18}O$ of surface waters. This suggestion is supported by the fact that, at key LGIT sites such as Fiddaun and Hawes Water declining $\delta^{18}O$ values are associated with decreasing temperatures whilst increasing $\delta^{18}O$ values are associated with increasing temperatures. That is not to say that temperature is the only environmental factor that affects this proxy but that it is a factor that exerts a particularly strong control.

If it is accepted that temperature, through its influence on the $\delta^{18}O$ of rainfall, plays a major role in controlling the $\delta^{18}O$ of lake water and, consequently, the $\delta^{18}O$ of lake carbonates then it can be proposed that the isotopic signal within the Tirinie Interstadial sequence comprises three climatic ameliorations (TIR-Oa, c and e) separated by 2 climatic deteriorations (TIR-Od and b). As has been noted by a number of authors the impact of temperature on the $\delta^{18}O$ of freshwater carbonates is always muted because the temperature control on the $\delta^{18}O$ of rainfall (Rozanski et al., 1992; 1993) is offset by the temperature control on isotopic fractionation (Hays and Grossman, 1991; Kim and O'Neil, 1997). This means a temperature increase of +1°C may generate an increase in carbonate $\delta^{18}O$ value of only ca 0.3‰ even though the impact of rainfall on such a temperature rise is closer to 0.6‰ (Andrews, 2006; Candy et al., 2011).

5.3 Inter-proxy comparison and sensitivity of Tirinie to climate change

There is a consistent pattern of response between the $\delta^{18}O$ signal, the pollen succession and the lithostratigraphy recorded in the Interstadial sequence at Tirinie (Figure 6). That is to say that each $\delta^{18}O$ peak occurs in association with a peak in % tree/shrub pollen and an increase in % carbonate within the sediment sequence. Low $\delta^{18}O$ values are associated with an increase in % herb pollen, at the expense of tree/shrub pollen, and an absence of carbonate, or major reduction, from the sediments in question. It is, however, important to note that shifts and peaks in $\delta^{18}O$ values slightly precede the changes in the pollen assemblage, suggesting that vegetation and landscape response lag behind the driver of the $\delta^{18}O$ signal. If it is assumed that climate, primarily temperature, is driving the $\delta^{18}O$ signal then two observations can be made: 1) that vegetation response lags behind climate forcing, and 2) the ecosystem and landscape in the Tirinie catchment is highly sensitive to climate forcing as there is a shift in the vegetation assemblage associated with every major oscillation in the $\delta^{18}O$ signal.

In such a model a climatic amelioration, as identified by an increase in $\delta^{18}O$ values, would produce an expansion of trees and shrubs within the landscape, a stabilisation of the catchment and, consequently, a reduction in the supply of allogenic material to the basin. The increase in temperature and its influence on productivity within the lake basin, coupled with a reduction in allogenic inputs, would result in conditions favourable for marl accumulation (Candy et al., 2015; Palmer et al., 2015).

Conversely a climatic deterioration, as identified by a decrease in $\delta^{18}O$ values, would produce a contraction of trees and shrubs within the landscape, a reduction in the stability of the catchment and, consequently, an increase in the supply of allogenic material to the basin (Candy et al., 2015; Palmer et al., 2015). Accelerated erosion in the landscape, combined with a reduction in productivity within the lake basin as a result of reduced temperatures, would result in a cessation of marl accumulation, producing a sediment sequence dominated by minerogenic sedimentation (Palmer et al., 2015).

The close coupling of climate forcing, as recorded in the $\delta^{18}O$ signal, and landscape response, as recorded by the pollen and sediment record, suggests that the Tirinie catchment is highly sensitive to abrupt and short-lived climatic perturbations. It is likely that this sensitivity is a result of the high altitude of the site which would potentially result in the site lying close to ecological boundaries. In such a setting even minor shifts in temperature could result in the upward or downward migration, within the landscape, of the boundaries between different ecotones. This may result in the vegetation response to climatic events within the Lateglacial Interstadial being more clearly expressed at Tirinie than at other sites that occur at less sensitive positions within the landscapes, i.e lowland locations. It is also interesting to note that, despite the long-term trend to decreasing $\delta^{13}C$ values that occurs across the Interstadial, peaks and troughs in $\delta^{13}C$ values that occur on top of this trend are broadly coincident with shifts in pollen assemblages. The peak in $\delta^{13}C$ values that occurs at 345cm depth coincides with a grass dominated landscape whilst the low in $\delta^{13}C$ values that occurs at 339-337cm depth coincides with a peak in *Betula* and *Juniperus*. Whilst it is not possible to prove a direct link between $\delta^{13}C$ values and vegetation assemblage it is possible that during cold intervals, when grassland is dominant, vegetation density declines allowing more atmospheric CO_2, which has relatively high $\delta^{13}C$ values, into the DIC pool (Carling and Quade, 1993). Conversely, during warm intervals, when trees/shrubs become more significant, vegetation density increases resulting in more plant respired CO_2, which has relatively low $\delta^{13}C$ values, being supplied to the DIC pool (Carling and Quade, 1993). This pattern is not observed in other LGIT lacustrine carbonate $\delta^{13}C$ records (Marshall et al., 2002; Van Asch et al., 2013), however, this may reflect the fact that Tirinie is unique among such sites in occurring at a high altitude. In such a setting vegetation cover, particularly during cold events, may be low resulting in subtle changes in temperature being sufficient to alter the biomass density enough to effect carbon sources in the lake basin.

6 $\delta^{18}O$ signal of the Tirinie sequence and its wider palaeoclimatic significance

6.1 The oxygen isotopic stratigraphy of the Lateglacial Interstadial at Tirinie and other sequences in the North Atlantic region

The $\delta^{18}O$ stratigraphy of the Lateglacial Interstadial sequence at Tirinie is consistent with that observable in all currently existing $\delta^{18}O$ lake carbonate records of this time interval in the British Isles (Marshall et al., 2002; Diefendorf et al., 2006; Van Asch et al., 2012). A comparison between the $\delta^{18}O$ record of the Lateglacial Interstadial at Fiddaun (western Ireland; Van Asch et al., 2012), Lough Inchiquin (western Ireland; Diefendorf et al., 2006), Hawes Water (northwest England; Marshall et al., 2002) and Tirinie (Scottish Highlands) show that they all have the following characteristics in common. Firstly, that all four sites record three discrete $\delta^{18}O$ peaks (labelled e, c and a on Figure 7). If, following the argument outlined above, these are inferred to reflect climatic ameliorations then, in the British Isles, the Lateglacial Interstadial is characterised by three separate warm phases. Secondly, that these $\delta^{18}O$ peaks are separated by two significant intervals characterised by low $\delta^{18}O$ values (labelled d and b on Figure 7), which are interpreted as representing climatic deteriorations or cold episodes. Finally, at all four sites there is evidence for a minor decline in $\delta^{18}O$ values during the middle $\delta^{18}O$ peak, suggesting that this warm episode was punctuated by a further short-lived cold event (labelled c1 on Figure 5). This stratigraphic pattern is also consistent with the $\delta^{18}O$ stratigraphy of GI-1 in the NGRIP record (Figure 8), i.e. GI-1e, 1c and 1a represent warm intervals, GI-1d and b represent cold intervals and GI-1c contains a short-lived cooling episode within it (Rasmuseen et al., 2006; 2008). Although there is insignificant chronological information to discuss the synchroneity of these warm and cold events across Britain or the North Atlantic the isotopic structure of the Interstadial across this region is consistent. Furthermore, the position of the Penifiler tephra in the Tirinie sequence, during the rise in $\delta^{18}O$ values from TIR-Od to TIR-Oc, is consistent with the increase in $\delta^{18}O$ values in NGRIP from GI-1d to GI-1c, the time interval with which this tephra is correlated (Matthews et al., 2011).

