December 2022

Holocene Sea-level Change in the Dysynni Valley

A thesis submitted in partial fulfilment of the requirements for the degree of Doctor of Philosophy

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Abstract

Holocene sea-level and vegetation changes were reconstructed for the Dysynni Valley, north Cardigan Bay, Wales. Stratigraphic surveys were completed, and sample cores recovered from three locations in the valley: Penllyn, Perfeddnant and Gesail. The cores were subsampled for loss on ignition, particle size, diatom, foraminifera, pollen, and radiocarbon dating. Transfer functions were applied, and sea-level index points (SLIPs) developed.

At Penllyn a basal peat was overlain by blue-grey silt then peat. A back barrier saltmarsh with mixed oak woodland and alder gave way to intertidal mudflats followed by low marsh. Marine conditions declined as alder carr colonised the infilled marsh, replaced by birch, before the landscape opened up. Two SLIPs were produced, indicating a rise in sea level from -7.82 to -4.91m OD MHWST during 7787-7665 cal. BP to 6402-6297 cal. BP. At Perfeddnant a basal freshwater peat was overlain by a brackish blue-grey silt followed by peat, with two freshwater limiting points established at 9885-9547 cal. BP and 4782-4420 cal. BP (-0.44 and -2.60m OD). At Gesail a short, freshwater core was dominated by high energy, fluvial environments.

Marine influence extended to Perfeddnant, with the maximum horizontal inundation at 7.76km inland. The data from the Dysynni were compared to regional patterns of sea-level change for Wales, indicating rising sea levels between 9330 to 685 cal. BP (-27.07m to-0.08m OD MTL). The research provided definition for the rapid sea-level rise of the early Holocene transitioning to the gradual sea-level rise of the mid-Holocene.

The environmental reconstructions revealed that the Dysynni was a wide tidal estuary with marsh, reed swamp and tall herb fen from 7787-7665 cal. BP. Prehistoric settlement existed on the high ground as islands in the estuary. Evidence of widespread agriculture in the Dysynni was limited until the post-medieval land improvements, reflected in the Penllyn pollen record.

Contents Page

List of Figures

List of Tables

List of Abbreviations

AMS	Accelerated mass spectrometry
BGS	British Geological Survey
CCA	Canonical correspondence analysis
DCCA	Detrended canonical correspondence analysis
GIA	Glacio-isostatic adjustment
GIS	Geographic Information Systems
GNNS	Global Navigation Satellite System
HAT	Highest astronomical tide
LGM	Last Glacial Maximum
LiDAR	Light detection and ranging
MAT	Modern analogue technique
MHHW	Mean higher high water
MHWST	Mean high water spring tide
MLLW	Mean lower low water
MTL	Mean tide level
ML	Maximum likelihood
OS	Ordnance survey
OD	Ordnance datum
PMSE	Palaeomarsh surface elevation
R^2	Coefficient of determination
RMSE	Root mean squared error
RMSEP	Root mean squared error of prediction
RSL	Relative sea level
RWL	Reference water level
SLIP	Sea-level index point
SWLI	Standardised water level index
TLP	Total land pollen
WA	Weighted averaging
WA-PLS	Weighted averaging partial least squares
WA-Tol	Tolerance downweighted weighted averaging

Acknowledgements

I feel very privileged to have received a studentship from the NERC ACCE DTP, I would never have succeeded without their funding and support. I am grateful for the training opportunities, as well as the academic support. I am grateful for my ACCE DTP cohort for their encouragement and friendship.

To my supervisors, Bob Johnston and Katherine Selby, I can't thank you enough for your unending patience and support throughout my PhD journey. Thank you for going above and beyond to help me, and for never giving up hope I would succeed.

I would like to thank the Snowdonia national park authority for support during fieldwork, and use of equipment, particularly John Roberts and Tomos Jones. I am thankful to Sarah Davies and her colleagues from Aberystwyth university for advice and support during fieldwork, especially for use of the percussion corer. I am grateful to the landowners and tenants for access to their land for fieldwork. I am thankful for all the help I have received during fieldwork through my PhD, I would not have succeeded without their help, particularly the Sheffield MA landscape archaeology students. I would like to thank the staff and PhD students from the York Department of Environment and Geography for help during laboratory and field work. I would also like to thank the staff at Sheffield Archaeology Department for help during laboratory and fieldwork particularly, Yvette, Angelos, Angela, Guglielmo, Giana and Emily.

I am immensely grateful to Sheffield DDSS and my mentors Hayley and Adele, without your support I wouldn't be who I am today. You have given me the knowledge, skills and mental strength to persevere, and taught me how to pick myself up when I was ready to give up. My thesis would not have been finished without you. I would like to thank my friends and family for supporting me throughout my life, and for giving me the strength to keep trying. Thank you for always being there when I needed you. I would also like to thank Terry Pratchett and his wonderful audiobooks for getting me through my laboratory work with most of my sanity intact.

Section A: Context of Research

Chapter 1: Introduction

The primary aim of the research is to reconstruct sea-level change through the Holocene in the Dysynni Valley, northern Cardigan Bay, Wales (Figure 1.1). Current understanding of sea-level variation is limited on the west coast of Wales. The valleys of the northern Cardigan Bay, such as the Dysynni and the Mawddach, are vital as they are affected by neither isostatic subsidence nor uplift (Shennan *et al.,* 2012). Understanding coastal changes through the Holocene is vital in order to provide context for archaeology research and provide guidance for future work. The Dysynni Valley has a rich history from prehistoric times, and therefor provides a valuable case study in the application of sea-level and environmental reconstruction to the archaeology of the west coast of Wales.

Figure 1.1: Photograph of the Dysynni valley from the top of Craig yr Aderyn looking west out to sea.

1.1 Study Area

The Dysynni valley is an over-deepened U-shaped glacial valley which stretches from the Cadair Idris ridge in the northeast to Cardigan Bay in the southwest. The Dysynni is one of the many east-west valleys in the north of Cardigan Bay, including the Dyfi and the Mawddach. The steep sided Dysynni valley widens dramatically at Bryncrug into a wide floodplain with sand dunes and shingle banks (Figure 1.2). The Bara fault line extends from the Tal-y-llyn lake, through a narrow valley, which connects to the Dysynni through a narrow ice-breached col at Ystumanner (Pratt *et al.,* 1995). The coast from the Dysynni to the Dyfi supports a variety of environments including saltmarshes, lagoons, mudflats, reedbeds and ditches.

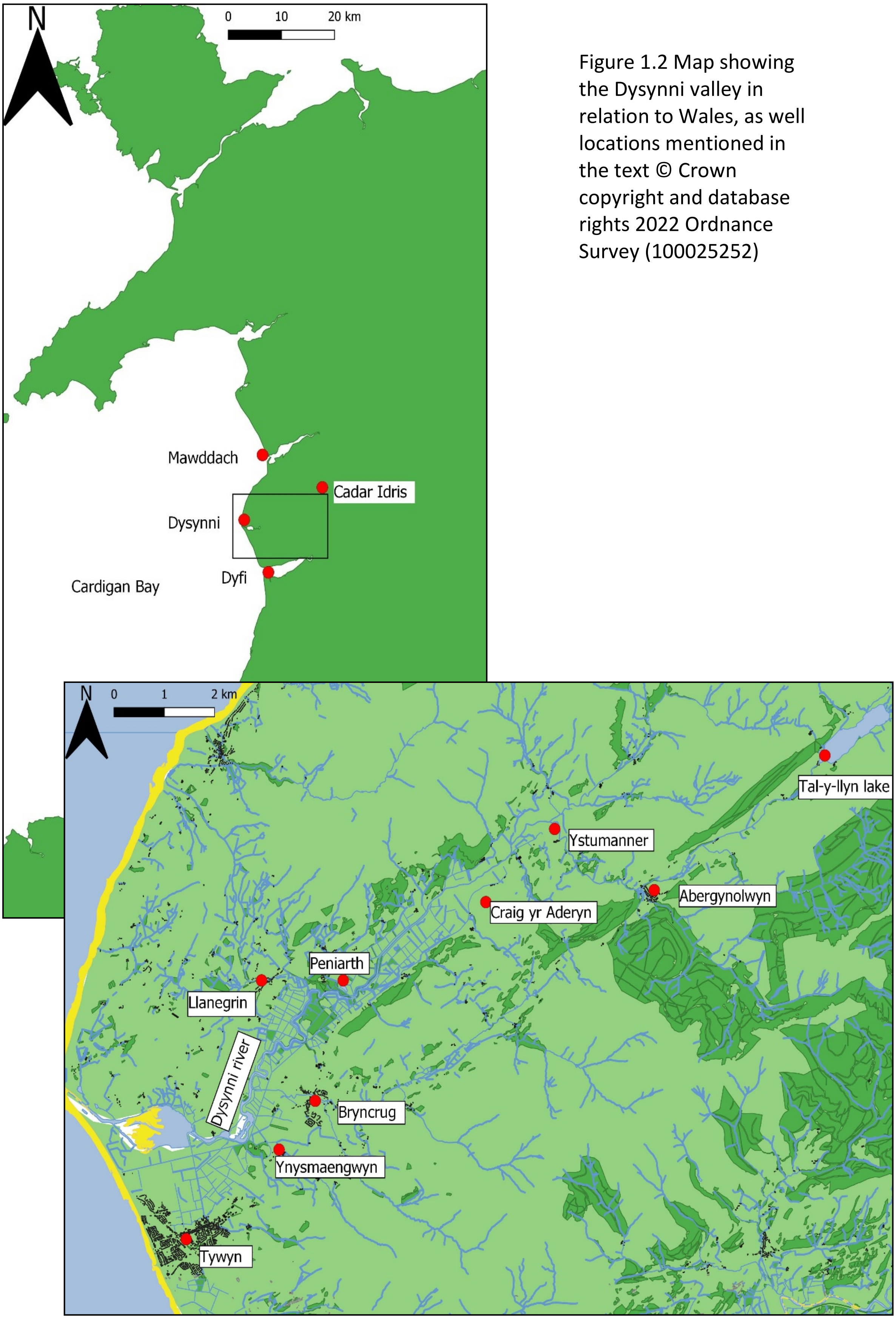

Figure 1.2 Map showing the Dysynni valley in relation to Wales, as well locations mentioned in the text © Crown copyright and database rights 2022 Ordnance Survey (100025252)

The bedrock of the Dysynni is largely Ordovician mudstones, sandstones, and siltstones, intermixed with volcanic intrusions such as the peak of Craig yr Aderyn (Pratt *et al.,* 1995) (Figure 1.3). There are extensive glacial features and deposits, including the drumlin at Tywyn and the moraine ridges in the valley bottom. Extensive silts and peats are present in the valley bottom, particularly in the valley northeast of Bryncrug (Pratt *et al.,* 1995) (Figure 1.4).

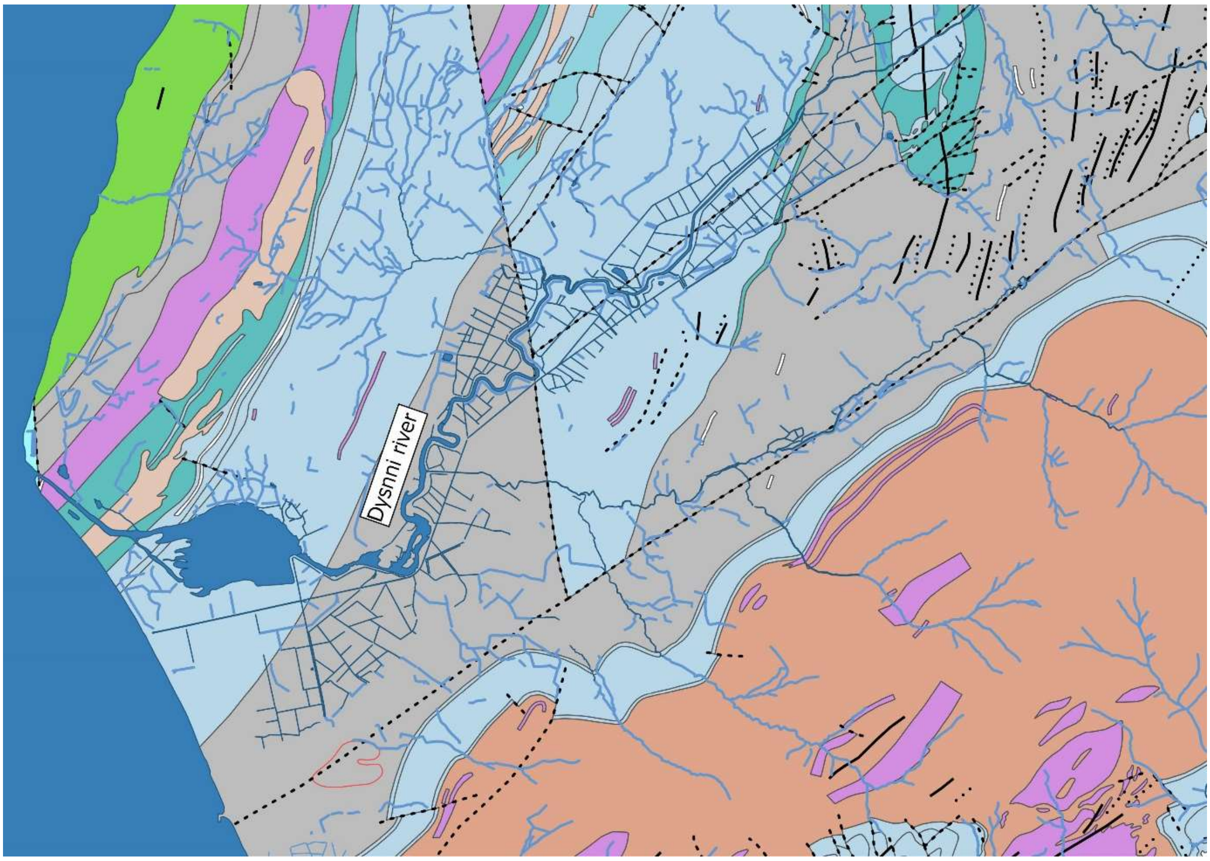

Figure 1.3: Map showing Bedrock geology of the Dysynni Valley. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Geological Map Data BGS © UKRI 2022Geological Map Data BGS © UKRI 2022

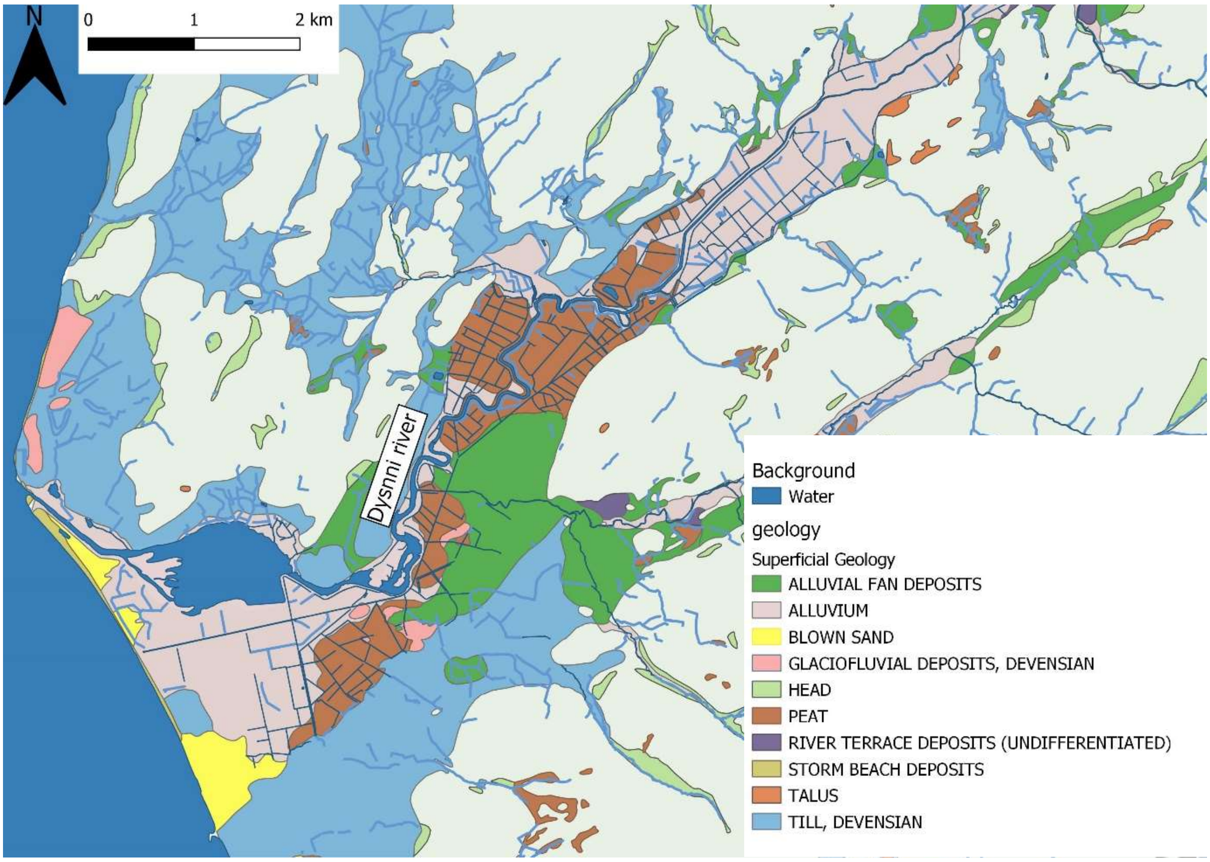

Figure 1.4: Map showing the superficial geology of the Dysynni Valley. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Geological Map Data BGS © UKRI 2022

The farmland in the valley bottom is primarily the large, regular enclosure typical of eighteenth and nineteenth century coastal reclamation, which gives way to smaller, irregular enclosure in the upland valleys (Frost, 2012). The steep open sides of the Dysynni are largely sheep-grazed ffridd[1] divided by stone walls. The vegetation of the valley is largely pasture, divided by drainage ditches, fences and hedgerows with frequent hedgerow trees in the upland valleys. There are areas of woodland and plantation, particularly along the valley sides. The slopes of Craig yr Aderyn consist largely of unimproved acid grasslands and bracken with nationally important populations of cormorants and chough nesting on the cliffs (Countryside Council for Wales, 2008).

The Dysynni is a landscape of Special Historic Importance, identified by Cadw as one of the best examples of historic landscapes in Wales (Cadw, 2022). There are many significant sites, such as the

[1] In north Wales, 'ffridd' means land with no borders or land in a large field surrounded by a wall or fence.