The comparison shown in Figures 7 and 8, also shows some key differences in the $\delta^{18}O$ structure of Interstadial records between Greenland and Britain and also between the British sites. Firstly, if it is assumed that $\delta^{18}O$ is an indicator of temperature then the highest $\delta^{18}O$ values within the Interstadial records can be used to identify the interval of maximum warmth. At Tirinie this is clearly the middle of the Interstadial (TIR-Oc). At Hawes Water, Lough Inchiquin and Fiddaun the isotopic data are less equivocal (Marshall et al., 2002; Diefendorf et al., 2006; Van Asch et al., 2012). Although at all three sites the highest individual $\delta^{18}O$ measurements are found within the middle of the Interstadial, at Fiddaun the difference between the maximum value in peaks c and a are minimal (0.2‰), whilst at Lough Inchiquin and Hawes Water the high value in peak c is an outlier with the majority of $\delta^{18}O$ measurements from this section of the Interstadial sequence being consistent with those from peaks a and e. Consequently, although the Tirinie sequence shows a clear $\delta^{18}O$ maximum in the middle of the

Interstadial, at Fiddaun, Hawes Water and Lough Inchiquin it is probably more robust to state that there is a general consistency in the $\delta^{18}O$ values that occur across all three of the Interstadial isotope peaks. The implication of this is that, for the Irish and English sites at least, there is no evidence for a single pronounced, or uniquely warm, peak within the Interstadial with all three ameliorations being of a broadly similar magnitude. Such an observation is consistent with the chironomid-based temperature reconstructions from both Hawes Water (Lang et al., 2010) and Fiddaun (Van Asch et al., 2012).

What is clear from this comparison is that none of the isotopic records from the British Isles shows evidence for the early Interstadial peak in $\delta^{18}O$ values (GI-1e at ca 14.6 kyr ago) that is clearly seen in NGRIP, and other Greenland records (Rasmussen et al., 2006; 2008). This early isotopic peak has been used to suggest that the thermal maximum of the Interstadial occurred immediately after its onset and was characterised by very high levels of warmth, a climatic pattern also seen in some palaeoecological-based temperature reconstructions from Britain (Atkinson et al., 1987; Walker et al., 1993). The absence of the very lowest sediments in the Interstadial sequence at Tirinie means that this site cannot provide conclusive evidence for an absence of an early Interstadial peak, however, the highest $\delta^{18}O$ values (TIR-Oc) occur after the Penifiler tephra, 13,939 ± 66 $\mu \pm \sigma$ cal BP (Bronk Ramsey et al., 2015) and, therefore, cannot be correlated with the early Interstadial peak seen in the Greenland records, ca 14.6 kyr ago.

6.2 The magnitude of $\delta^{18}O$ oscillations within the Lateglacial Interstadial at Tirinie and other British lacustrine carbonate records

The clearest difference between the Interstadial $\delta^{18}O$ record from Tirinie and those from Fiddaun, Hawes Water and Lough Inchiquin is the magnitude of the isotopic shifts that are found at this site (Table 2). The difference between the maximum and minimum values associated with the various isotopic shifts within the Tirinie record are, in all cases, >2.30‰ and, at their greatest, up to 3.30‰. It is important to highlight that as the minimum of the cold event TIR-Od contained no carbonate accumulation the isotopic shifts associated with it, the decline of -2.32‰ from TIR-Oe to d and the increase of +2.98‰ from TIR-Od to c, represent the minimum values of these isotopic variations. To place the magnitude of these isotopic shifts into the context of the other sites, at Fiddaun the isotopic decrease that occurred from the last peak of the Interstadial to the climatic minima of the Younger Dryas is ca 4.00‰, whilst at Hawes Water it is ca 2.00‰. The isotopic variations seen *within* the Interstadial at Tirinie are, therefore, as great, or greater, than the full isotopic shift associated with the Younger Dryas. On average the isotopic shifts that occur during the Interstadial sequences at Fiddaun, Hawes Water and Lough Inchiquin are 1.56‰, 1.40‰ and 1.25‰ in magnitude respectively compared to an average of 2.78‰ at Tirinie. As shown in Table 2 this difference cannot be a result of sampling resolution because there is no real difference in the sampling resolution, across the Interstadial interval, between any of these sites.

It is important to note that the magnitude of some of the Interstadial isotopic shifts seen at Tirinie is comparable to those seen in NGRIP (Table 2). For example the increase in $\delta^{18}O$ values from TIR-Ob to TIR-Oa is 2.53‰, whereas in NGRIP the increase in $\delta^{18}O$ values between GI-1b and GI-1a, the shift that occurs at a comparable point in the isotope stratigraphy of the Interstadial but is not necessarily synchronous, is 2.86‰. This similarity is surprising because isotopic shifts driven by climatic changes of a comparable magnitude will always be more muted in a carbonate record than in an ice-core record. This is because ice cores directly record the temperature-controlled change in the $\delta^{18}O$ of precipitation (+0.58‰/+1^{o}C for the modern), whereas a lake sequence will record the temperature controlled change in the $\delta^{18}O$ of precipitation, transferred to lake water through a number of processes, *plus* the temperature controlled fractionation of oxygen isotopes that occurs during carbonate mineralisation (Andrews, 2006; Candy et al., 2011). Hypothetically, a 1^{o}C temperature change could, therefore, be represented by a ca 0.6‰ shift in an ice-core sequence but a ca 0.3‰ shift in a lacustrine carbonate. When this is taken into account it is possible that the 2.53‰ shifts in the Tirinie sequence, quoted above, could actually reflect a bigger shift in the $\delta^{18}O$ of the source water than that represented by the 2.86‰ seen in NGRIP. Clearly directly comparing the isotopic shifts observable in a lacustrine carbonate sequence and an ice-core sequence is fraught with issues and assumptions, however, this comparison is made simply to highlight the large magnitude of the isotopic shifts that are seen in the Tirinie sequence. Not only are these isotopic shifts significantly greater than those observable in any other British Interstadial $\delta^{18}O$ record but, in some cases, these are the same as, or theoretically greater than, those seen in NGRIP.

Two possible explanations are put forward to account for high magnitude $\delta^{18}O$ shifts seen in the Tirinie record. First, the small size of the lake basin in question. Leng and Marshall (2004) suggest that the $\delta^{18}O$ of lacustrine waters in smaller lake basins have the potential to respond more rapidly to short-lived environmental change than larger bodies of water, and, consequently, the resulting sediment archive will record such changes more clearly. This is certainly a possible explanation as the basin at Tirinie is significantly smaller than Hawes Water, Fiddaun or Lough Inchiquin. Second, that the magnitude of the climatic shifts that occurred during the Interstadial were significantly greater at Tirinie than those that occurred at the other three sites and are, consequently, propagated into larger isotopic shifts. There may, therefore, be a regional difference in the magnitude of climatic change, during the Interstadial, between more northerly and more southerly sites.

Currently, it is not possible to demonstrate which of these explanations is more probable. The idea, however, that a more northerly site, such as Tirinie, experienced a greater degree of climatic variability during the Lateglacial Interstadial than more southerly sites, such as Fiddaun and Hawes Water, is supported by chironomid-based temperature reconstructions. Such reconstructions exist for both Fiddaun and Hawes Water and show that, during the Interstadial, temperature fluctuations are <2^{o}C. Although Tirinie itself has no associated quantified temperature records a number of chironomid-based Interstadial temperature records do exist from nearby sites in northern Britain, these include; Whitrig Bog (Brooks and Birks, 2000a), Loch Ashik and Abernethy Forest (Brooks et al., 2012). The magnitude of the temperature shifts in all of these records show much greater temperature variability during the Interstadial: Whitrig Bog = 5^{o}C (Brooks and Birks, 2000a), Loch Ashik = 4-5^{o}C (Brooks et al., 2012) and Abernethy Forest = 6-7^{o}C (Brooks et al., 2012). These shifts also indicate that cold oscillations observed in the Interstadial record the same magnitude of cooling as seen in the Loch Lomond Stadial (Younger Dryas) event in the same records. The suggestion that the magnitude of the isotopic shifts seen at Tirinie represent the occurrence of temperature changes during the Lateglacial Interstadial that were significantly greater than those experienced at more southerly sites is not, therefore, without supporting evidence.

If the above argument is accepted then it raises the question: why, in the British Isles, are $\delta^{18}O$ oscillations and chironmid inferred temperature variations during the Lateglacial Interstadial greater at more northerly sites than those observed in more southerly sites? In answer to this we propose that it is the geographical location of sites, such as Tirinie, that is crucial, in particular its relatively high-latitudinal position. With respect to latitude, it is frequently argued that climatic instability during the LGIT in the northeastern Atlantic is driven by changes in ocean forcing, particularly the release of latent heat into the atmosphere, elevating air temperatures, during Atlantic Meridional Overturning Circulation (AMOC)(Broecker et al., 1989; Lehman and Keigwin, 1992; Karpuz and Jansen, 1992; Koc et al., 1993; Clark et al., 2001; Nesje et al., 2004). It can be argued that within the region where AMOC is most strongly felt, ca 50 – 60^{o}N, the more northerly the location of a site the more significant even short-lived fluctuations in circulation will be (Broecker et al., 1989; Brooks and Birks, 2000b; Bakke et al., 2009). This is because weakening of AMOC may retard the transfer of heat into more northerly locations but maintain a supply of heat into more southerly locations (Bakke et al., 2009). It could, therefore, be argued that the location of Tirinie at 56.5^{o}N i.e. one of the most northerly situated Interstadial lacustrine $\delta^{18}O$ record in Europe, would make it more sensitive to short-lived climatic fluctuations than more southerly sites and, therefore, the magnitude of observed $\delta^{18}O$ fluctuations should be greater. It could also be suggested that significantly further north of this latitude the climatic influence of AMOC is less strongly felt reducing the potential for the expression of its variability to be preserved in palaeoenvironmental records of this region. This is certainly true in temperature reconstructions from northern Norway where climatic oscillations within the Interstadial are in the order of 1-2°C, i.e. within the uncertainty of the technique (Birks et al., 2012). The role of latitude may also be amplified by the impact of altitude as, in the upland regions of Scotland at the present day, the complex interplay of topography and local conditions can produce substantial temperature gradients across altitudinal ranges (McClatchey, 1996). Such altitude-based controls would have a much greater impact on a site such as Tirinie (>300 metres a.s.l.) than sites such as Hawes Water, Fiddaun and Lough Inchiquin (<20 metres a.s.l.).

6.3 Summary

The study of the Interstadial sediments from Tirinie highlights two key factors. Firstly, that the integration of the $\delta^{18}O$ record from Tirinie with other existing isotopic records of the Lateglacial Interstadial from the British Isles suggest a coherent oxygen isotopic stratigraphy for this time interval. Despite the lack of strong chronological controls on many of these sequences the number of isotopic events and the isotopic structure of the Interstadial appears to be regionally consistent. Secondly, despite this regional stratigraphic consistency, the magnitude of the $\delta^{18}O$ variations that occurred during the Interstadial varies across Britian. Although this could partly be a function of lake-basin size there is good reason to suggest that this could also be a function of regional differences in the magnitude of the temperature shifts that occurred within the Interstadial. Understanding the spatial variability of abrupt climate change has been identified as an important research question (Bakke et al., 2009; Lane et al., 2013) and the sensitivity of $\delta^{18}O$ as a proxy makes it an ideal medium through which to investigate this. The construction of a network of sites, with a resolution and precision comparable to that of Tirinie, along latitudinal and longitudinal climatic gradients is a key basis for addressing such questions and will allow a greater understanding of spatial variability in the expression of abrupt climatic events.

7 Conclusions

A detailed litho-, bio, tephro- and isotope-stratigraphy is constructed for carbonate rich lacustrine sediments from the site of Tirinie, Scotland, which span the Lateglacial Interstadial. The data generated from these sediments has resulted in the following conclusions being drawn:

- The $\delta^{18}O$ record of the Tirinie Interstadial sequence suggests that this interval was characterised by complex climate change consisting of three distinct climatic ameliorations separated by two pronounced cold episodes.
- The pollen and sediment record of this side indicate that the Tirinie catchment was highly sensitive to the climate changes that occurred during the Interstadial with each $\delta^{18}O$ shift correlating with a shift in pollen assemblage and sediment type.
- Although the $\delta^{18}O$ stratigraphy of the Lateglacial Interstadial at Tirinie is consistent with the $\delta^{18}O$ record of other Interstadial lacustrine carbonate records from the British Isles suggesting a regionally consistent pattern of climate change across this region.
- Despite the regional consistency of the isotope stratigraphy of the Interstadial, the magnitude of $\delta^{18}O$ variations found at Tirinie during this interval are significantly greater than those seen in any other British Lateglacial Interstadial $\delta^{18}O$ record.
- It is proposed that the reason for the high magnitude $\delta^{18}O$ shifts that occurred during the Interstadial at Tirinie is the regional variability in the magnitude of abrupt climatic events during the LGIT. In this model more northerly sites, such as Tirinie, experienced much greater temperature shifts, and, therefore, much greater $\delta^{18}O$ shifts, during the Interstadial than more southerly sites such as Fiddaun, Hawes Water and Lough Inchiquin
- $\delta^{18}O$ analysis of LGIT lacustrine carbonates has great potential for investigating climatic gradients across this time interval and future work needs to focus on the construction of regional networks of such sites along latitudinal and longitudinal gradients.

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Figure 1 – Location of the site of Tirinie in the context of other key sites in the North Atlantic and Nordic Seas region (a). Location of Tirinie in the context of other sites within the British Isles that have used the $\delta^{18}O$ analysis of lacustrine carbonate sequences and/or chironomid studies to reconstruct climatic variations during the Last Glacial to Interglacial Transition (LGIT) (b). (c) shows the topographic details of the study area (contours in metres), note that most of the catchment that drains into the modern mire is situated in the higher ground to the north.

Figure 2 – Stratigraphy of the Tirinie sequence; a) basic lithostratigraphy of the whole LGIT sequence showing generalised stratigraphic units, % carbonate content and tephra concentrations, b) Detailed stratigraphy of lowermost marl unit (355-327cm below surface) showing detailed % carbonate and tephra concentrations. Not that in b) the tephra peak occurs during a period of increasing carbonate content.

Figure 3 – Pollen percentage diagram from Tirinie. All land pollen percentages were calculated with total land pollen, whilst aquatics were calculated using TLP+Aquatics with TLP+Spores for spore data. A 10X exaggeration line has been shown indicating rare taxa. What is striking is the close resemblance of the biostratigraphy and the lithostratigraphic record.