Iron Age hillfort at Craig yr Aderyn and the medieval Castell y Bere. There is evidence for prehistoric activity including Neolithic stone axes and Bronze Age finds in the valley bottom, and an extensive area of cairns and standing stones in the uplands (Smith, 2004a; Frost, 2012). A series of undated cropmarks at Bryncrug are likely to be either prehistoric or early medieval (Cook, 2019). An early church community was established at Tywyn, which developed into a medieval town, along with the villages at Bryncrug, Llanegrin and Abergynolwyn (Frost, 2012). The post-medieval gentry estates at Peniarth and Ynysmaengwyn were instrumental in orchestrating valley wide land improvements including enclosure, drainage, and the canalisation of the river (Frost, 2012). There is also a rich industrial history in the Dysynni, with the narrow-gauge Tal-y-llyn railway and the remains of slate quarries as well as lead and tin mines (Frost, 2012). The landscape of the valley today reflects the land improvements in the post medieval periods, and it is unclear how the landscape has changed through the dynamic sea-level rise of the Holocene.

1.2 Location of study sites

Sampling was undertaken at three locations (Figure 1.5). The site of Penllyn is located at the coast to the south of Tywyn, where the southern edge of the drumlin gives way to a marsh which is bordered by sand dunes and gravel storm banks. This site was located following the British Geological Survey (BGS) survey by Leng and Pratt (1985) which identified a blue-grey clay, with potential for microfossil preservation. Gouge surveys up the valley identified a similar blue-grey silt in the valley northeast of Bryncrug. Perfeddnant is located 7km inland, in the base of the U-shaped narrow valley and represents the full sequence of basal peat, estuarine silt and overlying woody peat. At the base of Craig yr Aderyn, the site of Gesail was chosen as Pratt *et al.* (1995) documented estuarine sediments underlying freshwater silt, but no estuarine sediments were found. Above Craig yr Aderyn the valley rises above 10m OD, and it is unlikely sea level would have been above 10m OD in this region during the Holocene (Lambeck *et al.,* 2014).

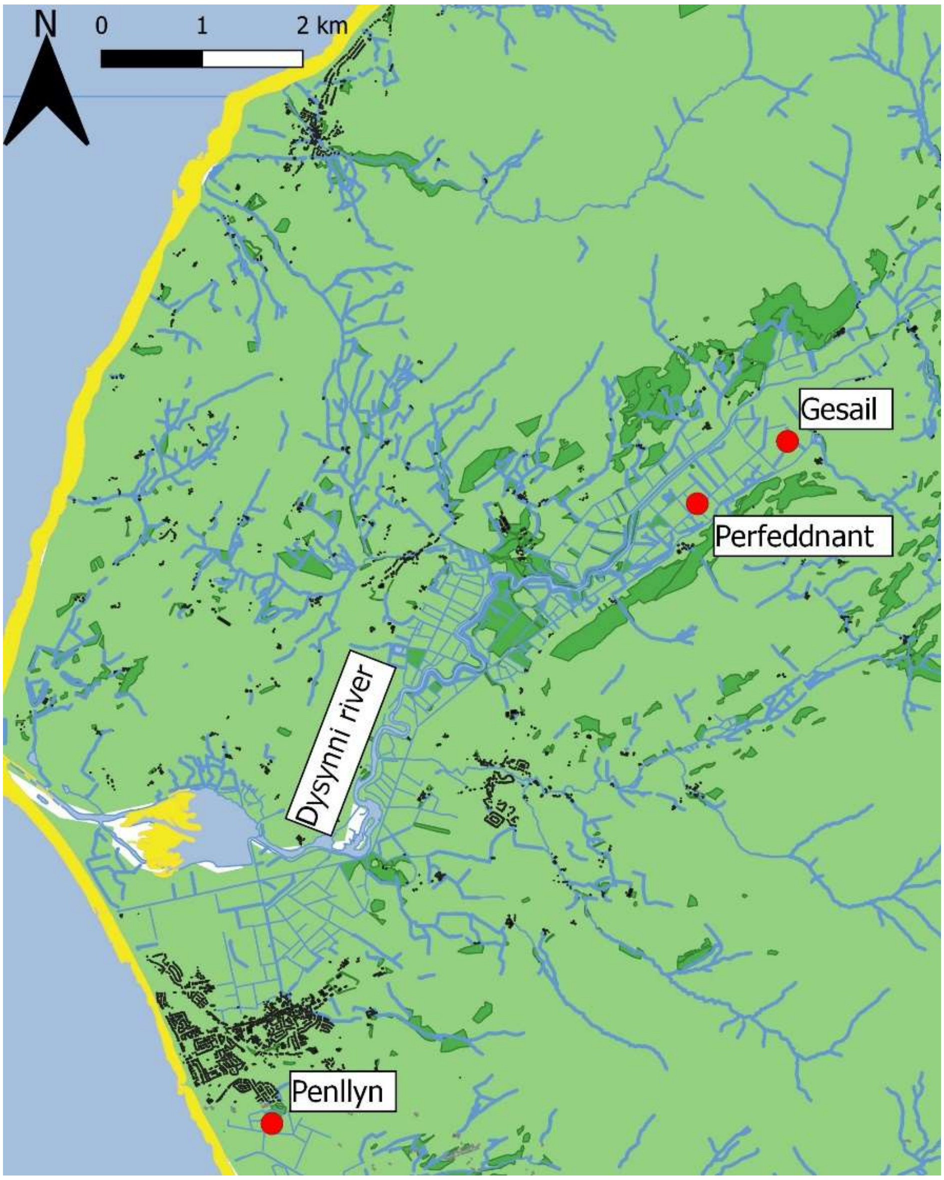

Figure 1.5: Map of study sites. © Crown copyright and database rights 2022 Ordnance Survey (100025252).

1.3 Justification for research

Paleoenvironmental reconstructions through the Holocene provide important contributions to models of future sea-level change (see Shennan *et al.,* 2018). In west Wales sea-level research is limited to sites at Borth, Ynyslas and Clarach (Adam and Haynes, 1965; Godwin, 1943; Godwin and Newton, 1938; Godwin and Willis, 1961; Godwin and Willis, 1964; Haynes and Dobson, 1969; Heyworth and Kidson, 1982; Heyworth *et al.,* 1985; Prince, 1988; Taylor, 1973; Wilks, 1977, 1979). The deep, estuarine sediments prevalent in the valleys of northern Cardigan Bay, such as the Dysynni or the Mawddach (Blundell, 1969), are valuable resources for sea-level research but remain under-studied. The Dysynni is one such valley that in addition has been unaffected by isostatic uplift or subsidence following deglaciation (Shennan *et al.,* 2012). Isostatic uplift in the north of Scotland, and subsidence across the south coast of England produces differential sea-level records across the UK. This can lead to discrepancies in GIA[2] models: for example, some GIA models predict a mid-Holocene high stand for the north coast of Wales (e.g., Peltier 1994, 2004; Peltier et al., 2002), although other GIA models do not identify this (e.g., Shennan *et al.,* 2018). More sea-level data are needed from northern Cardigan Bay to clarify the pattern of Holocene sea-level change. This research will address this through multiproxy analysis of sediments to establish the vertical and horizontal extent of sea-level change in the Dysynni Valley.

The Dysynni has a rich history stretching back to the last glacial period, with extensive archaeological remains. However, there is a distinct difference between the archaeology of the uplands and the valley lowlands, which remains unresolved. The valley bottom is filled with large, regular fields and a straight, canalised river, which reflects the land improvements of the post-medieval period. Echoes of the past landscape can be seen in the palaeo-river channels in the valley bottom, as well as place names such as "Glan-y-morfa", meaning coastal marsh, which now lie well inland. Furthermore, Craig yr Aderyn bears the UKs largest inland cormorant nesting site, which is quoted as evidence that the sea formerly reached the base of Craig yr Aderyn (Countryside Council for Wales, 2008). There is a rich archaeological history of the Dysynni, with many prehistoric sites such as three Iron Age hillforts and many cairns in the uplands (Frost 2012), however there is very little evidence to date of prehistoric land use in the valley bottom. Multiproxy environmental reconstruction of the dynamic coastal landscapes in the Dysynni is essential in order to fully understand the history of settlement and land use in the valley, as well as contribute to future archaeological investigations. Diatom and foraminifera analysis of material retrieved from three sites in the Dysynni valley will enable the

[2] glacio-isostatic adjustment

reconstruction of sea-level change. Pollen analysis will be undertaken to understand coastal change and to elucidate the potential human impact and land use in the Dysynni.

1.4 Aims and objectives

The primary aim of the research is to reconstruct sea-level change through the Holocene in the Dysynni Valley. Three objectives have been identified:

1. To reconstruct the Holocene sea-level changes in the Dysynni Valley.
2. To recreate the environmental and vegetation changes in the Dysynni Valley in order to fully understand past changes in the coastal environment.
3. To examine the impacts that the changing coastal environment had on human settlement in the area.

Sea-level change in the Dysynni will be reconstructed through sedimentological and biostratigraphical analyses of sediments. Sea-level index points (SLIPs) will be established enabling data from the Dysynni valley to be compared to the regional pattern of sea-level change for Wales. Biostratigraphic analyses will also be used to reconstruct the environmental and vegetation changes in the Dysynni. The results of the sea-level, coastal and vegetation reconstructions will be compared to the archaeological record in order to provide insight into human response to sea-level and coastal change.

1.5 Thesis outline

The thesis is organised into three sections (Figure 1.6). Section A outlines the context of the research, with Chapter 1 introducing the project. Chapter 2 reviews the Devensian glaciation in the UK and Wales before defining the concepts of sea-level change. The sea-level history of Wales is reviewed followed by the palynological research for Wales and the archaeology of the Dysynni Valley. Chapter 3 describes the methods and techniques of analysis including fieldwork, laboratory work and statistical analyses. Section B includes the presentation of results, with one chapter for each site (Penllyn, Perfeddnant and Gesail) presenting the fieldwork, laboratory work, archaeological data, and statistical analysis results for each site. Section C includes the conclusions of the research with Chapter 7 discussing the sea-level and coastal change as well as reviewing the archaeology. Chapter 8 concludes the research.

<table>
<tr><th colspan="3">Section A: Context of research</th></tr>
<tr><td colspan="3">Chapter 1: Introduction</td></tr>
<tr><td colspan="3">Chapter 2: Research context
Reviewing the glacial, sea-level and palynological research of Wales. The archaeology of the Dysynni is summarised.</td></tr>
<tr><td colspan="3">Chapter 3: Methodology and techniques of analysis
Describing the methods used including fieldwork, laboratory, and statistical analyses</td></tr>
<tr><td colspan="3">Section B: Presentation of results
Reviewing the results for each site including fieldwork, laboratory, and statistical analyses</td></tr>
<tr><td>Chapter 3: Penllyn</td><td>Chapter 4: Perfeddnant</td><td>Chapter 5: Gesail</td></tr>
<tr><td colspan="3">Section C: Conclusions of research</td></tr>
<tr><td colspan="3">Chapter 7: Discussion
Discussion of the sea-level change, coastal change, and archaeology in context of</td></tr>
<tr><td colspan="3">Chapter 8: Conclusions</td></tr>
</table>

Figure 1.6: Thesis outline

Chapter 2: Literature Review

2.1 Introduction

This chapter reviews the literature on the late Quaternary glacial history and sea level changes in the UK. It presents a detailed account of the glaciation of the UK and specifically Wales. This will be followed by an outline of the fundamental concepts of eustasy[3], isostasy[4] and relative sea-level change. Sea-level records for Wales are then reviewed and presented chronologically. Finally, the current research into palynology and vegetation changes for the Holocene is presented for northwest Wales.

2.2 Devensian Glaciation of the UK and Wales

This section will outline the Devensian glaciation in the UK. The Devensian glaciation of Wales will be described in detail.

2.2.1 The Devensian Glacial History of the UK

The Last Glacial Maximum (LGM) is defined as the most recent period during Marine Isotope Stage 2, when global ice sheets reached their maximum extent (Figure 2.7) (CLIMAP, 1976, 1981; Chiverrell and Thomas, 2010; Mix *et al*., 2001) around 26000-21000 BP (Peltier and Fairbanks, 2006; Patton *et al*., 2013 a, b). Patton *et al.* (2013 a, b) modelled the LGM and determined that the maximum glacial extent developed quickly in the north and west of the UK by 23500 BP. Absolute dates are not widely available, but it is probable that the maximum extent of ice at the LGM was not reached simultaneously across the UK (Bickerdike *et al*., 2016; Ballantyne, 2012; Golledge *et al*., 2010).

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[3] The world-wide change of sea level throughout, influenced by factors such as the melting of glaciers.
[4] the equilibrium between parts of the earth's crust, rising if mass is removed and sinking if mass is deposited.

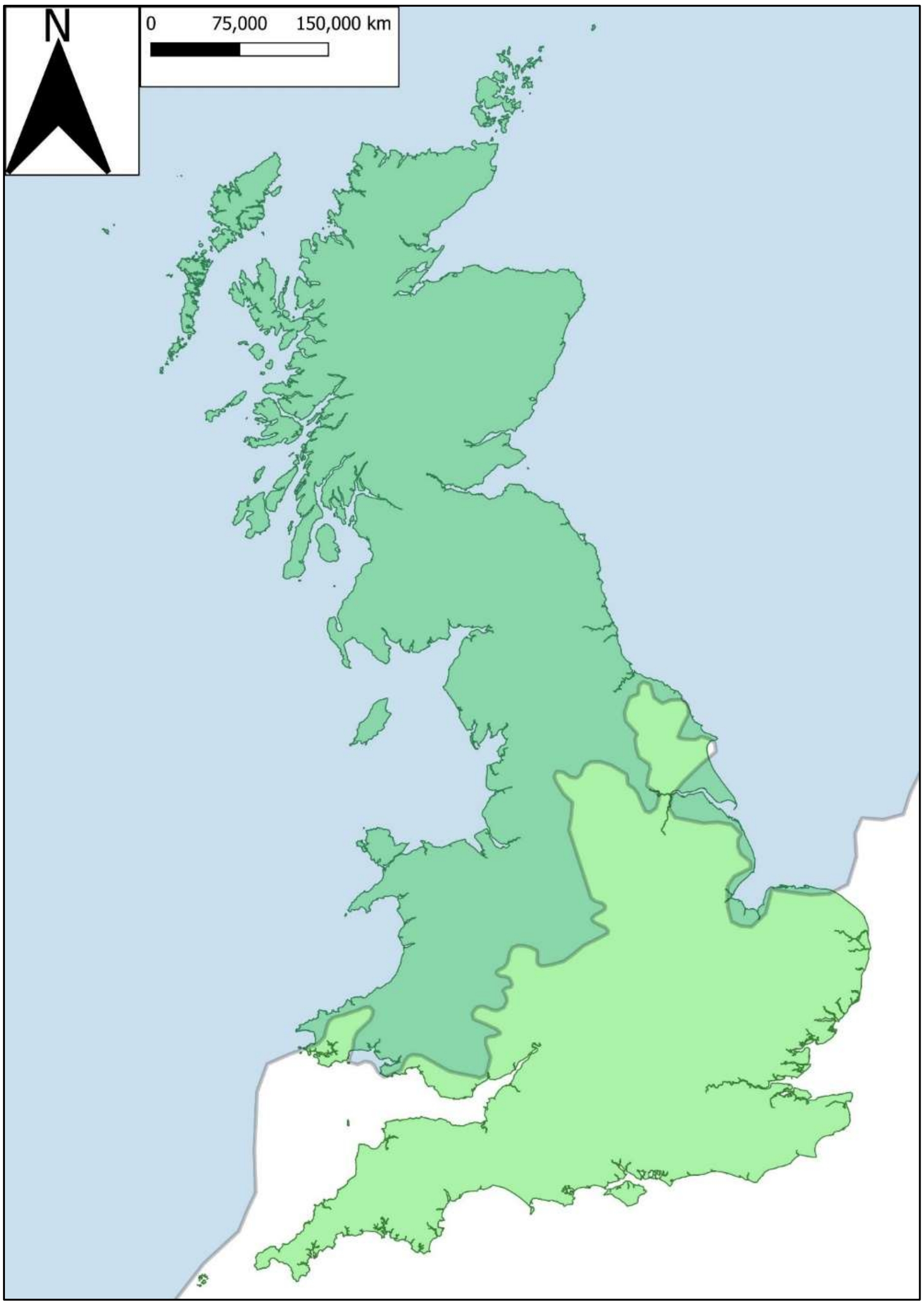

Figure 2.7: The full extent of the LGM for the UK indicated in blue. Compiled from Emery, 2020; Sejrup *et al.,* 2005; Bradwell *et al.,* 2008 and Sejrup *et al.,* 2016. © Crown copyright and database rights 2022 Ordnance Survey (100025252)

During the Devensian glaciation the British and Irish ice sheet consisted of four separate ice domes over Scotland, Wales, the Lake District, and Ireland (Clark *et al*., 2012; Jansson and Glasser, 2005). A predominantly marine based margin extended over the British and Irish continental shelves and at times this connected to the larger Scandinavian ice sheet in the North Sea (Clark *et al.,* 2012, 2017; Graham *et al*., 2007; Hughes *et al.,* 2014; Sejrup *et al.,* 1994).

The British and Irish ice sheet first developed over central Scotland, expanded to Northern Ireland, and then connected to an ice dome over the Lake District (Hughes *et al.,* 2014), causing sea level to fall as the ice caps formed (Patton and Hambrey, 2009). Throughout the Devensian glaciation the Welsh ice cap existed semi-independently from the ice caps over Scotland, England, and Ireland. The Welsh ice cap accumulated over the uplands of Snowdonia and the Cambrian Mountains as well as the Brecon Beacons in south Wales (Jansson and Glasser, 2005; Patton *et al.* 2013a; Sahlin, 2009). The thickest ice cap was located over Cadair Idris, between the Rhinog Mountains and Arenig Fawr (Campbell and Bowen, 1989; Foster, 1968, 1970; Hughes, 2002; Sahlin, 2009). There is evidence from ice flow patterns that the Welsh ice cap developed prior to the LGM (Jansson and Glasser, 2005; Hughes *et al.,* 2014). This is also reflected in numerical models by Patton *et al.* (2013 a, b), who modelled the Welsh ice cap in terms of ice flow dynamics, climate and the timing of events such as glacial advance and retreat. The model showed two large advances, the first at approximately 29100 BP and the second at the LGM (Hughes *et al.,* 2014; Patton *et al.,* 2013 a, b). During the second advance the Welsh ice cap joined the larger British and Irish ice sheet and expanded to the south coast of Wales (Jansson and Glasser, 2005).