Figure 4 - $\delta^{18}O$ and $\delta^{13}C$ variations within the lowermost marl unit at Tirinie (shown as variations in ‰ in relation to VPDB). TIR-Oe, c and a represent peaks in $\delta^{18}O$ values, whilst TIR-Od and b represent areas of low $\delta^{18}O$ values. Note the short-lived isotopic decline during the peak of TIR-Oc, labelled TIR-Oc1. The size of the error bars on each data point is smaller than the size of the point itself, for associated uncertainties please see supplementary information 2.

Figure 5 - Geochemical data for two tephra layers preserved in Tirinie. A = A Total Alkalis Vs Silica bi-plot using the divisions of Le Maitre and Streckeisen (2002). B = An FeO(t) Vs CaO bi-plot of the Silicic shards showing the close match of TIR-305 to the Vedde Ash and TIR-348 to the Penifiler tephra. The comparative data used for the Vedde Ash, Penifilier and Borrobol tephras was downloaded from the RESET database and presented as 95% kernel

density envelopes of the total distribution (Bronk Ramsey et al., 2015). The raw data in presented in supplementary data table 1.

Figure 6 – Comparison between the Tirinie $\delta^{18}O$ record and % variations in the key pollen taxa. Note that taxa indicative of landscape stability (*Betula*, *Juniperus* and *Empetrum*) occur in association with, but lag slightly behind, peaks in $\delta^{18}O$, whilst taxa that are indicative of disturbed ground and landscape instability (*Poaceae*, *Artemisia* and *Rumex*) occur in association with, but again lag behind, zones of low $\delta^{18}O$ values. This suggests that these taxa are responding to, and, therefore lag behind, climate forcing, as represented by the $\delta^{18}O$ signal.

Figure 7 – The $\delta^{18}O$ record of the Lateglacial Interstadial from the Tirinie sequence (a) compared to that of (b) Haweswater (Marshall et al., 2002), (c) Lough Inchquin (Diefendorf et al., 2006) and (d) Fiddaun (van Asch et al., 2012). In all cases the x-axis are scaled to the same amount so the observable magnitude of $\delta^{18}O$ variation is real. Note that the magnitude of $\delta^{18}O$ shifts in the Tirinie sequence are significantly greater than those in all of the other records. All sites show a similar $\delta^{18}O$ stratigraphy for the Lateglacial Interstadial comprising 3 distinct peaks (a, c and e) and 2 major lows (b and d). In all cases a decline in $\delta^{18}O$ values occurs in the middle of peak c. It is important to note that this correlation is not meant to suggest that these events are synchronous but simply that a similar $\delta^{18}O$ stratigraphy exists for the Interstadial at all four sites.

Figure 8 – The Lateglacial Interstadial $\delta^{18}O$ record of Tirinie compared with that of the GRIP Greenland ice core record (GRIP, 2003; Rasmussen 2006; 2008). The chronology of Tirinie is not secure enough to allow direct correlation, although the occurrence of the Penfiler tephra (14090-13650 cal yr BP; Matthews et al., 2011) on the warming limb between TIROx2 and TIROx3 allows this low and high in $\delta^{18}O$ values to be correlated, stratigraphically, with GI-1d and GI-1c respectively. This figure does not, therefore, suggest that these events are directly synchronous but simply that the $\delta^{18}O$ stratigraphy of the two records is comparable. The figure clearly shows that the magnitude of shifts in $\delta^{18}O$ values seen in the GRIP and Tirinie record are comparable even though the $\delta^{18}O$ of ice is being compared with that of lacustrine carbonates.

Table 1 – Summary of the main characteristics of the pollen zones recorded in this study (TIR-P*n*) and their correlatives from Lowe and Walker (1977) T*n*.

Table 2 – Comparison of the value of the quantified shifts in $\delta^{18}O$ values during the Lateglacial Interstadial between Tirinie, HawesWater (Marshall et al., 2002), Lough Inchiquin (Diefendorf et al., 2006) and Fiddaun (Van Asch et al., 2002). The letters of each climatic event correspond to those shown in Figure 6. The sampling resolution of all four lacustrine sequences are broadly comparable suggesting that the magnitude of the $\delta^{18}O$ variations seen at Tirinie cannot be explained by sampling strategy. This suggestion is also supported by the fact that the Tirinie contains the “shortest” Interstadial sequence (27 cm) of all of the sites (Hawes Water – ca 65cm, Lough Inchiquin - ca 70cm and Fiddaun – ca 90 cm). In a shorter sequence oscillations have a greater likelihood of being compressed and consequently

appearing more subdued in the derived $\delta^{18}O$ record. The value for shifts e/d and d/c in the Tirinie sequence is shown in italics because it represents the minimum value for this event as it is unlikely that the coring process recovered the earliest part of the Interstadial.

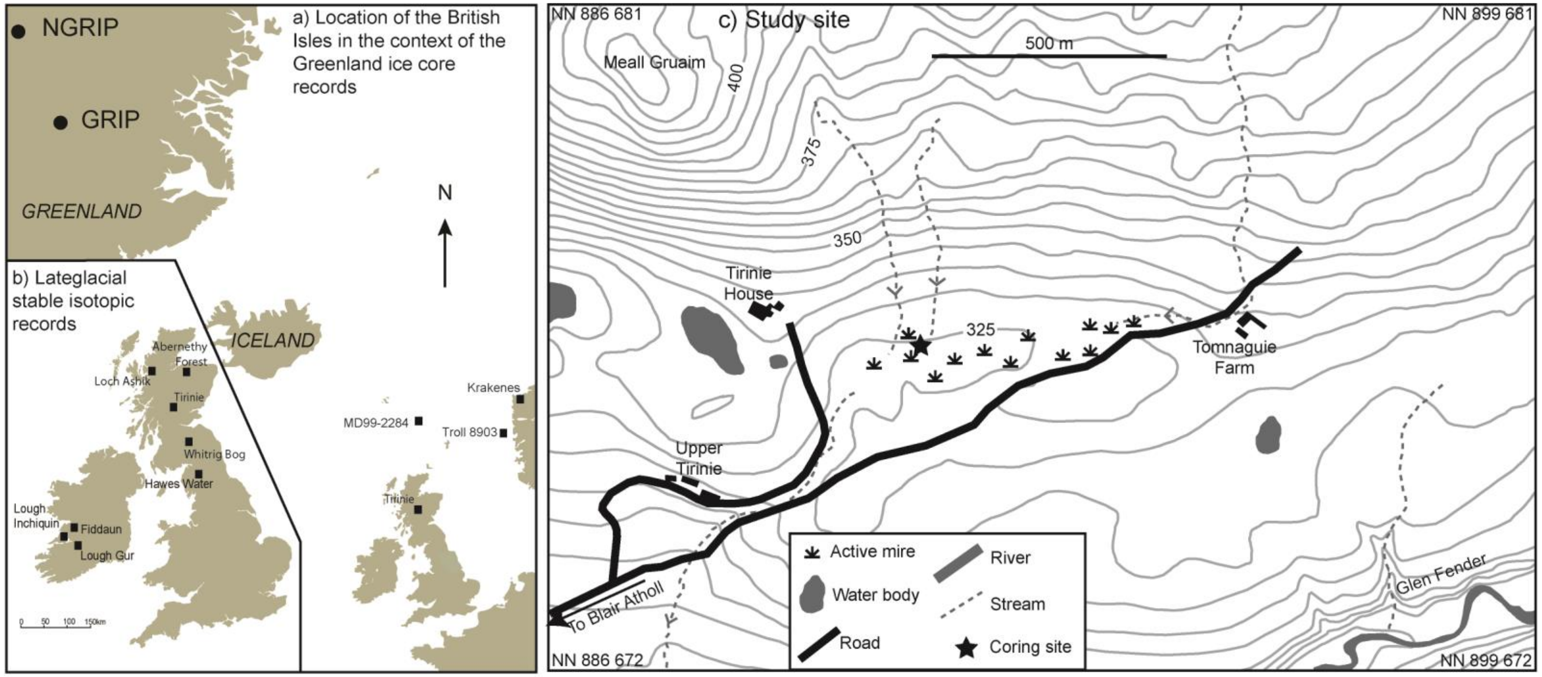
Table 1

Zone (from base up)	**Main taxa defining the zones**	**Equivalent zone in Lowe and Walker (1977)**
Zone TIR-P1a	Poaceae, Cyperaceae, *Empetrum*	Not present
Zone TIR-P1b	Poaceae, Cyperaceae, *Rumex, Artemisia*	T1
Zone TIR-P2	Poaceae, Cyperaceae, *Betula, Juniperus, Empetrum*	T2
Zone TIR-P3	*Betula*, Poaceae, *Juniperus*, Cyperaceae	T3
Zone TIR-P4	Poaceae, *Rumex*, Cyperaceae, *Pediastrum*	T4
Zone TIR-P5	*Juniperus*, *Betula*, Poaceae, Cyperaceae	
Zone TIR-P6	Poaceae, Cyperaceae, *Rumex, Pediastrum, Selaginella*	

Table 2

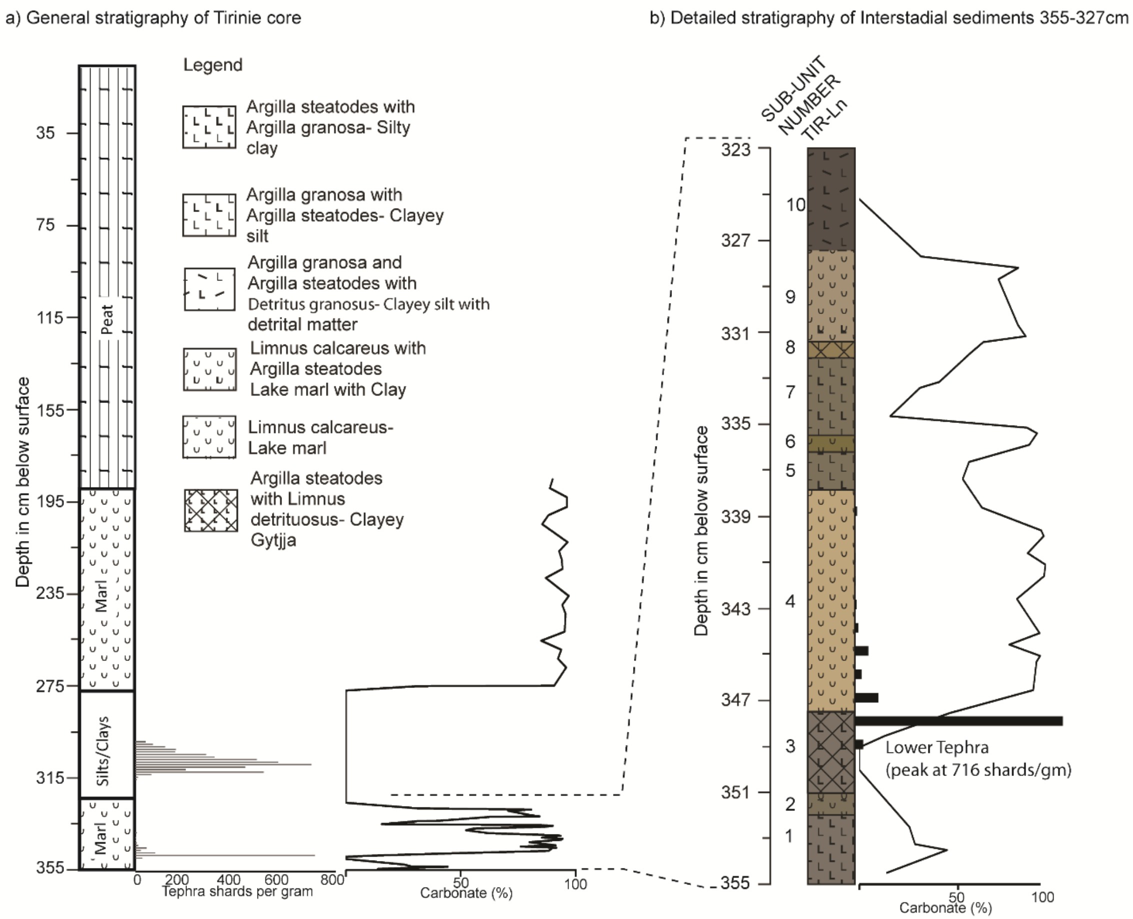
Event	Tirinie	Difference	Haweswater	Difference	Lough Inchquin	Difference	Fiddaun	Difference	GRIP	Difference
a	-6.82		-4.37		-5.04		-3.13		-37.68	
		-2.53		-0.91		-0.93		-1.36		-2.86
b	-9.35		-5.28		-5.97		-4.49		-40.54	
		-3.30		-1.32		-1.25		-1.40		-4.13
c	-6.05		-3.96		-4.72		-3.09		-36.41	
		-2.98		-1.85		-1.52		-2.06		-3.57
d	-9.03		-5.81		-6.24		-5.15		-39.98	
		-2.32		-1.52		-1.28		-1.42		-5.26
e	-6.71		-4.29		-4.96		-3.73		-34.72	
Resolution (samples per cm)	1.13		1.12		1.04		0.80			

Figure 1



NGRIP
GRIP
GREENLAND
a) Location of the British Isles in the context of the Greenland ice core records
N
ICELAND
b) Lateglacial stable isotopic records
Abernethy Forest
Loch Ashik
Tirinie
Whitrig Bog
Hawes Water
Lough Inchiquin
Fiddaun
Lough Gur
Krakenes
MD99-2284
Troll 8903
Tirinie
c) Study site
NN 886 681
NN 899 681
NN 886 672
NN 899 672
500 m
Meall Gruaim
400
375
350
325
Tirinie House
Upper Tirinie
Tomnaguie Farm
To Blair Atholl
Glen Fender
Active mire
Water body
Road
River
Stream
Coring site

Figure 2



a) General stratigraphy of Tirinie core
b) Detailed stratigraphy of Interstadial sediments 355-327cm
Legend
Argilla steatodes with Argilla granosa- Silty clay
Argilla granosa with Argilla steatodes- Clayey silt
Argilla granosa and Argilla steatodes with Detritus granosus- Clayey silt with detrital matter
Limnus calcareus with Argilla steatodes Lake marl with Clay
Limnus calcareus- Lake marl
Argilla steatodes with Limnus detrituosus- Clayey Gytjja
Peat
Marl
Silts/Clays
Depth in cm below surface
Tephra shards per gram
Carbonate (%)
SUB-UNIT NUMBER TIR-Ln
Lower Tephra (peak at 716 shards/gm)