As the LGM progressed, the British and Irish ice sheet joined the Scandinavian ice sheet in the North Sea (Hughes *et al*., 2014). The Irish Sea ice stream developed and began to move south. The Irish Sea ice stream was formed of ice from Scotland, Ireland, and the Lake District. The ice stream flowed south from Scotland through the Irish Sea basin and overran the western tip of the Llŷn Peninsula in Wales (Jansson and Glasser, 2005; Patton *et al.,* 2013 a, b). The Welsh ice cap expanded to meet the Irish Sea ice stream along the north coast of Wales (Glasser *et al*., 2001; Hambrey *et al.,* 2001; Thomas *et al.,* 1998; Thomas and Chiverrell, 2007). There is some evidence that the Irish Sea ice stream may have reached as far south as the Isles of Scilly (Hambrey *et al*., 2001; Hiemstra *et al.,* 2006; Hubbard *et al.,* 2009; Patton and Hambrey, 2009; Patton *et al.,* 2013a; Scourse and Furze, 2001).

Upon deglaciation, the ice in the North Sea and the Irish Sea ice stream retreated rapidly leading to the development of glacial lakes, such as at Llyn Teifi (Hambrey *et al.,* 2001) as well as smaller lakes in the Tonfannau and the Dysynni Valley (Patton and Hambrey, 2009). As the Irish Sea ice stream retreated, the British and Irish ice sheet decoupled from the Welsh ice cap, and the ice sheets shrunk down to centres over Wales, the Lake District and Scotland, before finally retreating into the Scottish Highlands (Hughes *et al.,* 2014). Following separation from the British and Irish ice sheet the dynamics of the Welsh ice cap changed, as ice flowed quickly through major valleys to the west and south, draining the Welsh ice cap (Jansson and Glasser, 2005).

The chronology of the Devensian glaciation is unclear, with very little evidence from absolute dating. Along the Aran Ridge Glasser *et al.* (2012) undertook ^{10}Be and ^{26}Al exposure-age dating on eight bedrock samples. Six of the dates yielded ages from 17200 to 34400 BP with six samples along the summit having very close ages (17500±600 BP, 17500±700 BP, 19700±800 BP and 20000±700 BP) and a mean age of 19140±1450 BP (Glasser *et al.,* 2012). Two samples produced anomalously old ages (27500±1000 BP, 33900±1200 BP), but the results indicated that the Welsh ice cap covered the Aran ridge until the summits were exposed between approximately 20000 and 17000 BP (Glasser *et al.,* 2012). Hughes *et al.* (2016) took eight samples from the summits of the Moelwyn, Rhinog and Arenig mountains that yielded a mean age of 19080±800 BP. The samples all came from vertical elevations between 824-581m, demonstrating that the ice cap thinned very quickly and uniformly (Hughes *et al.,* 2016). The mean age for all the dates from the Aran ridge and the Moelwyn, Rhinog and Arenig is 19210±1070 BP, which suggests that the summits of North Wales were exposed as nunataks[5] between 20000-19000 BP (Glasser *et al.,* 2012; Hughes *et al.,* 2016).

[5] the summit or ridge of a mountain that protrudes from a glacier that otherwise covers most of the mountain

Following deglaciation there was a return to glacial conditions during the Loch Lomond Stadial, or Younger Dryas, between 12900-11700 BP. Ice developed in upland areas of the UK, the most significant of which was the ice field in the Western Highlands of Scotland (Bickerdike *et al.,* 2016; Bennett and Boulton, 1993). The ice cover during the Loch Lomond stadial may have been more extensive for the UK than previously thought (Bickerdike *et al.,* 2016; Brown *et al.,* 2011, 2013; Lukas and Bradwell, 2010; McDougall, 2013). There were many smaller ice fields, ice caps, valley glaciers and cirques[6] or niche glaciers in the upland areas of the UK and Ireland (Bickerdike *et al*., 2016; Golledge, 2010). In Snowdonia there were several cirque glaciers which have been associated with the Loch Lomond Stadial, with 38 identified in total (Figure 2.8) (Evans, 1999; Bendle and Gasser, 2012; Gray and Lowe, 1982). There is also evidence for cirque glaciation elsewhere in North Wales during the Loch Lomond stadial, such as the Aran ridge (Hughes, 2002), the Berwyn mountains (Hughes, 2009) and Cadair Idris (Lowe, 1994).

[6] glaciers formed in a cirque, a bowl-shaped depression on the side of mountains.

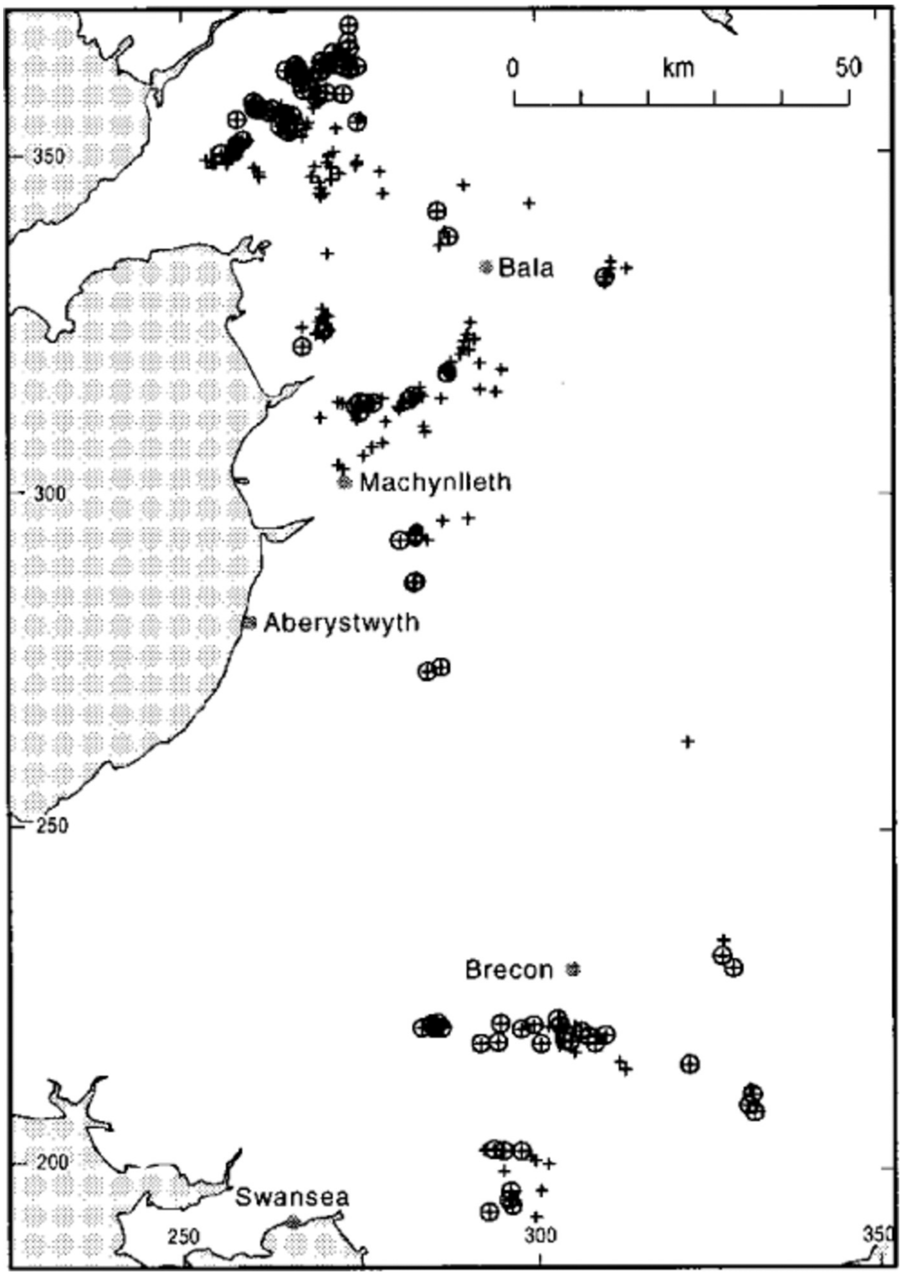

Figure 2.8: Distribution of cirques in Wales, indicated by crosses. Circles indicate possible late-glacial age. Reproduced from Evans, 1999.

Despite being relatively small, with estimates ranging from 357000 km^2 to 840000 km^2, the British and Irish ice sheet had a dramatic impact on sea-level change throughout the Holocene in the UK, affecting isostasy and land movement (Roberts *et al.,* 2011; Shennan *et al.,* 2009; Shennan *et al.,* 2012). There have been many numerical models of glacio-isostatic adjustment (GIA) developed which incorporate reconstructions of the ice sheets at the LGM (e.g., Bradley *et al.,* 2011), the response of the earth's mantle (e.g., Bradley *et al.,* 2009) and the volume of water in the world's oceans (e.g., Uehara *et al.,* 2006) (Shennan *et al.,* 2009; Shennan *et al.,* 2012). To test the accuracy of the models, comparisons are made to reconstructed relative sea level records and the model is adjusted until a good fit is obtained (Whitehouse and Bradley, 2013). There have been many GIA modelling studies of the UK (Brooks *et al*., 2008; Bradley *et al.,* 2009; Bradley *et al.,* 2011, Lambeck 1993, 1995, 1996; Peltier 1994, 2004; Peltier *et al.,* 2002; Shennan *et al*., 2006; Shennan *et al.* ,2012, 2018). Many models, however, predict that sea level would have been higher than present levels in areas such as north Wales, where this does not appear to fit the evidence (Peltier, 1994; 2004; Shennan *et al.,* 2006; Brooks *et al.,* 2006). The most recent GIA models by Bradley *et al.* (2011) and Shennan *et al.* (2018) provide the best fit for relative sea level records.

Shennan *et al*. (2012) produced a map of relative land uplift and subsidence (Figure 2.9) based on the GIA model of Bradley *et al*. (2011) with relative uplift represented by positive values and relative subsidence as negative. The maximum uplift is centred over Scotland, reflecting the continued impact of glacio-isostatic adjustment following the melting of the British and Irish ice sheet (Shennan *et al.,* 2012). However, there are three centres of relative subsidence in southwest England, the southern North Sea and the Shetland Isles, caused by other factors such as forebulge[7] collapse, meltwater pulses, the load of water on the Atlantic Ocean, and the proximity of the Fenno-Scandinavian ice mass (Shennan *et al.,* 2012).

[7] a flexural bulge as a result of a load on the lithosphere, often caused by glaciation

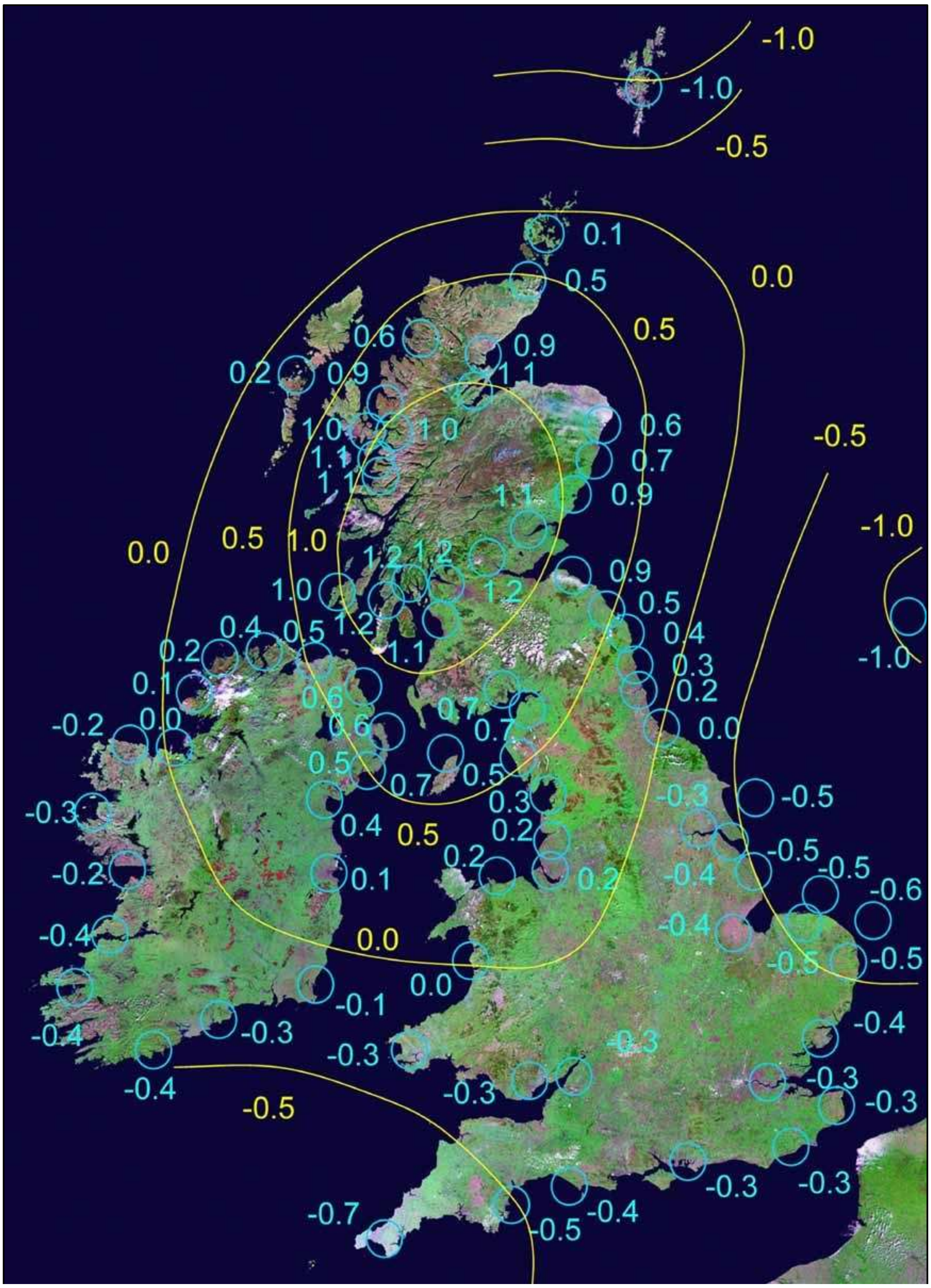

Figure 2.9: The rate of relative land-level change 10000 BP to the present day in Britain and Ireland (mm $^{a-1}$ [8]). Relative land uplift is shown as positive and relative subsidence as negative. The background image is 900x1300 km, courtesy of NASA, acquired under the scientific data purchase programme (from Shennan *et al.,* 2012). Blue numbers indicate rates of relative land-level change at specific locations, with yellow contour lines indicating general trends.

[8] Millimetres per year.

2.2.2 The Devensian Glacial History of Wales

There is a great deal of geomorphological evidence for Devensian glaciation in Wales. For example, in the north where the Welsh ice cap developed over the uplands in Snowdonia during the LGM and flowed through valleys to the west, east and north of Snowdonia (Figure 2.10) (Clarke *et al.,* 2018; Evans *et al.,* 2005). The Welsh ice cap met the Irish Sea ice stream during the LGM, with ice flowing from the Irish Sea towards the north and west coasts as well as the east in the Cheshire-Shropshire lowlands, leading to complex features where the two ice sheets interacted (Clarke *et al.,* 2018; Evans *et al.,* 2005). For instance, deposits of till from the Irish and Welsh Ice sheets record multiple advance and re-advance episodes in areas such as the Llŷn Peninsula and Anglesey (Clarke *et al.,* 2018; Evans *et al.,* 2005). In the south there are fewer geomorphological glacial features, and it is unclear where the southern limit of the ice would have been, although it has been argued that ice reached the south coast of Wales (Clarke *et al.,* 2018; Evans *et al.,* 2005).

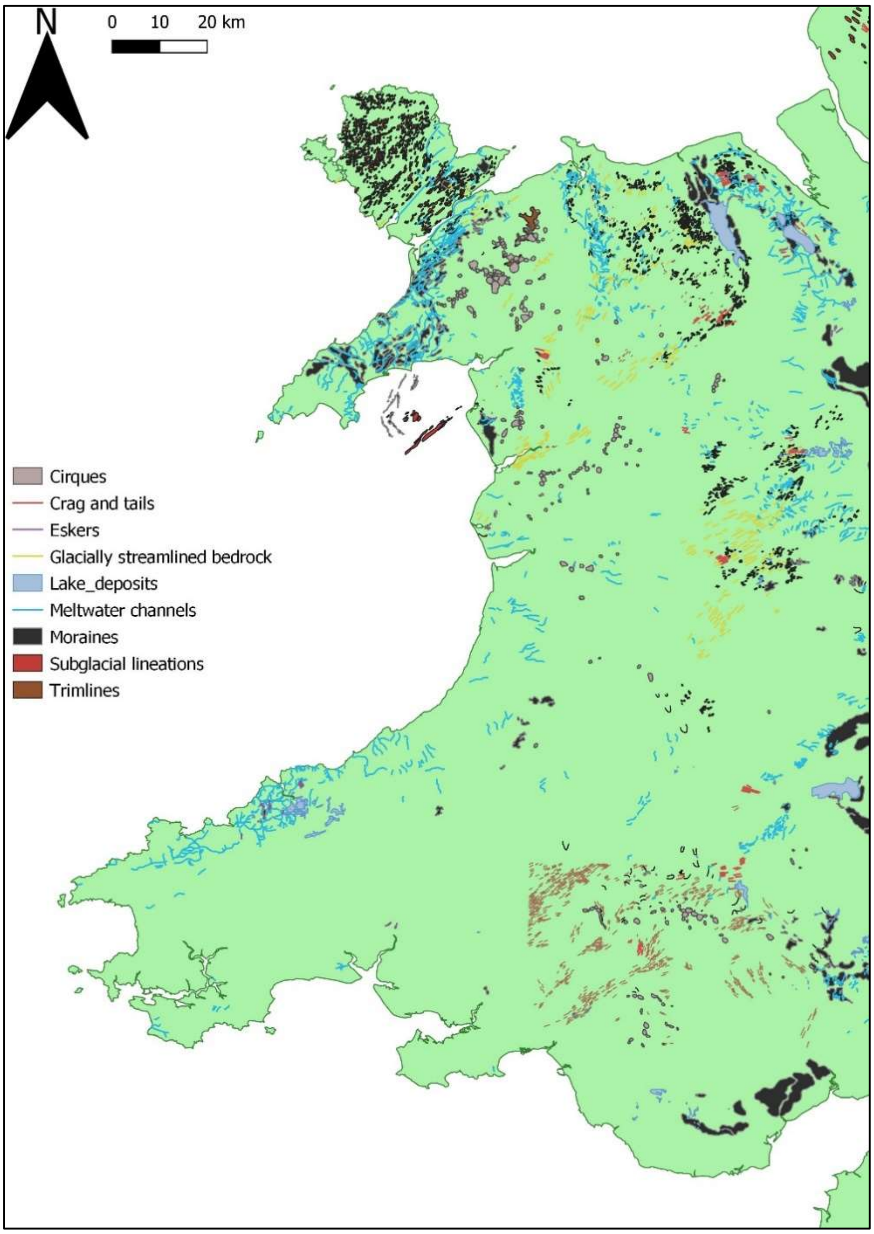

Figure 2.10: Map of Wales showing glacial features from BRITICE (Data sources from Clarke *et al.,* 2018). © Crown copyright and database rights 2022 Ordnance Survey (100025252)

2.2.2.1: Welsh Marches and the Cheshire Plains

In the Cheshire-Shropshire lowlands, the Welsh ice cap met the ice caps from the Irish Sea ice stream and the British ice sheet over the Pennines. To the west of the Pennines, valley glaciers drained the British ice cap towards the southwest, until a large lobe of the Irish Sea ice stream penetrated into the Manchester lowlands and south into the Cheshire-Shropshire lowlands (Figure 2.11) (Chiverrell and Thomas, 2010; Wills, 1924).

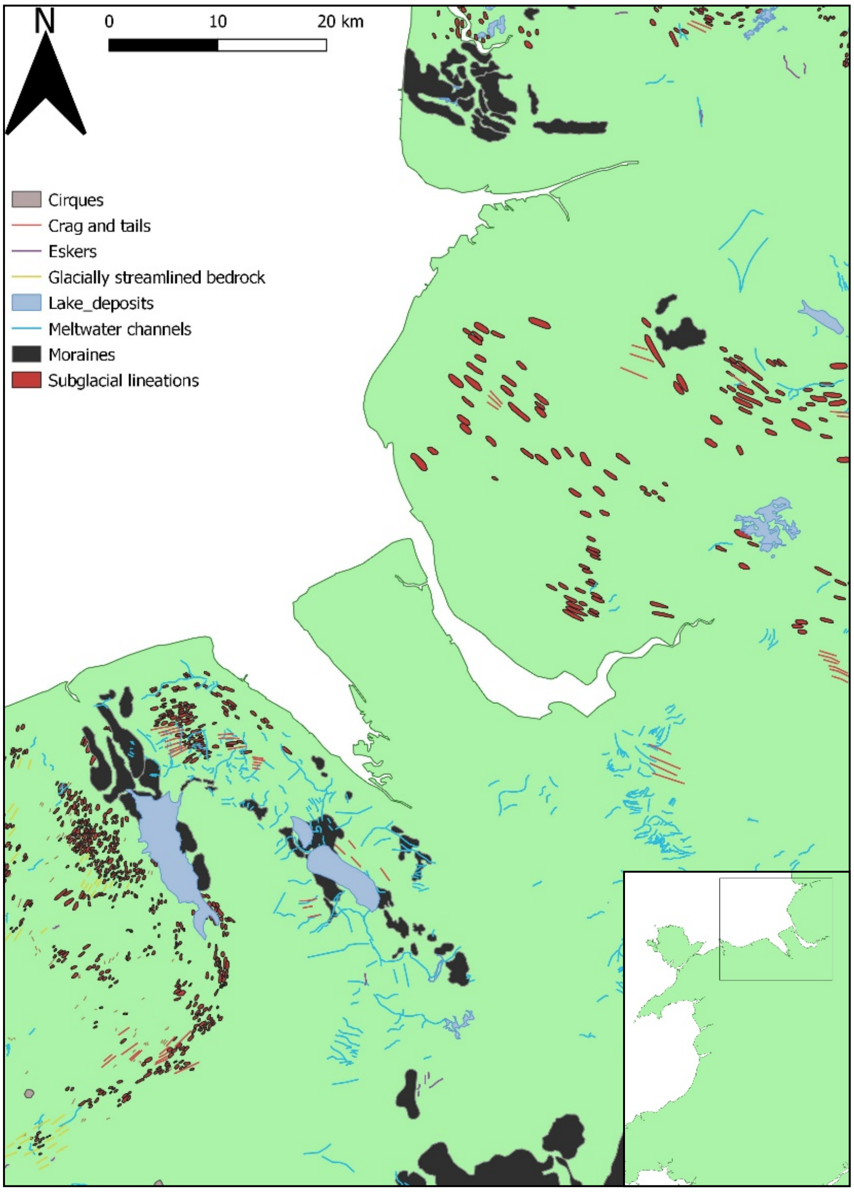

Figure 2.11: Map of east Wales showing glacial features from BRITICE (Data sources from Clarke *et al*., 2018). © Crown copyright and database rights 2022 Ordnance Survey (100025252)

After the retreat of the Irish Sea ice stream, the Welsh ice cap advanced into the lowlands as a series of piedmont glaciers[9], however there is very little dating evidence (Chiverrell and Thomas, 2010). There is a large moraine[10] ridge known as the Wrexham-Ellesmere-Whitchurch-Barhill Moraine which curves across the Cheshire-Shropshire lowlands in a series of ridges (Boulton and Worsley, 1965; Paul 1983; Paul and Little, 1991; Thomas, 1989; Thomas and Chiverrell, 2010; Worlsey, 1967,

[9] a valley glacier which has spilled out onto relatively flat plains, spreading into bulb-like lobes.
[10] a mass of rocks and sediment carried down and deposited by a glacier, typically as ridges.

1970). There is evidence for glacial lakes, originally one large glacial lake known as Lake Lapwas was identified, stretching from Manchester to Wrexham (Murton and Murton, 2012). More recently, a series of lakes have been found including Lake Prees (20x5km) and Lake Bangor (30x10km) (Murton and Murton, 2012; Thomas, 1985, 1989).

2.2.2.2 Anglesey and the Llŷn Peninsula

The Late Devensian deposits along the coast on Anglesey and the Llŷn Peninsula reveal evidence for the advance and retreat of the eastern edge of the Irish Sea ice stream which met, joined, and then decoupled from the Welsh ice sheet (Figure 2.12) (Thomas *et al.,* 1998; Thomas and Chiverrell, 2007). There are two distinct diamicts[11] which characterise the glacial sediments of Anglesey and the Llŷn Peninsula, an Irish Sea ice stream diamict from the north Irish Sea and a Welsh diamict from Snowdon (Thomas and Chiverrell, 2007; Thomas *et al.,* 1998).

[11] a sediment resulting from dry-land erosion that is unsorted to poorly sorted and contains particles ranging in size from clay to boulders, suspended in an unconsolidated matrix of mud or sand.

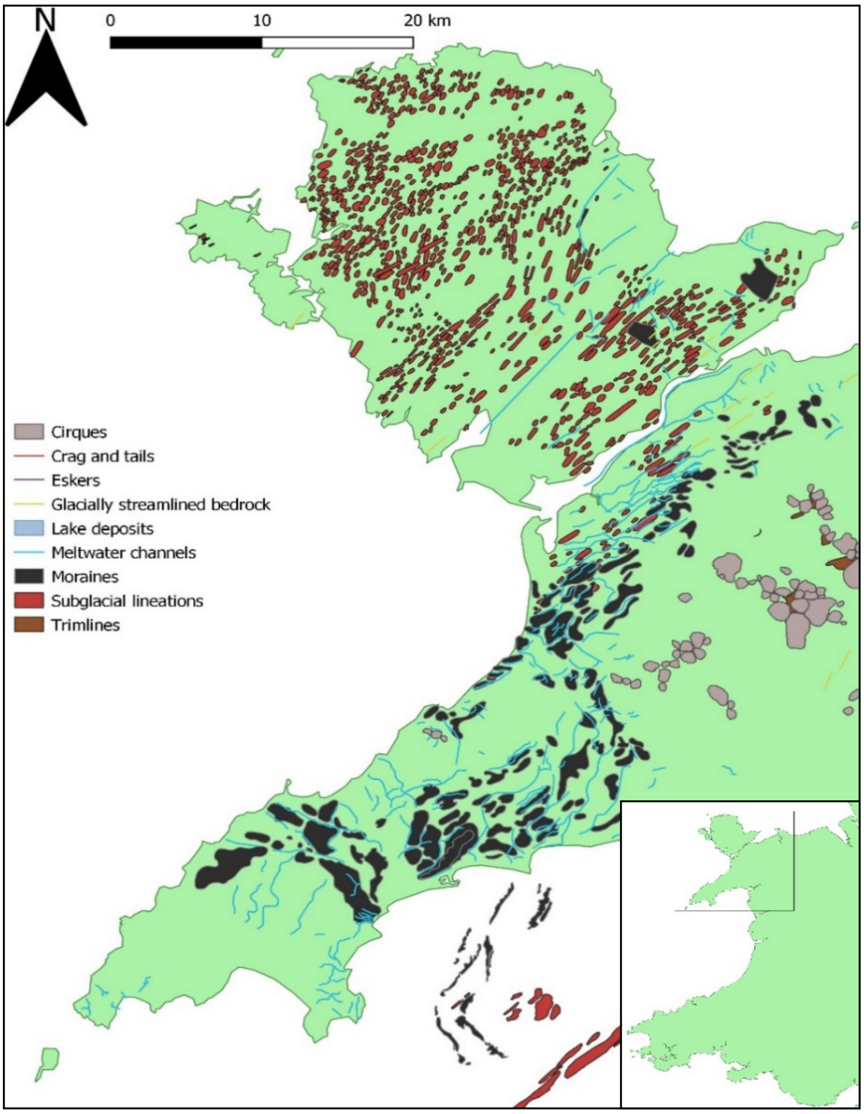

Figure 2.12: Map showing the glacial features of the Llŷn peninsula and Anglesey from BRITICE (Data sources from Clarke *et al.,* 2018). © Crown copyright and database rights 2022 Ordnance Survey (100025252)

On Anglesey, Thomas and Chiverrell (2007) identified three geomorphological assemblage zones from the pattern of drumlins, which indicate the flow direction and the effect of sub-glacial action (Figure 2.12) (Phillips *et al.,* 2010; White and Beamish, 2014). Philips *et al.* (2010) found marked differences in the size and shape of the features in each zone which closely reflected bedrock geology. Philips *et al.* (2013) found that the diamict on Anglesey had undergone both periglacial[12]

[12] processes that result from seasonal thawing of snow in areas of permafrost, the runoff from which refreezes in ice wedges and other structures.

and glaciotectonic deformation[13], indicating that Anglesey was covered by permafrost[14], not ice, prior to the arrival of the Irish Sea ice stream.

Thomas *et al.* (1998) proposed that during the Late Devensian, ice from Snowdonia moved southwest across Tremadoc Bay. The ice reached north to Tai Morfa, on the eastern Llŷn Peninsula, where the Welsh ice joined the Irish Sea ice stream moving south across the north Llŷn coast (Crimes *et al.,* 1992; Thomas *et al.* 1998). Thomas *et al.* (1998) estimated that maximum glaciation was reached by approximately 18000 BP with the Llŷn peninsula covered in ice. During deglaciation, the two ice sheets uncoupled across the coast between Pwllheli and Criccieth prior to 14468 BP (Thomas *et al.,* 1998). The separation of the ice sheets led to meltwater drainage and glaciofluvial sedimentation south of Pant-y-glas, in particular the Rhoslan Fan, which has been interpreted as an ice front alluvial fan and sandur[15] (Thomas *et al.,* 1998). The meltwater produced by deglaciation formed an ice marginal lake which drained southwards and the Rhoslan fan was abandoned (Thomas *et al.,* 1998). Thomas and Chiverrell (2007) identified 30 ice marginal limits in the Llŷn peninsula, eleven of which were re-advance episodes. The high numbers of re-advance episodes were endemic to the ice margins of retreating ice sheets, influenced by local conditions such as topography and bedrock, rather than regional climate (Chiverrell and Thomas, 2010; Thomas and Chiverrell, 2007).

2.2.2.3 Snowdonia

In the upland areas of Wales there is limited evidence for glacial erosion. The uplands of Wales were covered in cold-based ice, which led to limited glacial erosion and the preservation of pre-glacial landscapes (Jansson and Glasser, 2005; Patton *et al.,* 2013a, b). McCarroll and Ballantyne (2000) found a radial flow of ice from the Snowdon massif, which was deflected to the southwest by Irish Sea ice flowing over Anglesey. Jansson and Glasser (2005) found geomorphological evidence that the ice cap was thick enough to cover the mountain summits during the LGM, but other studies

[13] The structural deformation of rock and sediment due to glacial movement or loading
[14] ground that continuously remains below 0 °C (32 °F) for two or more years
[15] an outwash plain formed by meltwater from glaciers

indicated that the ice would have been much lower, with the peaks standing as nunataks above the glacier (Lambeck, 1996; McCarrol and Ballantyne, 2000).

2.2.2.4 Valleys of the West Coast

During the LGM, ice flowed from Snowdonia through the valleys to the west, in particular the Mawddach, Dysynni and Tal-y-llyn valleys, causing glacial over-deepening (Figure 2.13) (Jansson and Glasser, 2005; Patton and Hambrey, 2009; Patton *et al.,* 2013a, b; Sahlin *et al.,* 2009). Offshore glacial diamicts are widespread in Cardigan Bay and the Irish Sea, possibly originating from the Irish sea ice stream (Garrard and Dobson, 1974). A seismic profile recorded Welsh till consistently below the Irish till, which was confirmed by a borehole offshore from the Dyfi estuary (Garrard and Dobson, 1974). In contrast, Tremadoc bay contained Welsh drift both under and over the Irish till, which may imply a re-advance of the Welsh ice after the Irish ice retreated (Garrard and Dobson, 1974). The Dysynni and the Tal-y-llyn valleys are both glacial troughs with hanging valleys and faceted spurs (Campbell and Bowen, 1989; Sahlin *et al*., 2009; Watson, 1962). At Tonfanau, near the Dysynni valley, there was evidence for six lithofacies[16], with the lowest section representing the Welsh Ice overlain by sediment from the Irish Sea ice stream (Patton and Hambrey, 2009).

[16] a subdivision of a stratigraphic unit characterised by particular lithologic features.

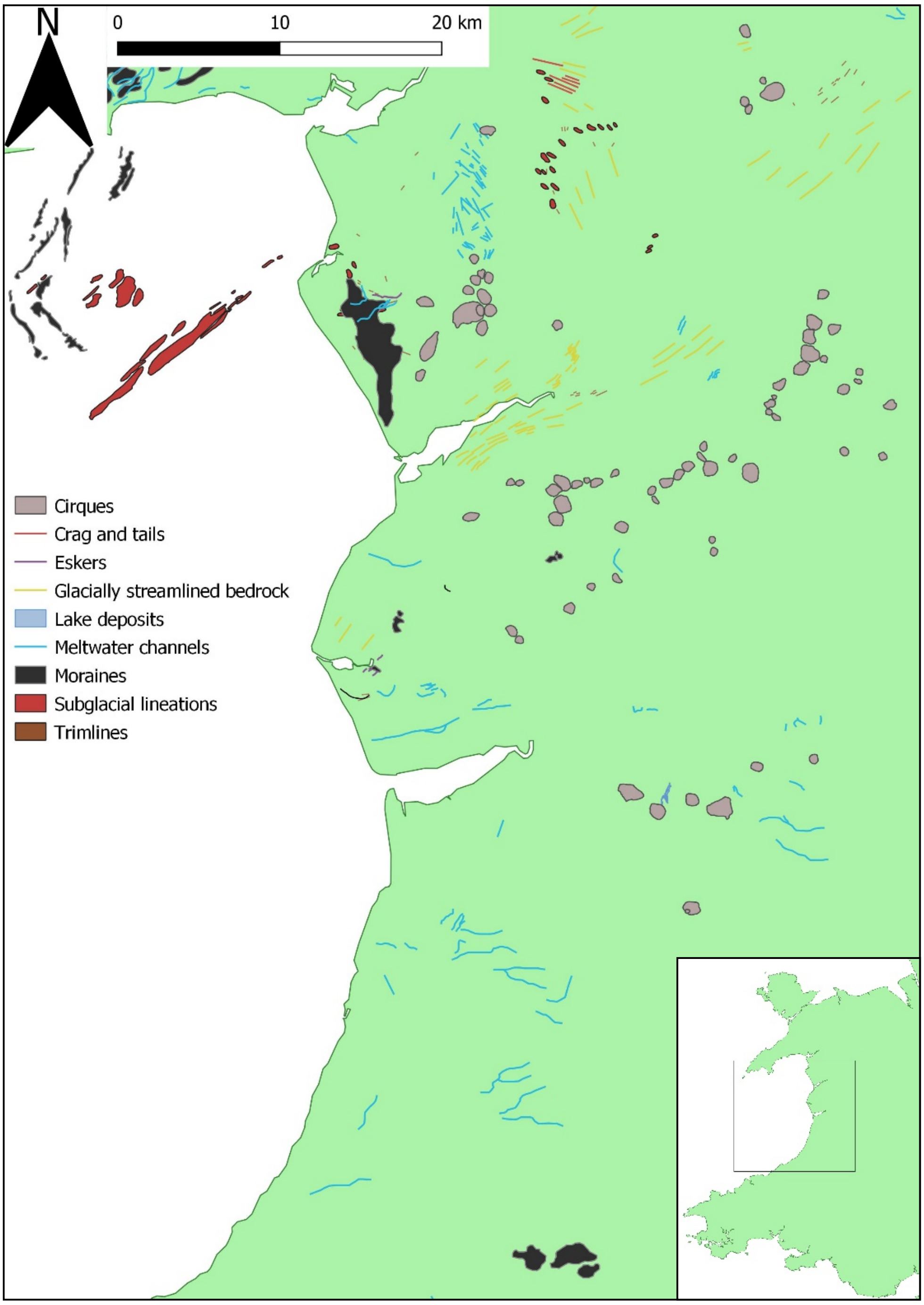

Figure 2.13: Map of west Wales showing glacial features from BRITICE (Data sources from Clarke *et al.,* 2018). © Crown copyright and database rights 2022 Ordnance Survey (100025252)

In contrast to the other valleys along the west coast of Wales, the Dyfi Valley is unusual as there is little evidence for glacial erosion (Sahlin *et al.,* 2009). The Dyfi upper catchment was shaped by glacial erosion, but the lower areas of the basin had a dendritic pattern more typical of fluvial erosion[17] (Sahlin *et al.,* 2009). The preservation of the pre-glacial, fluvial landscape implies there was little glacial erosion despite the area being subject to glacial activity throughout the Late Devensian. It is possible that either the area was covered in cold-based, slow-moving ice under an ice divide, or that the majority of ice movement was along the line of the Pennal fault which is located along the Dyfi valley (Sahlin *et al.,* 2009). The model by Patton *et al.* (2013 a, b) indicated multiple episodes of fast flowing ice in the Dyfi valley with much of the ice channelled through the Pennal fault (Patton *et al.,* 2013 a, b).