Figure 3

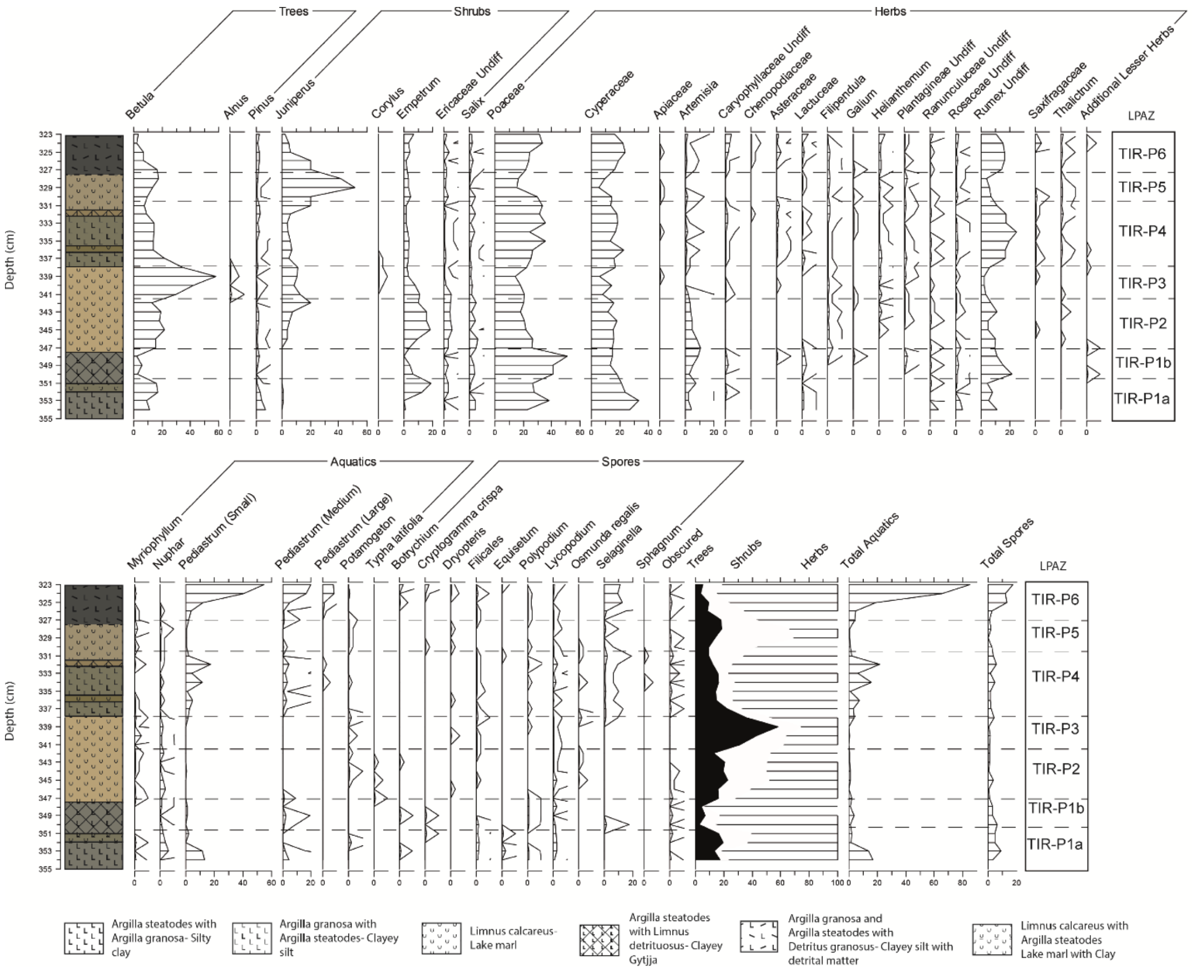
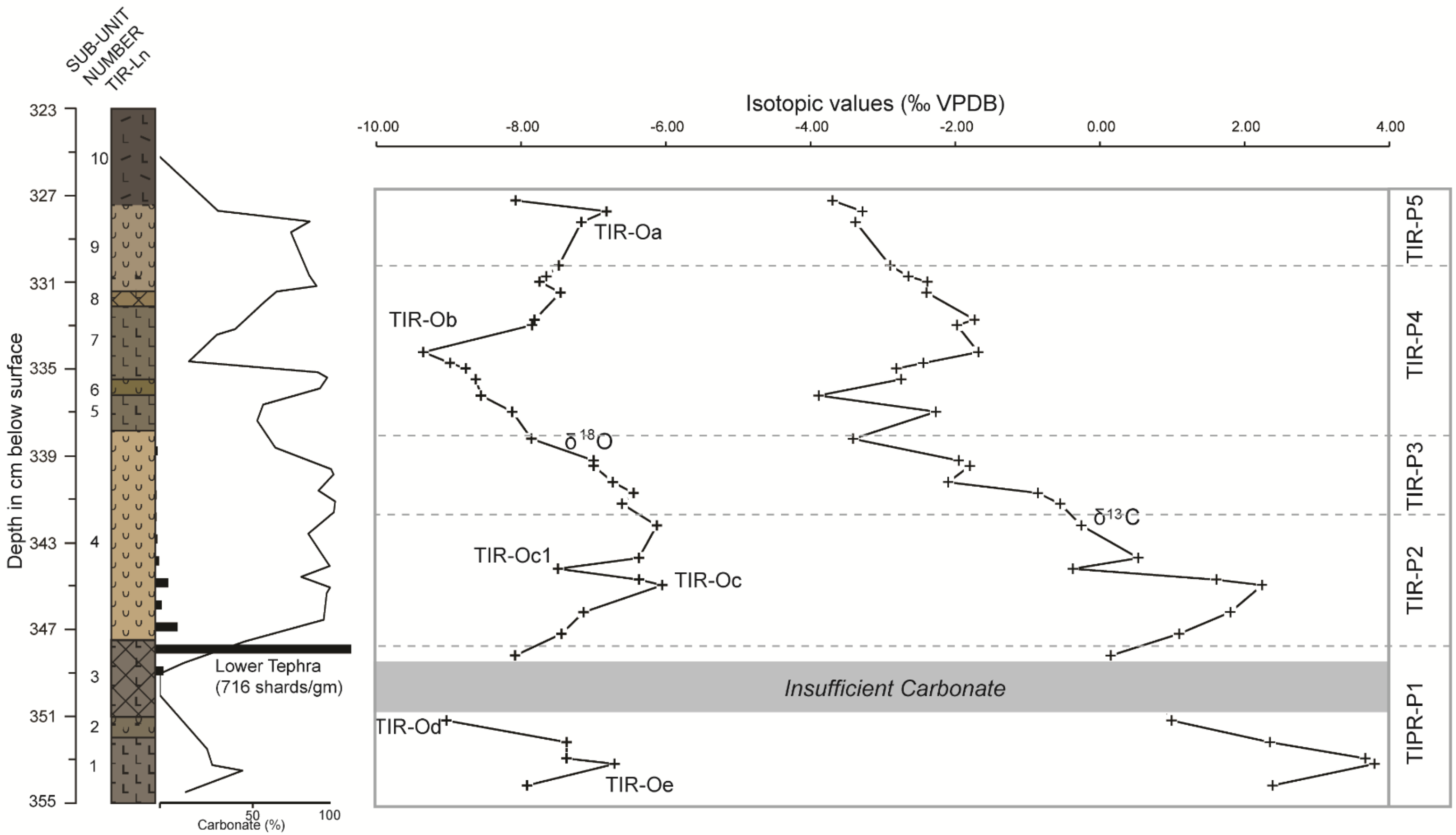
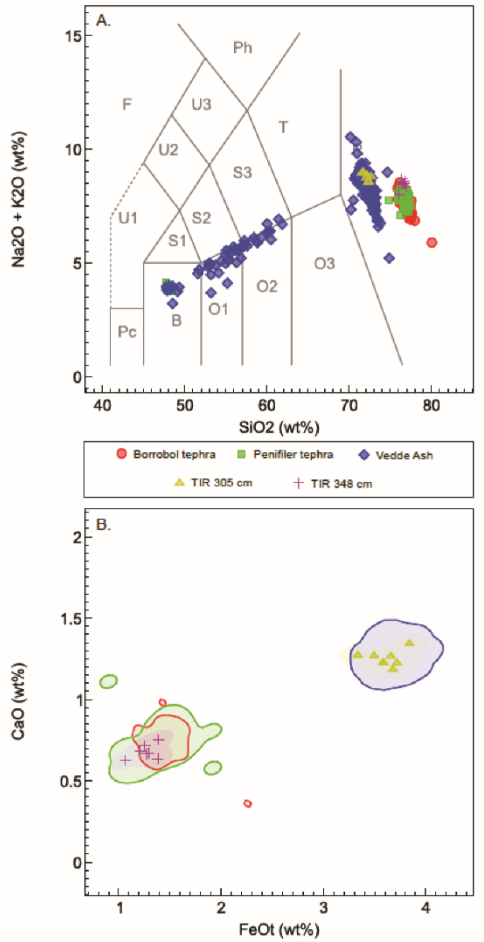