There are a number of moraines offshore in Cardigan Bay that form a series of lobes, known as sarnau[18]: Sarn Badrig, Sarn-y-Bwlch and Sarn Cynfelin (Figure 2.13) (Evans *et al.,* 2005; Foster, 1970; Garrard and Dobson, 1974; Patton and Hambrey, 2009; Patton *et al.,* 2013c; Tappin *et al.,* 1994). The sarnau form three low, smooth ridges covered in gravel, cobbles and boulders and are formed of diamict. Although there is no evidence of landward continuation, they are aligned with the valley sides onshore (Patton and Hambrey, 2009; Tappin *et al*., 1994). The sarnau are unique and the origins of these features are not fully understood (Patton *et al.,* 2013b). Foster (1970) suggested that they were deposited between the Welsh ice cap and the Irish Sea ice stream, while Allen and Jackson (1985) found an outcrop of Sarn Badrig on land, deposited by the combined glaciers which occupied the Dwyryd and Ysgethin valleys (Evans *et al.,* 2005). Garrard and Dobson (1974) interpreted the sarnau as deposits from local piedmont glaciers which reached the coast before the Irish ice arrived. However, Tappin *et al.* (1994) suggested they may instead be the remnant of a late-glacial sandur. Patton *et al.* (2013c) also proposed that the sarnau could be features that have been reworked over numerous glaciations during the Pleistocene.

2.2.2.5 South Wales

In south Wales there is less direct evidence of glacial depositional features (Figure 2.14). Early work by Charlesworth (1929) identified the South Wales end moraine through overflow channels,

[17] erosion by river
[18] 'sarn' is Welsh for causeway, plural 'sarnau'

discontinuous moraines, and postulated ice dammed lake boundaries. This work has been reviewed by many researchers who have estimated that the limit would have been further south with all of Wales covered in ice except Pembrokeshire (Bowen, 1981; Campbell and Bowen, 1989). Relatively little recent work has been done on the limit of the LGM in south Wales (Chiverrell and Thomas, 2010).

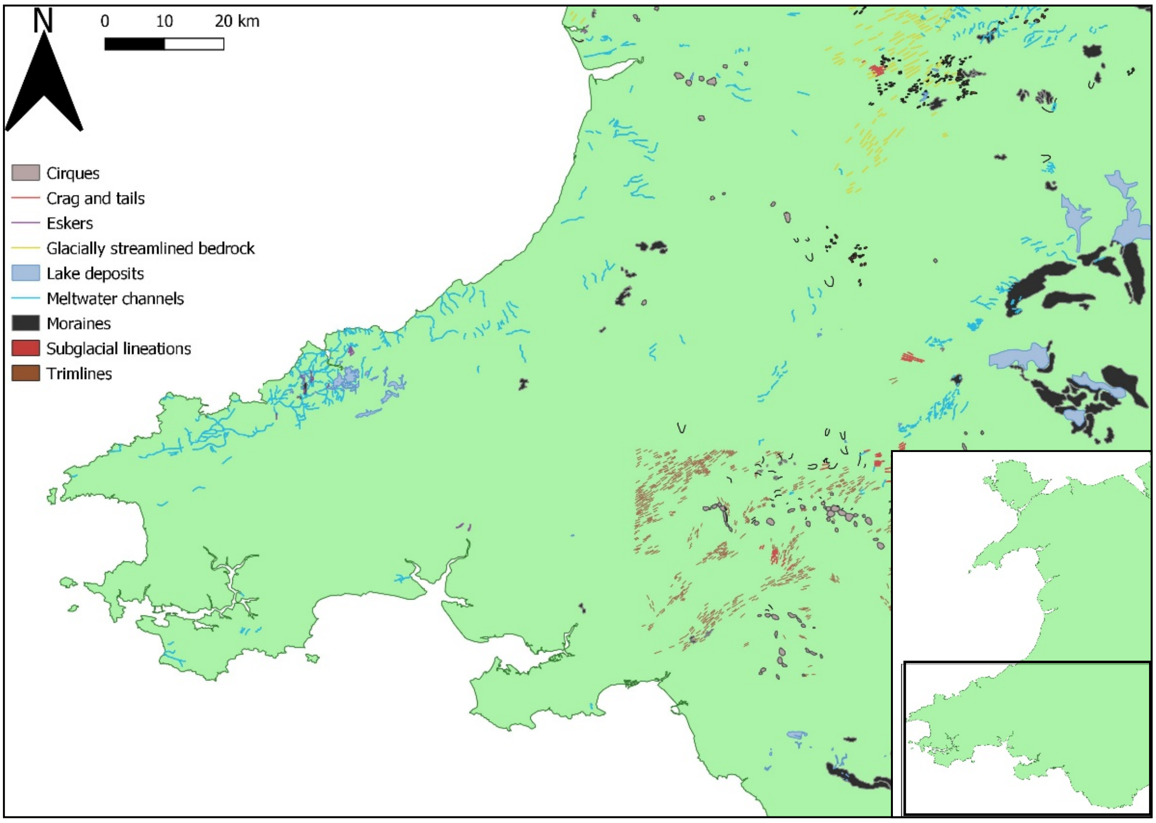

Figure 2.14: Map of south Wales showing glacial features from BRITICE (Data sources from Clarke *et al.,* 2018). © Crown copyright and database rights 2022 Ordnance Survey (100025252)

The chronology of glaciation in south Wales is unclear. There is evidence that ice reached the Welsh limit by 23000 BP with the retreat of the Welsh ice cap dated to between 25200-21200 BP (Clarke *et al.*, 2012; Philips *et al.,* 1994). At Broughton Bay there is a date of 17000 BP from a till overlying raised beach deposits, which provides a maximum date for the LGM (Evans *et al.,* 2005). The Irish Sea was ice free before 14000 BP and the Brecon Beacon ice cap melted before 13500 BP, but there are no dates constraining the deglaciation on the north Pembrokeshire coast (Hughes, 2009; Hughes *et al.,* 2010).

2.3 Sea-level change

There are many variables which influence sea-level changes. These include eustasy and isostasy as well as tectonic and local processes (Rovere *et al*., 2016; Murray-Wallace, 2013; Whitehouse and Bradley, 2013). Sea-level change is not uniform, it does not simply rise or fall as water is added or removed from the oceans. Eustatic sea-level change describes the global average variation in sea level surface caused by modifications in the volume and mass of water in the ocean basins (Rovere *et al.,* 2016; Murray-Wallace 2013; Whitehouse and Bradley, 2013). Many factors influence changes in eustatic sea level at different spatial and temporal scales, such as glacio-eustasy[19] (Murray-Wallace, 2013; Rovere *et al.,* 2016), plate tectonics (Garrett, 2013), changes in tidal regimes, sediment supply (Gehrels and Long, 2008) and thermo-steric[20] changes. Other factors also have an impact, such as the gravity and rotation of the earth, as well as meteorological changes in wind and ocean currents (Shennan *et al.,* 2012).

The primary driver affecting sea-level change since the Last Glacial Maximum (LGM) approximately 20,000 years ago was the melting of the ice sheets, which resulted in large amounts of meltwater entering the world's oceans, triggering a dramatic rise in eustatic sea level (Murray-Wallace, 2013; Whitehouse and Bradley, 2013). The build-up and subsequent melting of ice also had a significant effect on isostasy, particularly in areas formerly covered by ice sheets (Kemp *et al.,* 2013; Rovere *et al.,* 2016). Changes in mass loading led to the flexing of the crust as the mass of water (hydro-isostasy) or ice (glacio-isostasy) exerted force onto the solid surface of the Earth, causing subsidence beneath the ice sheet. This in turn caused uplift around the periphery of the ice sheet and formed a zone known as the forebulge (Gehrels and Long, 200; Kemp *et al.,* 2013; Rovere *et al.,* 2016). Upon

[19] ice sheet growth and decay
[20] thermal expansion of water

deglaciation the mass of ice was removed, causing isostatic uplift in the centre of the former ice sheet, while the forebulge collapsed, triggering subsidence around the periphery (Kemp *et al.,* 2013; Khan *et al.,* 2015; Peltier, 2004; Rovere *et al.,* 2016). Deglaciation following the LGM was not linear, with a number of meltwater pulses occurring, for example as a result of an ice-dammed lake being breached (Harrison, *et al.* 2019). These meltwater pulses led to dramatic increases in sea level over relatively short periods up to the mid-Holocene (Harrison, *et al.* 2019). Sea-level rise can also cause shorelines to move, for example a marine transgression, where shorelines move landward (Hein *et al.* 2014). Deglaciation and postglacial uplift or readjustment could result in a time-transgressive or diachronous shoreline, where marine transgression occurred at different time periods along the coast (Cattaneo and Steel, 2003).

As ice sheets grew, gravitational attraction drew sea water towards them, making the sea level appear to rise. Upon deglaciation, the gravitational force decreased leading to a fall in sea level close to the former ice sheets. In contrast, sea level rose in locations at a distance from glaciated areas i.e., far field locations (Khan *et al.,* 2015; Rovere *et al.,* 2016). In areas far from ice sheets the removal of water caused uplift in ocean basins (hydro-isostasy) and subsidence along the continental margins due to lithospheric flexure[21], known as continental levering (Milne and Mitrovica, 2008, Rovere *et al.,* 2016). Ocean water migrated towards the space created by the processes of continental levering and forebulge collapse, leading to a fall in sea level in equatorial locations far from former ice sheets. This is known as ocean siphoning (Gehrels and Long, 2008; Mitrovica and Milne, 2002; Mitrovica and Peltier, 1991). Mitrovica and Milne (2002) have estimated that ocean siphoning causes present day sea level to fall at a rate of approximately 0.3mm a^{-1} [22] in far field, low latitude locations (Gehrels, 2009, 2010; Mitrovica and Milne, 2002; Mitrovica and Peltier, 1991).

There are further factors affecting sea level in addition to the processes of isostasy and eustasy. Tectonic activity in ocean basins can cause dramatic changes, particularly at active plate margins

[21] process by which the lithosphere (rigid, thin outer layer of the Earth) bends under the action of forces such as glaciation
[22] The units mm a^{-1} indicate mm per year

(Rovere *et al.,* 2016). Local changes in sedimentation rates can also lead to an apparent rise or fall in sea level that does not correlate with the global changes in eustatic sea level (Shennan *et al.,* 2006). For example, if peat formation occurs at a faster rate than a slowly rising sea level, it appears that marine influence is decreasing.

Post-depositional factors, such as sediment compaction, can also have a dramatic impact on sea level, on a local and regional scale (Brain, 2016). Sediment volume is reduced through time due to the pressure of sediment above, or by the removal of water from the sediment matrix (Shennan *et al.,* 2006). In coastal wetlands, compaction results in a lowering of the surface elevation, therefore affecting records of sea-level change (Cahoon *et al*., 2002; Shennan *et al.,* 2009).

Sediment compaction is affected by factors such as the properties of the sediment, the loading history, and changes in the water content (Shennan *et al.,* 2009). Layers of intercalated peat and clay are more susceptible to compaction, and the rate is determined by the thickness of the overburden, the depth of sediment below and the thickness of the entire Holocene sequence (Shennan *et al.,* 2009). Saltmarsh sediments are also prone to compaction as the peat is waterlogged, highly organic and has low bulk densities (Brain, 2017). On the south coast of England, Holocene deposits are particularly susceptible to compaction as they are characterised by long sequences of unconsolidated sediments, particularly intercalated peat and clay (Edwards, 2006; Massey *et al.,* 2006; Waller and Long, 2003). However, peat at the base of a sequence which is resting on a hard substrate, such as gravel, will be resistant to compaction (Massey, *et al.,* 2006). There is currently no accepted method to correct for sediment compaction (Bird *et al.,* 2004; Brain *et al.,* 2012, 2015, 2016, 2017; Edwards, 2006; Massey *et al.,* 2006a, b; Pizzuto and Schwendt, 1997; Paul and Barras, 1998; Rybczyk, *et al.,* 1998; Tovey and Paul, 2002; Williams, 2003). It may be possible to make a quantitative assessment of compaction if detailed lithological and geotechnical models can accommodate the complex stratigraphy of coastal wetlands; however, this is often not possible (Allen *et al.,* 1999, 2000; Massey *et al.,* 2006a, b; Pizzuto and Schwedt, 1997; Paul and Barras, 1998;

Shennan *et al.,* 2006). An initial assessment of the impact of sediment compaction can be made by distinguishing between sea level data which originated from intercalated and basal samples (Shennan *et al.,* 2009).

2.4 Late Quaternary Sea-level Change in Wales

This section will outline the evidence for late Quaternary sea-level change in Wales. The review will discuss the evidence chronologically, from the north, west and south coast of Wales.

2.4.1 Post Late Glacial Maximum (18000 to 10500 cal. BP)

2.4.1.1 North Wales

Roberts *et al.* (2011) investigated three cores from the Menai Straits (SH479 644): two from the main channel (CJSC1 and CJSC2), and a third from intertidal mudflats near Bangor Pier (CJSC3) (Figure 2.15 and 16). Sea-level index points (SLIPs) were obtained from the contacts of transgressive and regressive episodes, spanning the period between the end of the LGM and the start of the early Holocene (18000 to 10500 cal. BP). Sedimentological, stratigraphic and low-resolution foraminifera analyses were used to establish indicative meaning: i.e. whether samples formed near the level of mean high water spring tides (MHWST, Chapter 3.4.1). Errors due to sediment compaction and changes in former tidal range were taken into account (Roberts *et al.,* 2011). Samples which did not have accurate indicative meanings were considered limiting points (Roberts *et al.,* 2011, Chapter 3.4.1).

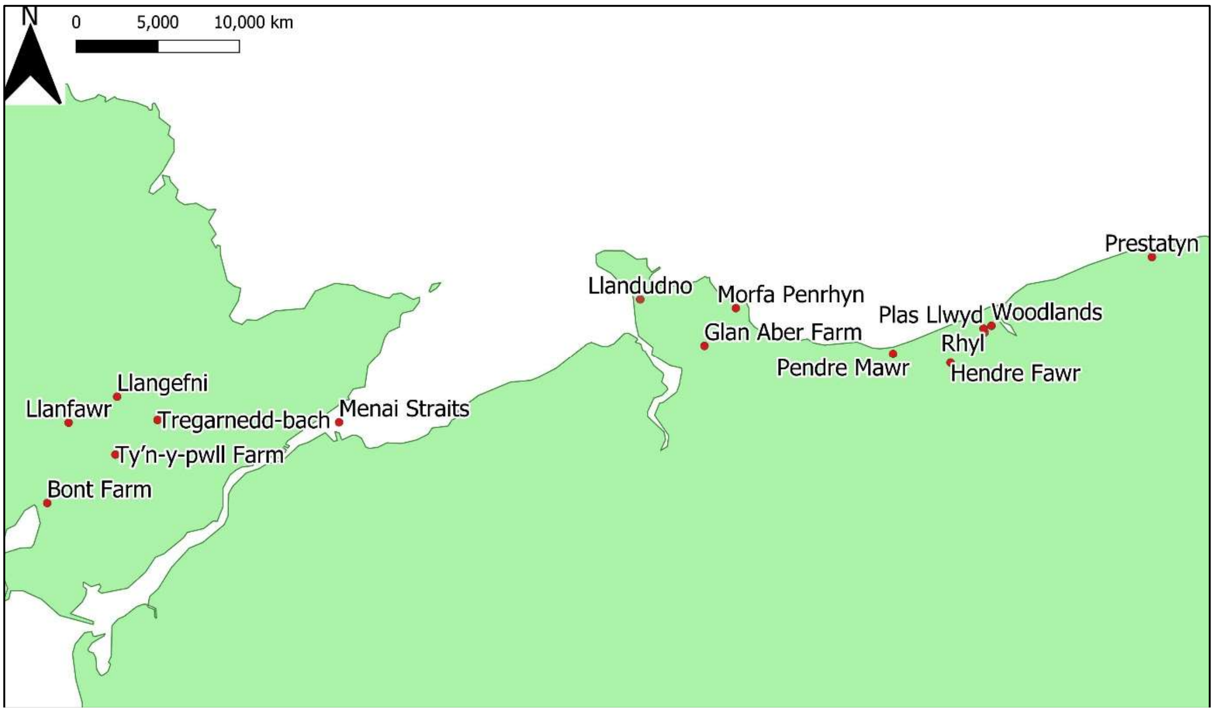

Figure 2.15: Sea-level sites for north Wales, indicating sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252).

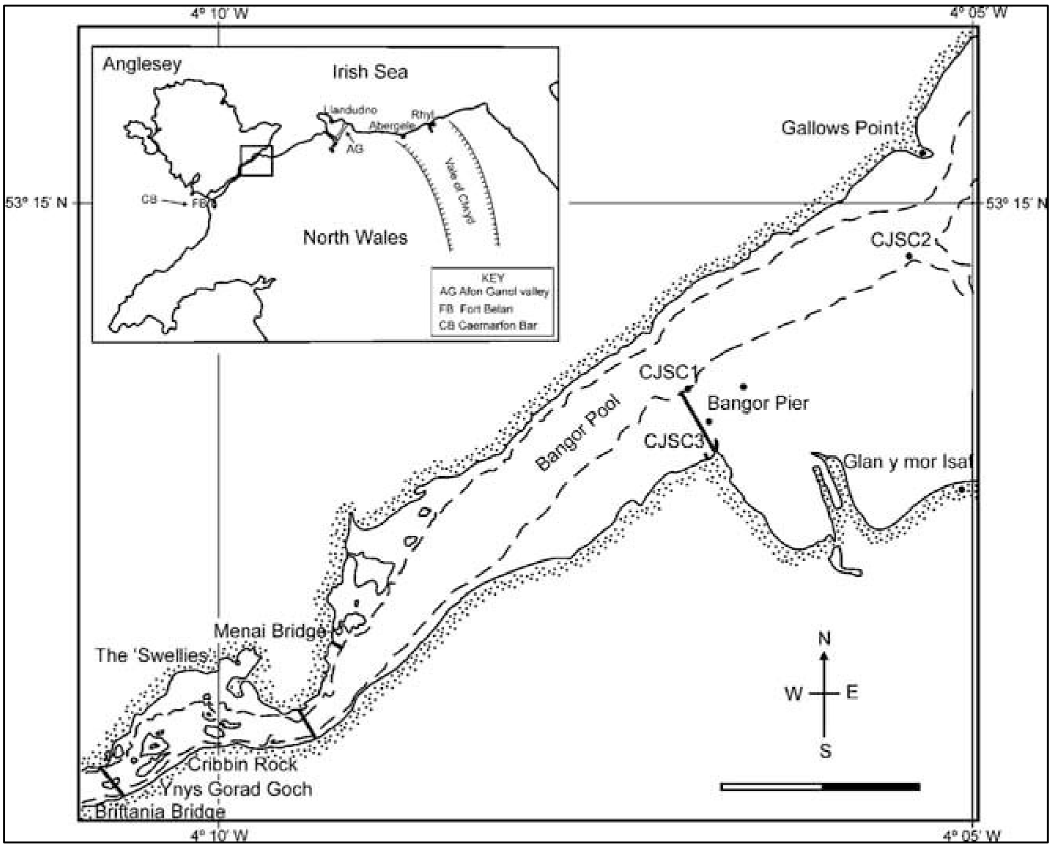

Figure 2.16: Map of the Menai Straits showing locations of coring sites mentioned in the text, scale bar = 1km (reproduced from Roberts *et al.,* 2011).

Core CJSC1 from the Menai Straits produced limiting points and SLIPs between the end of the LGM and the early Holocene (Table 2.1). Diamict was recovered from the base of the core between -35.42 to -27.02m OD, overlain by angular gravels up to -23.62m OD (Roberts *et al.,* 2011). Light grey silts and sands overlaid the gravel to -4.12m OD. The silts contained the marine molluscs of the genera *Mytilus, Cerastoderma, Tapes* and *Littorina*, with interbedded peats labelled 1a and 1b. Peat 1a incorporated two horizons of brown fibrous peat between -23.37 to -23.36m OD and -23.22 to -23.17m OD (Roberts *et al.,* 2011). Peat 1a produced four limiting points (SUERC2579, SUERC2578, SUERC2493, SUERC2496) between 11687-11246 and 11399-11099 cal. BP (Roberts *et al.,* 2011). As peat 1a was overlying gravel, the basal date was unlikely to have been affected by compaction. The second intercalated peat, 1b, was a dark brown, compact fibrous peat layer between -20.84 to -20.66m OD, which produced two SLIPs that indicated RSL rose from -24.44m OD to -24.08m OD between 11321-10825 cal. BP and 10652-10291 cal. BP (Roberts *et al.,* 2011) (Figure 2.17).

Table 2.1: Radiocarbon dates from Menai Straits core CJSC1 (Roberts *et al.,* 2011)

Site Name	Lab code	Calibrated Date (cal. BP)	Time	Altitude (m OD)	RSL (m)	Nature of contact	Index or Limiting point
Menai 1b (upper contact)	SUERC 2491	10652-10291	Early Holocene	-20.68	-24.08	Transgressive	Index Point
Menai 1b (lower contact)	SUERC 2492	11321-10825	Post LGM	-20.84	-24.44	Regressive	Index Point
Menai 1a (upper contact)	SUERC 2493	11607-11241	Post LGM	-23.08	-26.68	Transgressive	Limiting point
Menai 1a (lower contact)	SUERC 2579	11399-11099	Post LGM	-23.37	-26.97	Regressive	Limiting point
Menai 1a (upper contact)	SUERC 2578	11595-11123	Post LGM	-23.08	-26.68	Transgressive	Limiting point
Menai 1a (lower contact)	SUERC 2496	11687-11246	Post LGM	-23.37	-26.97	Regressive	Limiting point

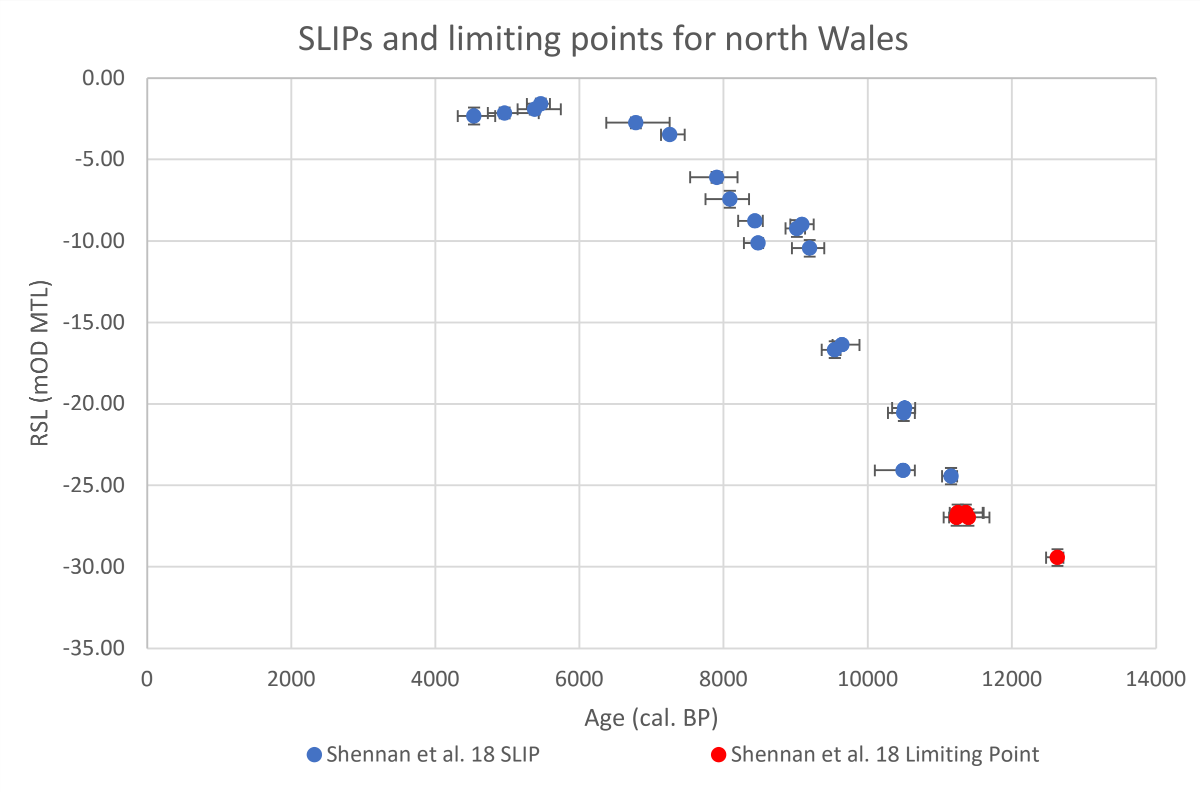

Figure 2.17: Sea-level curve for the north coast of Wales, with data from Shennan *et al.,* 2018 (blue: SLIP, red: limiting point).

Core CJSC2 also produced SLIPs from the period between the LGM and the early Holocene (Table 2.2). This core was dominated by fine, light-grey, and orange-brown silty clay. The clay contained sand and marine shells including the genus *Mytilus,* but no gravel or diamict were recovered (Roberts *et al.,* 2011). Five intercalated organic horizons were found within the silty clay (2a-e). An organic laminated clay (bed 2a) was dated to between 15700–14340 and 13820–13160 cal. BP (SUERC2508, -28.14m OD, SUERC2507, -28.21m OD), although these dates were rejected as they were contaminated by detrital coal (Roberts *et al.* 2011). A coarse, dark brown detrital mud at -25.83m OD (bed 2b) produced a limiting point of 12950–12360 cal. BP (SUERC 2506) from a transgressive boundary at -29.4m OD (Roberts *et al.,* 2011). Bed 2c, a detritus mud, produced two SLIPs from the start of the early Holocene (SUERC2503, SUERC2502) between 10654-10294 and 10655-10300 cal. BP when RSL was between -20.54 and -20.26m OD (Roberts *et al.,* 2011). An organic clay at -15.86 to -15.72m OD (bed 2d) produced two post LGM dates (SUERC2501, SUERC2500), however they were out of sequence compared to beds 2 a, b, c, and e, and so were rejected. The final organic horizon, 2e, consisted of composite horizons of detritus and mud, which produced early Holocene SLIPS (see section: early Holocene). The SLIPs from Menai cores CJSC 1 and 2 indicate that RSL rose from -26.97m OD at 12715-12544 cal. BP (SUERC 2506) to -20.54m OD at 10654-10294 (SUERC 2501), at a rate of 2.66mm a^{-1} (Roberts *et al.,* 2011).

Table 2.2**:** Radiocarbon dates from Menai Straits core CJSC2 (Roberts *et al.,* 2011)

Site Name	Lab code	Calibrated Date (cal. BP)	Time	Altitude (m OD)	RSL (m)	Nature of contact	Index or Limiting	Rejected?
Menai 2e (lower contact)	SUERC 2498	9622- 9489	Early Holocene	-13.07	-16.67	Regressive	Index Point	No
Menai 2e (upper contact)	SUERC 2497	9883- 9543	Early Holocene	-12.97	-16.37	Transgressive	Index Point	No
Menai 2d (lower contact)	SUERC 2501	12350-11660	Post LGM	-15.86	-	Regressive	Limiting point	Rejected: too old
Menai 2d (upper contact)	SUERC 2500	12110-11260	Post LGM	-15.72	-	Transgressive	Limiting point	Rejected: too old
Menai 2c (lower contact)	SUERC 2503	10654-10294	Early Holocene	-16.94	-20.54	Regressive	Index Point	No
Menai 2c (upper contact)	SUERC 2502	10655-10300	Post LGM	-16.86	-20.26	Transgressive	Index Point	No
Menai 2b (upper contact)	SUERC 2506	12715-12544	Post LGM	-25.83	-29.43	Transgressive	Limiting point	No
Menai 2a (upper contact)	SUERC 2508	13820-13160	Post LGM	-28.14	-	Transgressive	Limiting Point	Rejected: contaminated
Menai 2a (lower contact)	SUERC 2507	15700-14340	Post LGM	-28.21	-	Regressive	Limiting Point	Rejected: contaminated

Prince (1988) studied sites on the north coast of Wales including Anglesey (SH456 720) and the Clwyd Plain (SH846 786). Prince (1988) utilised multiple biostratigraphic proxies including pollen, diatoms, foraminifera and ostracods to reconstruct environmental change. Many radiocarbon dates were obtained; however the limiting points were rejected by later researchers due to lack of evidence for sea-level change at the dated horizons (Bedlington, 1994; Shennan *et al.,* 2018).

In the Vale of Clwyd, two cores at Plas Llwyd and Woodlands were extracted and produced pre-Holocene dates (Prince, 1988). At Plas Llwyd, basal freshwater gravel was overlain by blue grey clay between -10.58m OD to -6.67m OD. Two intercalated organic rich horizons were found between -10.25 to -9.59m OD and -9.59 to -9.08m OD (Prince, 1988). A date of 11192- 10725 cal. BP was obtained from the organic horizon at -9.54m OD. Pollen evidence indicated a coastal reed swamp environment with fresh-brackish diatoms (Prince, 1988). Microfossil evidence from the blue-grey silt, between -9.22 to -6.32m OD, indicated fresh-brackish standing water and a coastal reed swamp. A radiocarbon date was taken at -6.19m OD, where diatom analysis indicated marine-brackish conditions. However, the date of 13427- 13105 cal. BP was considered too old as the pollen evidence represented an open coastal environment typical of the early Holocene (Prince, 1988).

At the site of Woodlands nearby, a similar stratigraphy was recorded, with gravel overlain by blue-grey silt which was more organic than at Plas Llwyd, and grey-brown clay in the upper levels. Two samples of clay with *Phragmites* were submitted for radiocarbon dating. A date of 18985- 17575 cal. BP was taken of at -6.87m OD, where diatom evidence indicated a marine-brackish high saltmarsh environment. The date was considered too old, however, as the pollen assemblage showed a mixed oak-hazel woodland with a nearby saltmarsh, which is not typical of a late-glacial sequence (Prince, 1988).

Prince (1988) also investigated several sites in Anglesey including Bont Farm, which produced two pre-Holocene dates. The till surface at -47.97m OD was overlain by gravel and then clay from -46.53 to -23.81m OD. Organic horizons were recovered between -23.81 to -23.55m OD, which were dated to 23705- 22486 cal. BP (SRR2644). Sands and shells dominated the stratigraphy from -23.55 to -11.14m OD, with a piece of wood at -11.14m OD dated to 42236-39064 cal. BP (SRR2505) (Prince, 1988). These dates were rejected as they were inverted, and no detailed biostratigraphic analysis was undertaken on the core (Prince 1988).

2.4.1.2 West Wales

The only pre-Holocene sea level data for west Wales was produced by Prince (1988) at Harlech (SH 582 311), Llanbedr (SH 585 268) and Ynyslas (SN 606 928). However, the data was rejected by later researchers due to no evidence for sea-level change (Bedlington, 1994; Shennan *et al.,* 2018).

At Harlech, dark-grey silty clay was recovered from -47.55m OD, overlain by blue-black organic clay to -41.48m OD, which contained the marine shell *Turritella communis* (Prince, 1988). Four radiocarbon dates were obtained between -47.55 and -45.29m OD (SRR2501, SRR2640, SRR2641, SRR2502). The date SRR2501 (13278-12803 cal. BP) was considered too old as it was out of sequence, and the ostracod and foraminifera assemblages did not reflect a pre-Holocene environment (Prince, 1988). Between -39.05 to -37.52m OD, blue grey clay was recovered, overlain by grey, silty clay to -27.46m OD. The upper layers were dominated by sand containing shell fragments and seaweed (Prince, 1988). Overall, the microfossil evidence indicated a deep channel deposit throughout the core which formed in response to rapid sea-level rise, with some local changes in dryness and salinity (Prince, 1988).

Nearby at Llanbedr a similar sequence was recovered, with late-glacial, freshwater gravel overlain by clay containing marine and brackish foraminifera and well-preserved marine molluscs (Prince, 1988). Above this, alternating sequences of clay and gravel within silt contained evidence for continued sea-level influence and freshwater flooding, evidenced by saltmarsh pollen and low marsh foraminifera. Between -6.34 to -6.19m OD a peaty clay layer was found containing macrofossil remains of *Phragmites*, *Carex* and *Selaginella.* A date of 14744-13757 cal. BP (SRR2512) was obtained at -6.19m OD, with a late-glacial pollen assemblage and microfossil evidence for freshwater pools. However, Prince (1988) rejected the date as it was considered too old.

At Ynyslas, a core was taken at Ynys Tachwedd which produced a similar sequence of basal gravel overlain by organic clay, then sand and shells (Prince 1988). A date of 11750-11103 cal. BP (SRR 2504) was obtained from organic clay between -28.62 to -28.01m OD. The organic clay contained low marsh foraminifera and reed swamp pollen (Prince, 1988). Between -24.40 to -23.98m OD a date of 11238-10811 cal. BP (SRR2503) was obtained from organic clay with evidence provided by willow and hazel pollen, and a decline in marine diatoms indicating an increase in terrestrial, drier conditions (Prince, 1988).

2.4.1.3 South Wales

There are three limiting points from the south coast of Wales between the LGM and the start of the early Holocene (Heyworth and Kidson, 1982) (Figure 2.18, Table 2.3). These limiting points show sea level rose from -34.86m OD to -23.76m OD between 14279-13294 cal. BP (Q664) and 12116-10862 cal. BP, at rate of 3.25mm a^{-1} (Heyworth and Kidson, 1982). Care must be taken, however, as no details about stratigraphy or indicative meaning were published, and so the relationship to sea-level is unclear.

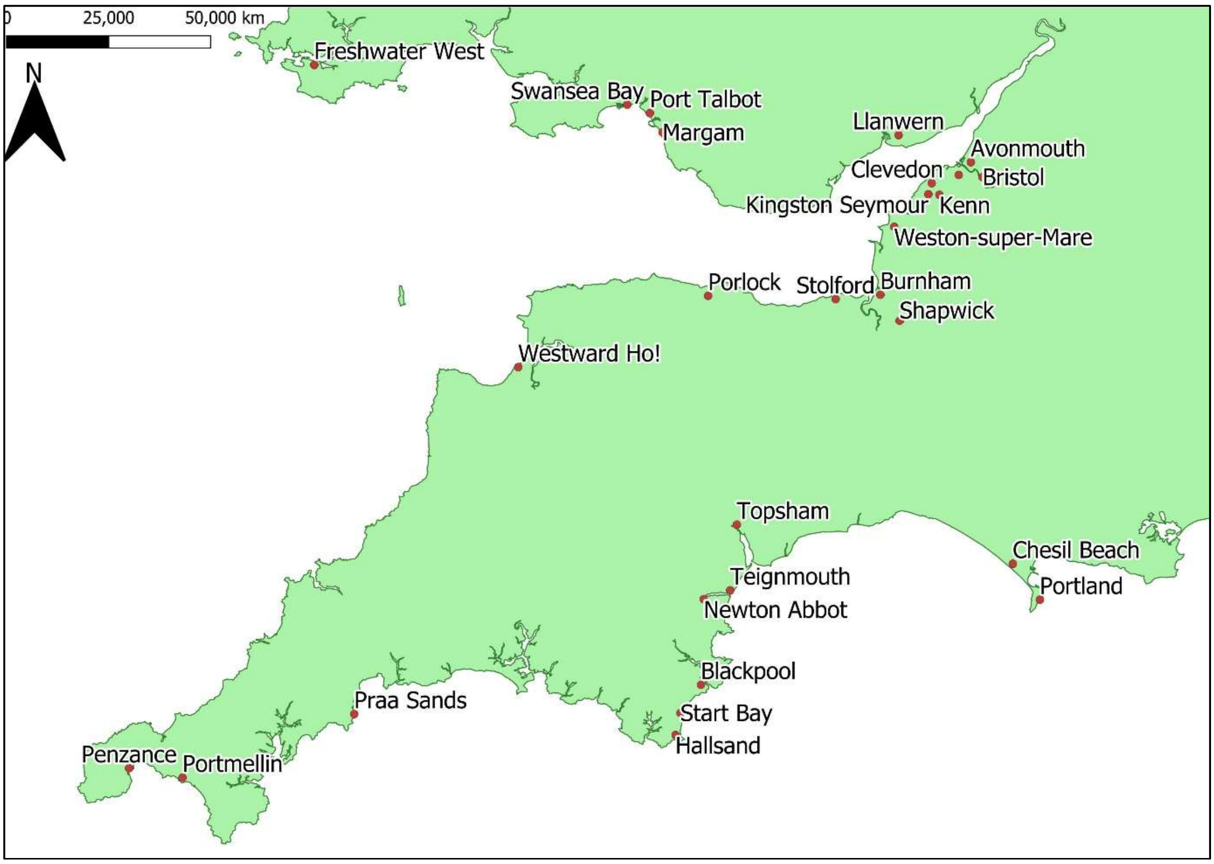

Figure 2.18: Sea-level sites for south Wales, data from Heyworth and Kidston (1982). © Crown copyright and database rights 2022 Ordnance Survey (100025252)

Table 2.3: Radiocarbon dates from Heyworth and Kidson (1982) between the LGM and the early Holocene, from south Wales

Reference	Lab code	Calibrated Date (cal. BP)	Time	Altitude (m OD)	RSL (m)	Nature of contact/ material dated	Index or limiting point	Rejected?
Heyworth and Kidson 1982	Q661	12116-10862	Post LGM	-18.90	-23.76	Unknown	Limiting Point	No
Heyworth and Kidson 1982	Q660	12901-11646	Post LGM	-18.95	-23.81	Unknown	Limiting Point	No
Heyworth and Kidson 1982	Q664	14279-10862	Post LGM	-30.00	-34.86	Unknown	Limiting Point	No

2.4.1.4 Summary: post Late Glacial Maximum

There is very little sea-level data from Wales between the end of the LGM and the onset of the early Holocene. Prince (1988) investigated several sites on the north and west coast of Wales and reconstructed environmental change, which included many pre-Holocene dates. However, the data was not included in later syntheses of sea-level change in North Wales (Bedlington, 1994) and the UK (Shennan and Horton, 2002; Shennan *et al.,* 2018) as there was no accurate biostratigraphic indicative meaning.