Figure 4



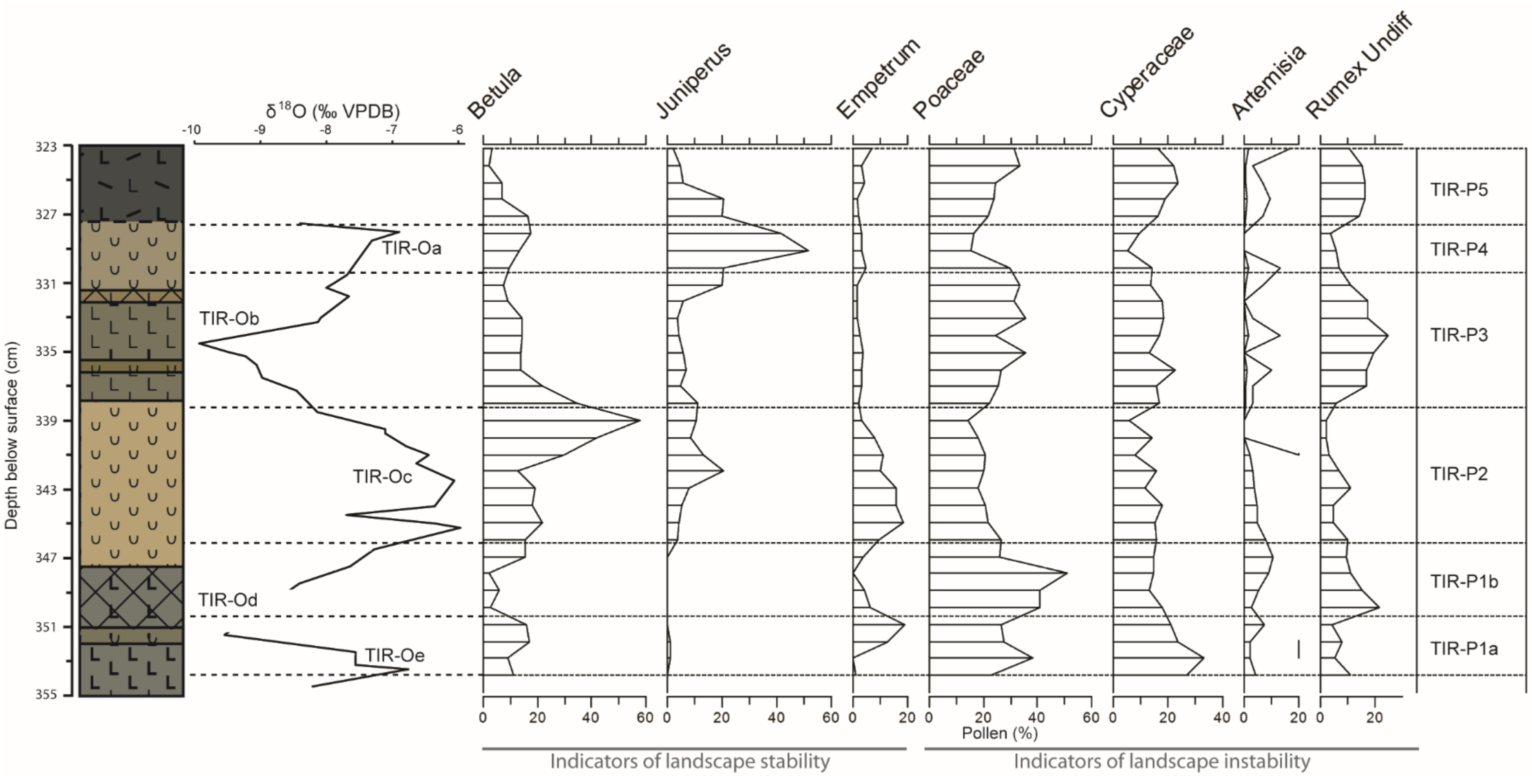
SUB-UNIT NUMBER TIR-Ln
Depth in cm below surface
323
327
331
335
339
343
347
351
355
10
9
8
7
6
5
4
3
2
1
Lower Tephra
(716 shards/gm)
50
100
Carbonate (%)
Isotopic values (‰ VPDB)
-10.00
-8.00
-6.00
-4.00
-2.00
0.00
2.00
4.00
TIR-Oa
TIR-Ob
δ18O
δ13C
TIR-Oc1
TIR-Oc
TIR-Od
TIR-Oe
Insufficient Carbonate
TIR-P5
TIR-P4
TIR-P3
TIR-P2
TIPR-P1

Figure 5



A.
Ph
F
U3
T
U2
S3
U1
S2
S1
O3
O2
O1
B
Pc
Na2O + K2O (wt%)
SiO2 (wt%)
Borrobol tephra
Penifiler tephra
Vedde Ash
TIR 305 cm
TIR 348 cm
B.
CaO (wt%)
FeOt (wt%)

Figure 6



δ18O (‰ VPDB)
-10
-9
-8
-7
-6
Betula
Juniperus
Empetrum
Poaceae
Cyperaceae
Artemisia
Rumex Undiff
Depth below surface (cm)
323
327
331
335
339
343
347
351
355
TIR-Oa
TIR-Ob
TIR-Oc
TIR-Od
TIR-Oe
TIR-P5
TIR-P4
TIR-P3
TIR-P2
TIR-P1b
TIR-P1a
0
20
40
60
Pollen (%)
Indicators of landscape stability
Indicators of landscape instability