Limiting points and SLIPs from the Menai straits on the north coast indicate RSL rose at a rate of 2.66mm a^{-1} between 12715-12544 cal. BP (SUERC 2506) and 10654-10294 cal. BP (SUERC 2501) (Roberts *et al.,* 2011). Three limiting points from the south coast imply a possible rate of sea-level rise of 3.25mm a^{-1} between 14279-13294 cal. BP (Q664) and 12116-10862 cal. BP (Heyworth and Kidson, 1982). Care must be taken, however, as there is very little data from the south coast and no details of indicative meaning or stratigraphy were published (Heyworth and Kidson, 1982). The evidence from the Menai Straits and the south coast of Wales would suggest a very rapid rate of sea-level rise around the Welsh coast after the LGM (Roberts *et al.,* 2011, Heyworth and Kidson, 1982). This reflects the rapid rate of global sea-level rise due to the influx of meltwater from deglaciation (Lambeck *et al.,* 2014). It is possible that the rate of sea-level rise was higher on the south coast due to differential isostatic uplift, however much more data is needed to confirm this pattern.

2.4.2 Early Holocene (10500 to 7000 cal. BP)

2.4.2.1 North Wales

For north Wales, there is evidence for early Holocene sea-level rise in the Menai straits from core CJSC3 as well as the upper sequences of cores CJSC1 and 2 (Table 2.4, Roberts *et al.,* 2011). Core CJSC3 was located in intertidal mudflats at the landward end of Bangor pier (SH5845 7335) (Roberts *et al.,* 2011). Core CJSC3 consisted of partially laminated silty clay with interbedded organics (peat beds 3a and 3b) overlain by silty sand and clay (Roberts *et al.,* 2011). Peat 3a contained foraminifera indicating upper marsh conditions and was dated to between 9394-9031 and 8548-8407 cal. BP. (-7.89 to -7.77mOD, SUERC2511 and SUERC2512). Peat 3b contained fresh/brackish aquatic macrofossils and was dated to between 9250-9008 and 9129-8780 cal. BP. (-6.68 to -6.63mOD, SUERC2509 and SUERC2510).

Core CJSC2 was dominated by silty clays, with five interbedded organic horizons (beds 2a-e, Chapter 2.4.1). Bed 2c, a detritus mud, produced two SLIPs from the start of the early Holocene (SUERC2503, SUERC2502) between 10654-10294 and 10655-10300 cal. BP. (Roberts *et al.,* 2011). The detritus mud contained the brackish foraminifera *Trochammina inflata* and macrofossils of freshwater/brackish aquatic taxa *Ruppia maritima, Zannichellia palustris* and *Chara oospore* (Roberts *et al.,* 2011). The organic deposit was overlain by clay which contained abundant ostracods that indicated intertidal brackish conditions (Roberts *et al.,* 2011). The final organic layer, 2e, consisted of composite horizons of brackish detritus and mud, which produced two early Holocene SLIPS of 9874-9543 cal. BP. (SUERC2497, -12.97m OD) and 9622-9489 cal. BP. (SUERC2498, -13.07m OD).

The early Holocene SLIPs from all three cores in the Menai straits indicate sea-level rose at a rate of 6.72mm a^{-1} between 10654 to 8407 cal. BP. (Roberts *et al.,* 2011). Roberts *et al.,* (2011) determined that the tidal causeway in the Menai straits formed between 8600 and 8200 cal. BP. if the palaeotidal range was considered (Uehara *et al.,* 2006). This indicates that during the Mesolithic there would have been a land bridge between Anglesey and the Welsh mainland (Roberts *et al.,* 2011).

Table 2.4: Radiocarbon dates from the Menai Straits cores CJSC2 and 3 for the early Holocene (Roberts *et al.,* 2011)

Site Name	Lab code	Calibrated Date (cal. BP)	Nature of contact	Altitude (m OD)	RSL (m)	Index or limiting point
Menai 3a (upper contact)	SUERC2511	8548-8407	Transgressive	-6.72	-10.12	Index
Menai 3a (lower contact)	SUERC2512	9394-9031	Regressive	-6.84	-10.44	Index
Menai 3b (upper contact)	SUERC2509	9250-9008	Transgressive	-5.58	-8.98	Index
Menai 3b (lower contact)	SUERC2510	9129-8780	Regressive	-5.63	-9.23	Index
Menai 2e (upper contact)	SUERC2497	9874-9543	Transgressive	-12.97	-16.37	Index
Menai 2e (lower contact)	SUERC2498	9622-9489	Regressive	-13.07	-16.67	Index
Menai 2c (upper contact)	SUERC 2502	10655-10300	Transgressive	-16.86	-20.26	Index
Menai 2c (lower contact)	SUERC2503	10654-10294	Regressive	-16.94	-20.54	Index
Menai 1b (upper contact)	SUERC2491	10652-10291	Transgressive	-20.68	-24.08	Index

Prince (1988) produced three early Holocene limiting points from the site at Woodlands on the Clwyd Plains. Basal gravel was overlain by blue-grey silt and grey-brown clay (Chapter 2.4.1). A woody detrital peat was dated to 9400-9000 and 9690-9330cal. BP. between -12.95 and -13.07m OD (SRR 2510 and SRR 2511) (Prince, 1988). Pollen from the peat indicated an open habitat with *Cyperaceae, Poaceae* and *Chenopodiaceae*, which transitioned to an arboreal landscape with a saltmarsh pollen assemblage in the blue-grey clay above (Prince, 1988). The upper date at -2.71 to -2.64m OD (SRR 2642, Table 2.5) contained upper brackish diatoms with pollen evidence for a coastal reed swamp, however there was evidence for reworking and so the date was rejected (Prince, 1988).

Prince (1988) also studied the site of Ty'n-y-pwll Farm at the Malltreath marshes on Anglesey and found a similar stratigraphy to Bont Farm (Chapter 2.4.1). The till surface was found at -11.94m OD overlain by sand and gravelly clay to -9.35m OD. A *Typha* layer between -9.35 to -9.50m OD produced two limiting points between 8378-8189 and 8979-8380 cal. BP. from *Typha* (SRR 2506) and shell fragments (SRR 2507). Biostratigraphical analysis was undertaken on the dated horizon and revealed low concentrations of pollen dominated by Cyperaceae as well as marine and brackish diatoms (Prince, 1988). The *Typha* layer was overlain by sands, clays and shells including a bed of marine *Cerastoderma* shells at -8.95m OD (Prince, 1988).

Table 2.5: Radiocarbon dates for the early Holocene in north Wales (Prince, 1988; Bedlington, 1994; Heyworth and Kidson, 1982)

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Nature of contact/ material dated	Altitude (mOD)	RSL (m)	Index or limiting point	Rejected?
Vale of Clwyd Woodlands	Prince 1988	SRR 2642	9549-9327	Clay with *Phragmites*	-2.71 to -2.64	-	Limiting	Rejected: reworked
Vale of Clwyd Woodlands	Prince 1988	SRR 2510	9400-9000	Peat Regressive	-12.95	-	Limiting	Rejected: no evidence of sea level
Vale of Clwyd Woodlands	Prince 1988	SRR 2511	9690-9330	Peat Regressive	-13.07	-	Limiting	Rejected: no evidence of sea level
Anglesey Ty'n-y-pwll Farm	Prince 1988	SRR 2506	8378-8189	*Typha* layer	-9.35	-	Limiting	Rejected: no evidence of sea level
Anglesey Ty'n-y-pwll Farm	Prince 1988	SRR 2507	8979-8380	Shell fragments in *Typha* layer	-9.35	-	Limiting	Rejected: no evidence of sea level
Clwyd Plain Hendre Fawr (4)	Bedlington 1994	HV 17814	8192-7609	Basal peat Transgressive	-2.48	-6.09	Limiting	
Malltraeth Tregarnedd Bach (3)	Bedlington 1994	HV 17818	8600-7840	Transgressive	-3.96	-	Limiting	Rejected: Inverted
Malltraeth Tregarnedd Bach (2)	Bedlington 1994	HV 17819	8352-7839	Basal peat overlaying gravel Transgressive	-4.11	-7.44	Limiting	
Llandudno Station	Heyworth and Kidson 1982; Radiocarbon, v16/2	SRR 61	8542-8369	Unknown	-5.21 to -4.16	-8.76	Limiting	

A transect at the site of Tregarnedd Bach in the Malltreath marshes revealed a similar sequence to nearby sites such as Bont Farm and Tyn-y-pwll farm (Bedlington, 1994; Prince, 1988). A basal peat was found between -4.42 and -3.15m OD, overlain by silty clay between -3.15 and +1.00m OD (Bedlington, 1994). In the centre of the valley, sands and silts were recovered between -2.0 and 0m

OD and a humified organic horizon was found between +1 and +1.5mOD. Overlying the whole site was 1.5m of fresh *turfa* (Bedlington, 1994). No diatoms were found, and pollen concentrations were very low, but from the lithostratigraphy Bedlington (1994) inferred marine conditions with terrestrial phases. The radiocarbon dates from Tregarnedd Bach were inverted, but the pollen evidence for mixed-oak woodland indicated that the basal date of 8352-7839 cal. BP at -4.11m OD was correct (Bedlington, 1994). The sites from the Malltreath marshes (Bont Farm, Tyn-y-pwll farm and Tregarnedd Bach) indicate that the area was dominated by marsh and tidal flats throughout the Holocene with marginal areas, such as Tregarnedd Bach, more sensitive to sea-level fluctuations (Bedlington, 1994; Prince, 1998).

At Hendre Fawr on the Clwyd Plain, Bedlington (1994) found a similar sequence to the nearby sites of Woodlands and Plas Llwyd (Prince, 1998), with a basal peat at -2.48m OD dated to 8192-7609 cal. BP. Diatoms in the overlying clay indicated a transition from a terrestrial environment through an upper marsh to lower marsh conditions, indicating an increase in marine influence (Bedlington, 1994). This is similar to the sequence found at Llandudno station, which contains a basal peat overlain by estuarine clay, followed by beach deposits and blown sand (Harkness and Wilson, 1974). At Llandudno station the peat was dated to 8542-8369 cal. BP between -5.21 to -4.16m OD, but due to the wide altitudinal range, and lack of paleoenvironmental context, it was rejected by Roberts *et al.* (2011).

2.4.2.2 West Wales

The only data from mid Wales during the early Holocene is from a site called Trawling End or Aberayron, which is located 5km offshore between Aberystwyth and Newquay (Figure 2.19, Table 2.6; Wilks, 1977; Heyworth and Kidson, 1982; Kitely, 1975). Kitely (1975) undertook foraminifera analysis, while Wilks (1977) analysed diatoms and pollen. The basal till was overlain by banded clays containing wood fragments and no foraminifera, indicating a terrestrial, freshwater environment, interpreted as the post-glacial period (Wilks, 1977). Overlaying this terrestrial clay, a thin band of peri-marine clay was recovered, with pollen evidence for local hazel-willow carr and regional mixed oak-birch woodland. The clay was overlain by peat which contained increasing quantities of saltmarsh pollen implying tidal submersion (Wilks, 1977). The peat was dated to 10154-9537 cal. BP at -20.50m OD, however the date was rejected as the altitude was estimated using echo soundings, resulting in an error of ±1.5m (Wilks, 1977; Kitely, 1975). Diatoms from the dated altitude indicate an upper-brackish water environment, with *Diploneis elliptica*, *Diploneis interrupta*, *Navicula peregrina* and *Navicula mutica* (Wilks, 1977). Wilks (1977) compared the diatom assemblage to modern samples, and found they reflected the saltmarsh in the area between Borth Bog and the Dovey Estuary (Wilks, 1977).

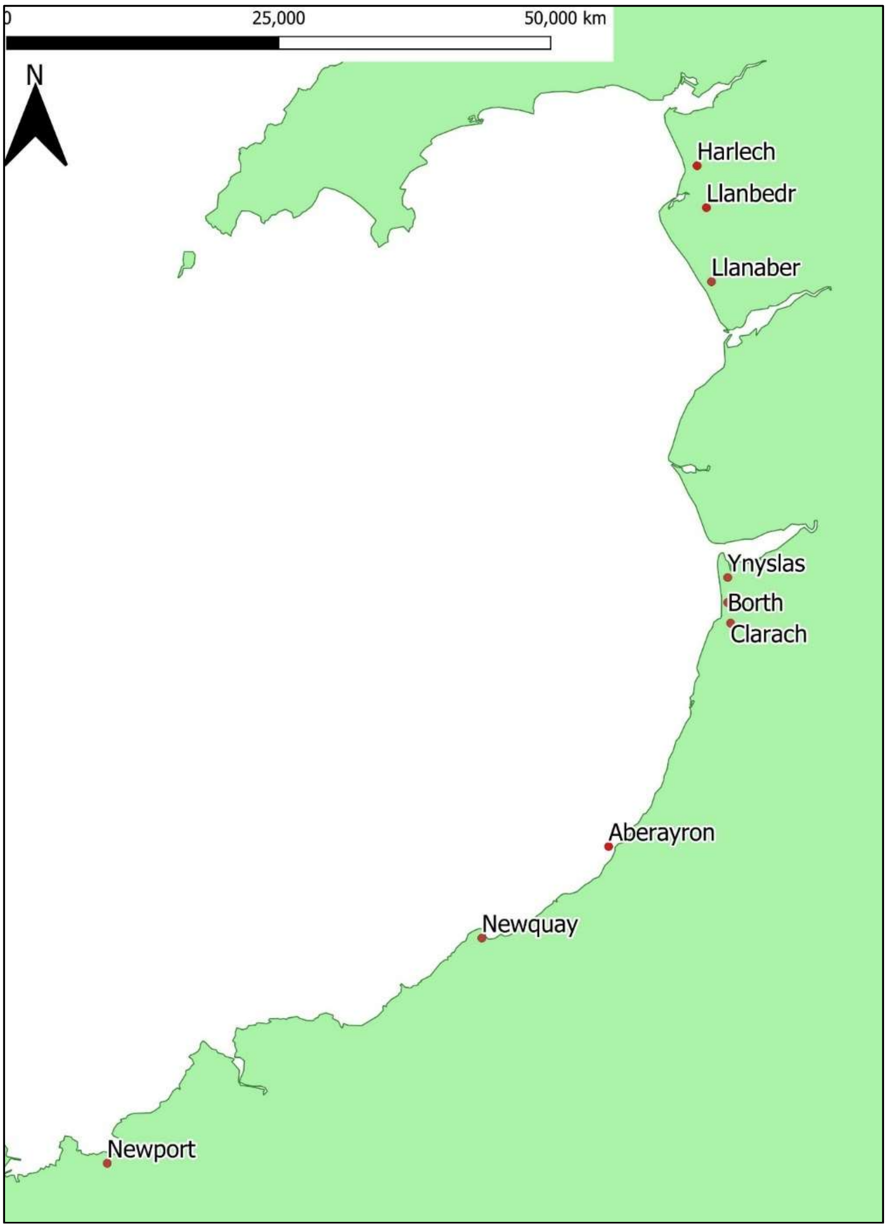

Figure 2.19: Sea-level sites for west Wales indicating sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252).

Overlaying the peat, a laminated silty clay was recovered containing brackish foraminifera indicating estuarine conditions (Kitely, 1975). Very little pollen was recovered, but diatoms showed increasing salinity from the peat horizon into the silty clay, with increasing marine and strongly brackish forms (Wilks, 1977). The proportion of marine foraminifera increased through the clay (Kitely, 1975), similar to the sequence found at Borth (Haynes and Dobson, 1969). At the top of the clay there was a thin shell layer containing estuarine and open marine foraminifera, indicating an estuary mouth environment (Kitely, 1975; Wilks, 1977). Above this, muddy sands with marine foraminifera and a shallow marine environment overlain by fluid muds with an open, diverse marine foraminifera assemblage was recorded (Kitley, 1975; Wilks, 1977). The results indicate a steadily drowning post glacial environment throughout the rapid sea-level rise of the early Holocene (Kitely, 1975; Wilks, 1977).