Figure 7

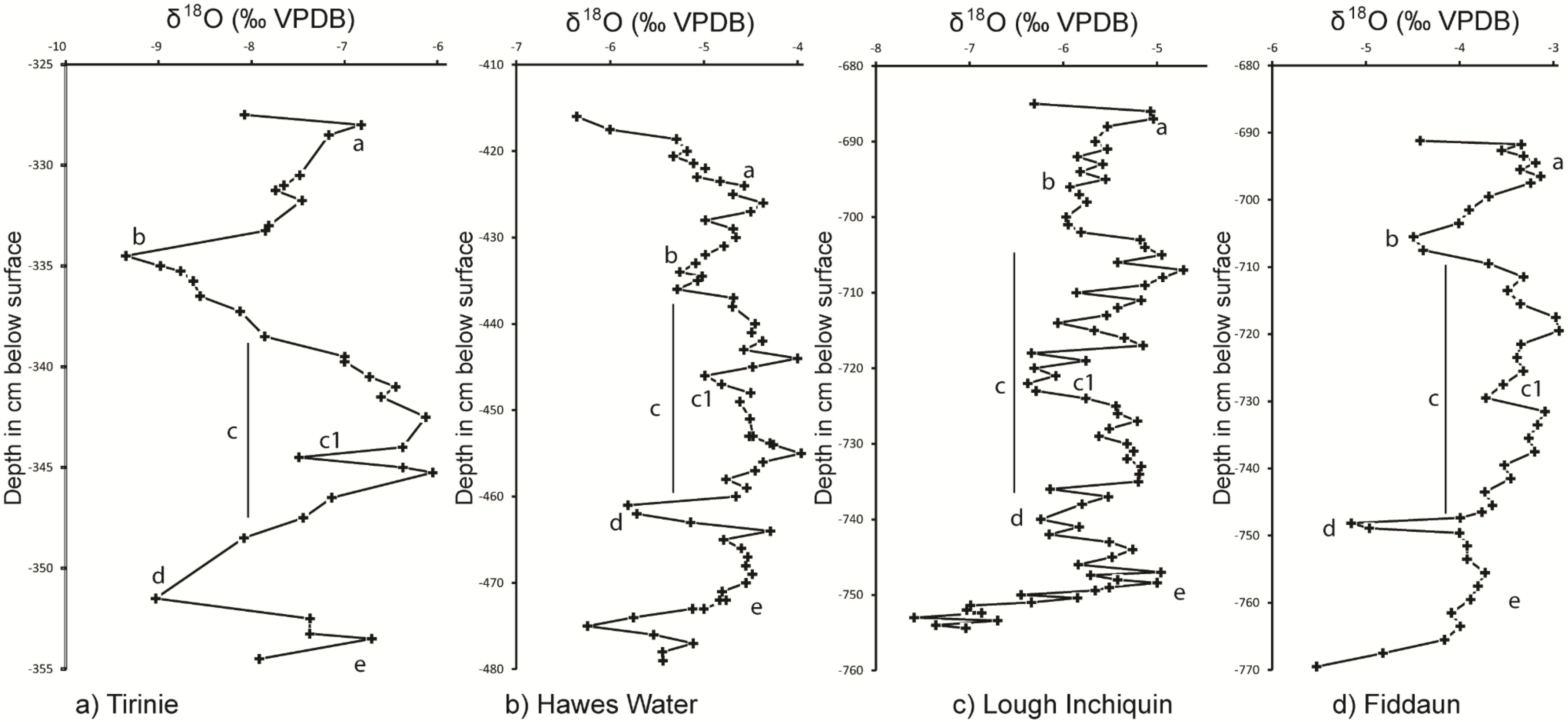
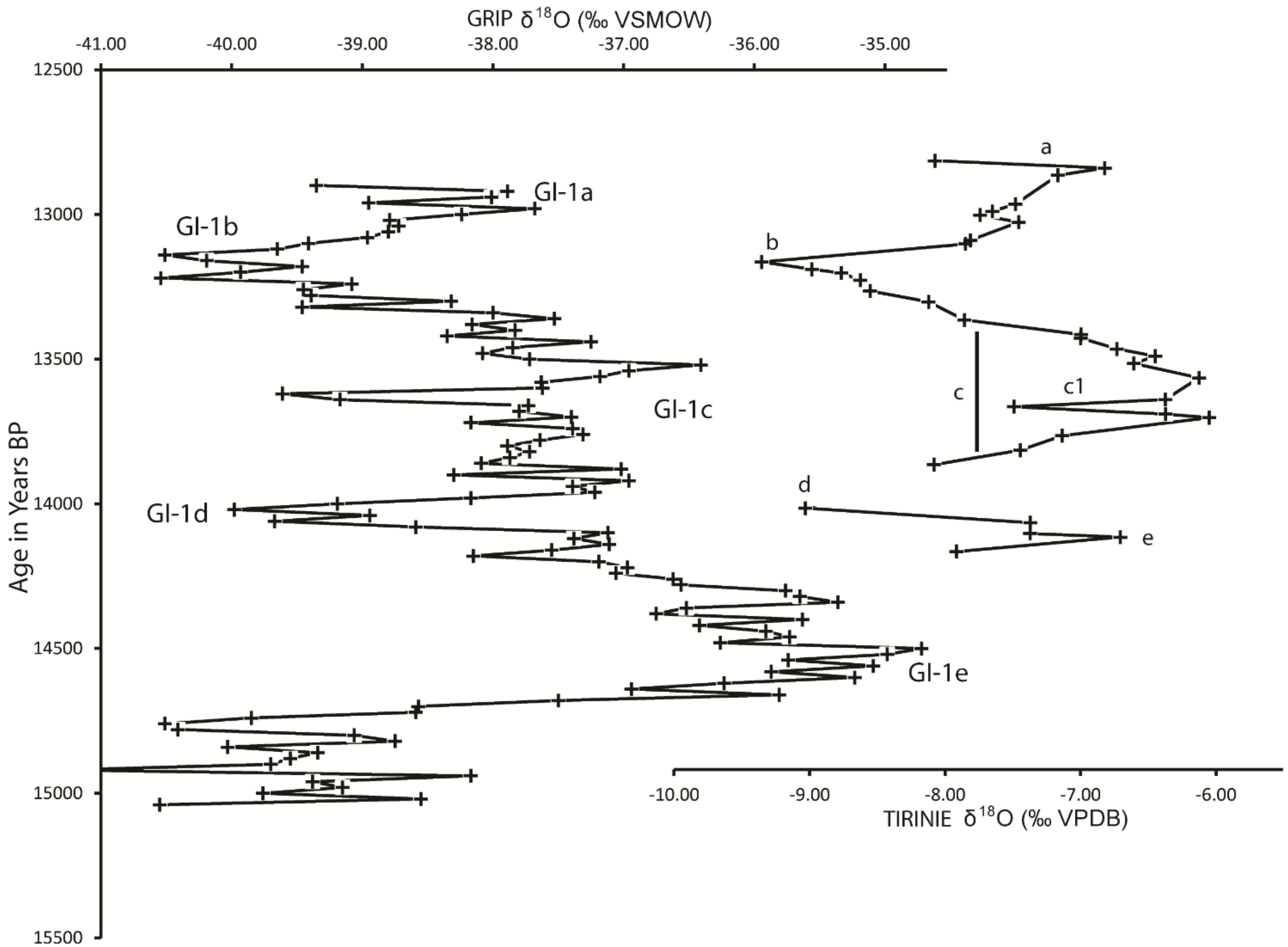

δ18O (‰ VPDB)
Depth in cm below surface
a
b
c
c1
d
e
a) Tirinie
b) Hawes Water
c) Lough Inchiquin
d) Fiddaun

Figure 8



GRIP δ18O (‰ VSMOW)
-41.00
-40.00
-39.00
-38.00
-37.00
-36.00
-35.00
12500
13000
13500
14000
14500
15000
15500
Age in Years BP
GI-1a
GI-1b
GI-1c
GI-1d
GI-1e
a
b
c
c1
d
e
-10.00
-9.00
-8.00
-7.00
-6.00
TIRINIE δ18O (‰ VPDB)