Table 2.6: Radiocarbon date from Aberayron for the early Holocene in west Wales (Heyworth and Kidson, 1982; Wilks, 1977; from Kitely, 1975)

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Nature of contact/ material dated	Altitude (m oD)	RSL (m)	Index or limiting point	Rejected?
Trawling End aka Aberayron aka Newquay, Dyffedd	Heyworth and Kidson 1982; Wilks 1977; from Kitely 1975	BIRM400	10154-9537	Basal peat	-20.50	-22.16	Limiting	Rejected: Altitude error: ±1.5m

2.4.2.3 South Wales

In south Wales, sea-level studies have been focussed on the macrotidal Severn Estuary, which has one of the largest tidal ranges in the world at 14.65m (Proudman Oceanographic Lab, 2018; Hill *et al.,* 2007). Deep sequences of estuarine silts and brackish/freshwater peats have been found along the edges of the estuary which developed in response to postglacial sea level (Allen and Rae, 1987; Culver and Banner, 1978; Devoy, 1979; Evans and Thomspon, 1979; Hawkins, 1971; Heyworth and Kidson, 1982; Shennan, 1983). There is a distinct tripartite sedimentary sequence known as the Wentlooge Formation which has been recognised across the Severn Estuary, particularly in the Gwent levels (Allen and Rae, 1976; Sturt *et al.,* 2014). The sedimentary sequence in the Bridgwater Bay area, however, differs from the Wentlooge formation, making direct comparisons challenging (Sturt *et al.,* 2014). There are many intercalated peats on the margins of Bridgwater Bay, for example at Burnham on Sea (Druce, 2000, Chapter 2.4.3) and Stolford (Heyworth and Kidson, 1982; Kidson and Heyworth, 1976; Heyworth, 1985; Hillam *et al.,* 1990).

There are a series of limiting points and SLIPs published in Heyworth and Kidson (1982) for the south coast of the Bristol Channel, including the site at Stolford (Table 2.7). These data appear to show that between 10511 to 7520 cal. BP sea level rose at a rate of 5.3mm a^{-1} (Heyworth and Kidson, 1982) (Figure 2.20). However, care must be taken as no details of stratigraphy were published, and the relationship to sea level is unclear.

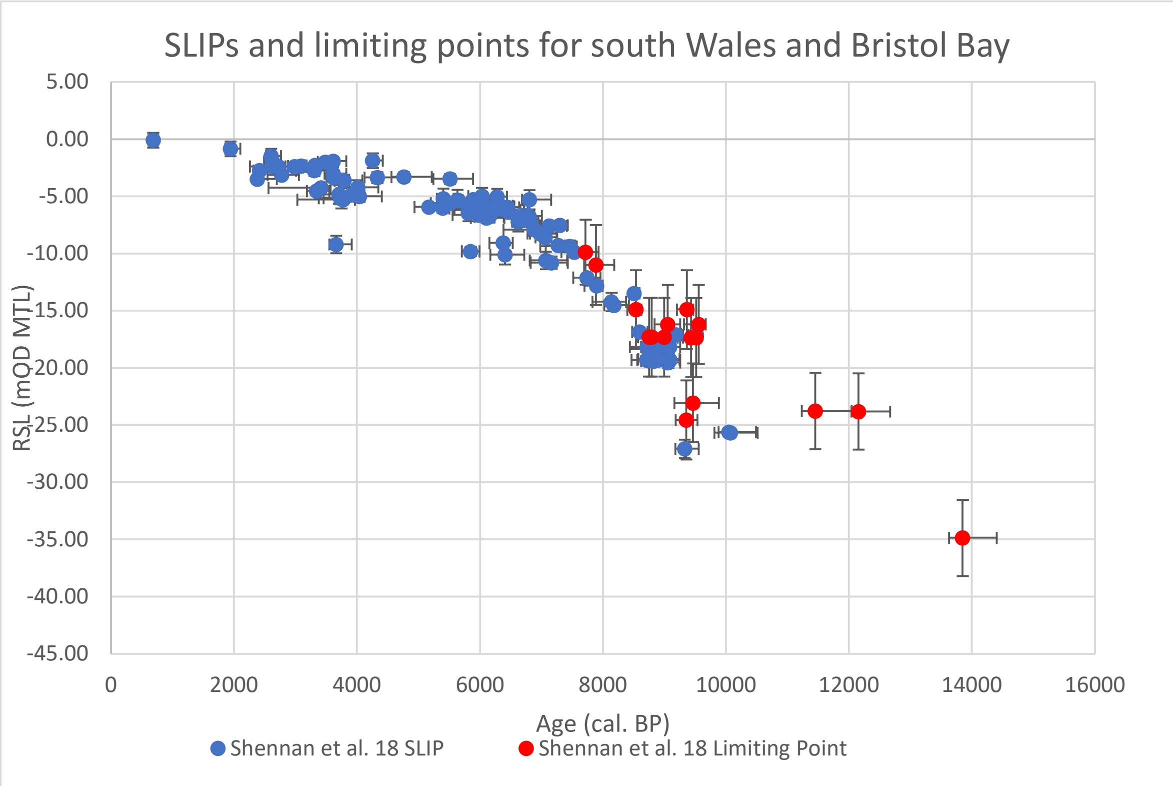

Figure 2.20: Sea-level curve for south Wales and Bristol Bay, with data from Shennan *et al.,* 2018 (blue: SLIP, red: limiting point).

Table 2.7: Radiocarbon dates from Heyworth and Kidson (1982) for the early Holocene in South Wales

Site Name	Lab code	Calibrated Date (cal. BP)	Nature of contact/ material dated	Altitude (m OD)	RSL (m)	Index or limiting point
Stolford	I2688	8182-7593	Basal peat	-7.30	-11.00	Limiting
Burnham	I3713	8374-7943	Unknown	-8.45	-14.22	Index
-	IGS53	9535-9092	Basal peat	-21.00	-24.55	Limiting
Highbridge	I4402	9885-9034	Basal peat	-19.50	-23.05	Limiting
Highbridge	I4403	9558-9001	Unknown	-21.30	-27.07	Index
Wall Common	I2689	7956-7520	Unknown	-5.92	-12.10	Index
Wall Common	I2690	8419-7933	Unknown	-8.95	-14.52	Index
Baglan Burrows, Port Talbot	Q663	10489-9564	Unknown	-18.80	-25.60	Index
Baglan Burrows, Port Talbot	Q662	10511-9564	Unknown	-18.86	-25.66	Index

Further west along the Somerset coast, intercalated peats and submerged forests occurs, for example at Minehead (Jones *et al.,* 2005) and Porlock (Jennings *et al.,* 1998; Sturt *et al.,* 2014). Jennings *et al.* (1998) assessed the Porlock marsh area, from 71 cores with 17 radiocarbon dates as well as pollen and diatom analyses, and produced 6 SLIPs (Table 2.8, Chapter 2.4.3). The stratigraphy consisted of Holocene alluvial deposits and intercalated peats, overlying Pleistocene glacial deposits (Jennings *et al.,* 1998; Lawrence, 2010). The results showed a gradual change from a freshwater environment with an increase in first brackish, then marine taxa, and a return to freshwater conditions at the top of the profile due to anthropogenic flooding (Jennings *et al.,* 1998; Lawrence, 2010). The SLIPs indicate that during the early Holocene, sea level rose at a rate of 3.38mm a^{-1} at Porlock Marsh (Jennings *et al*., 1998).

Table 2.8: Radiocarbon dates for Porlock Marsh for the early Holocene (Jennings *et al,.* 1998).

Site Name	Lab code	Calibrated Date (cal. BP)	Nature of contact	Altitude (m OD)	RSL (m)	Index or limiting point	Notes
Porlock Forest Bed	Beta 61544	7891-7490	-	-6.67	-9.86	Limiting	Base of basal peat
Porlock Forest Bed	Beta 86775	7937-7727	Transgressive	-8.00	-12.80	Index	
Porlock Forest Bed	Beta 81655	8559-8375	Transgressive	-8.67	-13.47	Index	

A series of offshore boreholes were investigated at Hinkley Point on the north Somerset coast, located at the west end of Bridgwater Bay in the inner Bristol Channel (Figure 2.21) (Sturt *et al*., 2014; Griffiths *et al.,* 2015). Data from offshore geophysical and geotechnical surveys were analysed to provide context for the 23 boreholes and 62 vibrocores obtained, and swath bathymetry was used to establish the altitude (Sturt *et al.,* 2014). A subsample of six cores were used for analyses, which included pollen, diatom, foraminifera, plant macrofossils and molluscs (Sturt *et al*., 2014).

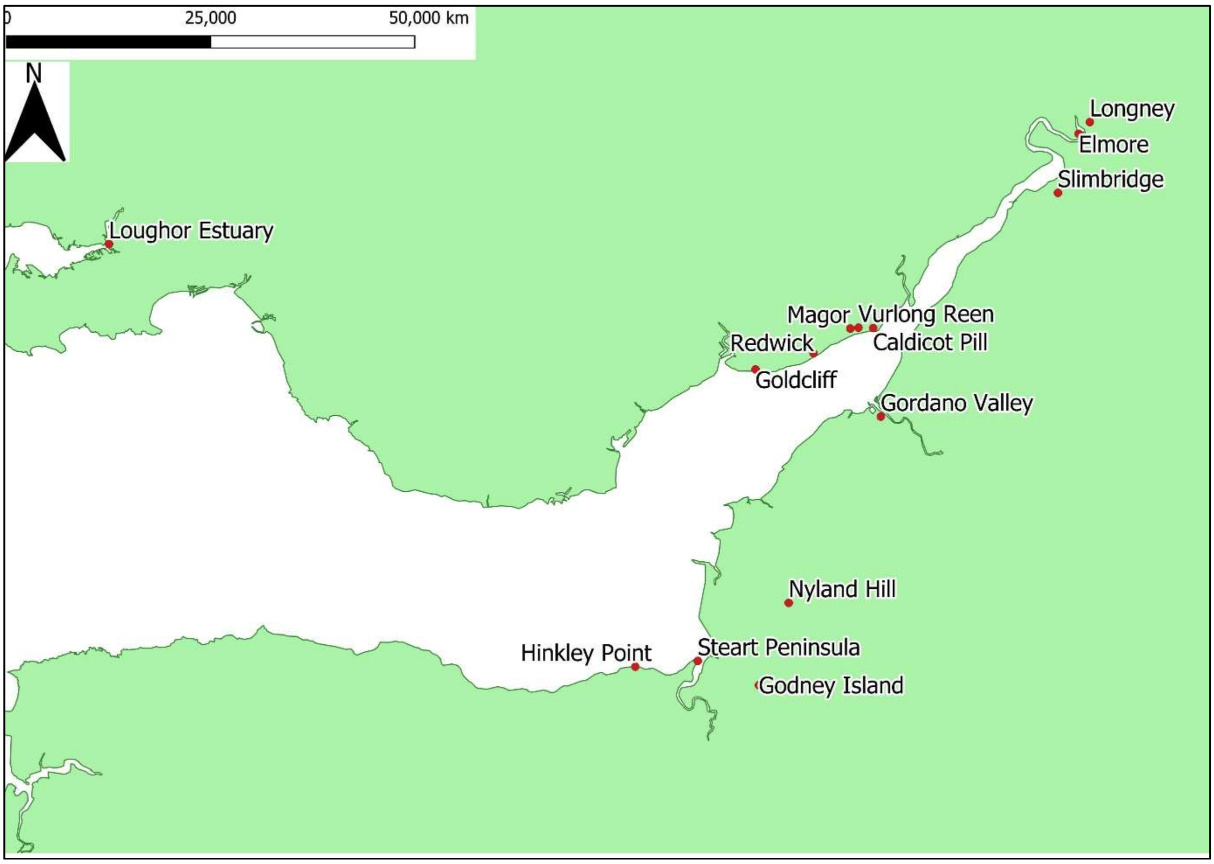

Figure 2.21: Sea-level sites for south Wales, including sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252). For sites mention in Heyworth and Kidston (1982) see Figure 2.18.

The geophysical and geotechnical survey revealed a widespread, but fragmentary, buried early Holocene landscape. An open, meandering shallow channel was revealed in the Severn Estuary, broadening into a steep sided channel under Bridgwater Bay, before becoming buried under sand and gravel in the Bristol channel (Sturt *et al*., 2014). The results of the borehole survey revealed a uniform stratigraphy, with the bedrock overlain by gravel and silty clay, which had low sediment $^{\delta 13}$C values and C/N ratios indicating a terrestrial, freshwater environment (Sturt *et al*., 2014). There was evidence for reworked clastic sediments and differential preservation, although occasional pollen grains were found including *Pinus, Corylus* and fern spores which were typical of the early Holocene vegetation of the area (Sturt *et al*., 2014).

Overlying the silts was a basal peat between-13.74m OD to -12.58m OD, with an organic silt at -11.24m OD in core VCJ15. Sturt *et al.,* (2014) determined that the peat formation was driven by sea level rise which forced the ground water upwards, leading to organic sedimentation. Multiple dates were taken from three different fractions where available: humic and humin from bulk peat, and identifiable, short-lived macrofossils (Sturt *et al*., 2014; Griffith *et al*., 2015). Where these dates were statistically comparable, a weighted mean was calculated. In core VC15 peat formation at -13.80m OD produced a weighted mean age of 9530-9430 cal. BP (GrA53747 and GrA53627) (Sturt *et al.,* 2014, Griffith *et al.,* 2015). In core VC25 two weighted mean dates were produced at -13.73 and -13.75 of 8,990-8,630 (Beta288890, GrA53517, GrA53518) and 8990-8640 cal. BP (GrA53523, Beta288891) (Sturt *et al*., 2014). In some cases, however, the dates were not statistically comparable, and a mean could not be calculated. In core V06, for example, two limiting points were produced at -12.64m OD from the humic and humin fractions: 9671-9521 cal. BP (GrA53526) and 9252-8810 cal. BP (GrA53525) (Sturt *et al*., 2014; Griffith *et al*., 2015). The dates from cores VCJ15 and VC24 were from sub-optimal material and were therefore rejected (Sturt *et al*., 2014; Griffith *et al*., 2015).

From the peat there is pollen evidence for a shift from pine and hazel dominated woodland to a mixed oak woodland with hazel and elm, which reflects the regional changes in woodland composition at this time (e.g., Birks, 1989; Hosfield *et al.,* 2007a). The plant macrofossils indicate the presence of a freshwater marsh-fen environment, with reeds, sedges, and grass present (Sturt *et al.,* 2014). There is also evidence of charred remains indicating burning episodes, for example charred fragments of *Populus* spp. which were dated to 9252-8982 cal. BP (GrA53628) (Sturt *et al*., 2014; Griffith *et al*., 2015).

In most cores from Hinkley Point, the contact with the overlying estuarine silts appeared to be erosional, but in core V06 the peat appeared to have an intact regressive contact, with the sediment $^{\delta 13}$C values and C/N ratios indicating progressive change (Sturt *et al.,* 2014). This contact was dated to 9072-8724 cal. BP (GRA53519) at -12.58m OD and marks the onset of local brackish conditions (Sturt *et al.,* 2014). Overlying the basal peat were marine silts, with all proxies across the six cores showing a transition from freshwater to saltwater conditions. In cores VC06 and VC15 there were intercalated organic deposits, which indicate it was not a smooth transition, but there were brief periods where sediment supply into the wetlands overtook sea-level rise, allowing saltmarsh to develop over the mudflats (Sturt *et al*., 2014). The combined results from Hinkley Point show a gradually drowning terrestrial early Holocene landscape, similar to the sequence found at Aberayron (Sturt *et al.,* 2014; Griffiths *et al.,* 2015; Wilks, 1977; Kitely, 1975).

Table 2.9:Radiocarbon dates from the early Holocene at Hinkley Point, cores VC06, VC15, VC24 (Griffiths *et al.,* 2015)

Core	Lab code	Calibrated Date (cal. BP)	Nature of contact/ material dated	Altitude (m OD)	RSL (m)	Index or limiting point	Notes
VC06	GrA53520	8976-8597	Top of peat horizon, humic fraction	-12.58	-18.15	Index	Statistically inconsistent
VC06	GrA53519	9072-8724	Top of peat horizon, humin fraction	-12.58	-18.15	Index	
VC06	Beta288884	9255-8997	Top of peat horizon, wood fragment	-12.58	-18.15	Index	Statistically too old
VC06	Beta288885	8998-8655	Wood fragment within peat	-12.60	-18.37	Index	
VC06	GrA53525	9252-8810	Bottom of peat horizon, humin fraction	-12.64	-16.19	Limiting	Base of basal peat
VC06	GrA53526	9671-9521	Bottom of peat horizon, humic fraction	-12.64	-16.19	Limiting	Base of basal peat
VC15	GrA53625	8998-8655	Top of lower peat horizon, humin fraction	-13.74	-19.31	Index	
VC15	GrA53746	9254-9005	Top of lower peat horizon, humic fraction	-13.74	-19.31	Index	Statistically older
VC15	GrA53628	9252-8982	Burning horizon within lower peat, charred macrofossils of *Populus* spp.	-13.76	-19.53	Index	
VC15	GrA53747	9502-9298	Bottom of lower peat horizon, humic fraction	-13.80	-17.35	Limiting	Base of basal peat Statistically comparable
VC15	GrA53627	9553-9456	Bottom of lower peat horizon, humin fraction	-13.80	-17.35	Limiting	Base of basal peat Statistically comparable
VC24	Beta289591	9245-8996	Organic Sediment	-13.46	-19.23	Index	Statistically older
VC24	Beta288889	8978-8593	Plant macrofossil within peat	-13.46	-19.23	Index	

Table 2.10: Radiocarbon dates from the early Holocene at Hinkley Point, cores VC25, VC34, VCJ15 (Griffiths *et al.*, 2015)

Core	Lab code	Calibrated Date (cal. BP)	Nature of contact/ material dated	Altitude (m OD)	RSL (m)	Index or limiting point	Notes
VC25	Beta288890	8976-8594	Phragmites leaves, Dates reed swamp conditions at end of peat formation	-13.73	-19.30	Index	Statistically comparable
VC25	GrA53517	8980-8640	Top of peat horizon, humic fraction	-13.73	-19.30	Index	Statistically comparable
VC25	GrA53518	9021-8729	Top of peat horizon, humin fraction	-13.73	-19.30	Index	Statistically comparable
VC25	GrA53523	8980-8640	Bottom of peat horizon, humin fraction	-13.75	-17.30	Limiting	Base of basal peat Statistically comparable
VC25	Beta288891	8977-8607	Phragmites, Dates reed swamp conditions	-13.75	-17.30	Limiting	Base of basal peat Statistically comparable
VC25	GrA53521	9091-8776	Bottom of peat horizon, humic fraction	-13.75	-17.30	Limiting	Base of basal peat Statistically older
VC34	Beta288892	8982-8647	Phragmites leaves within peat, dates reed swamp conditions	-13.62	-19.39	Index	
VCJ15	Beta288886	8721-8479	Wood fragment in organic silt	-11.28	-16.85	Index	Statistically inconsistent
VCJ15	Beta289589	9302-9028	Plant macrofossil	-11.32	-17.09	Index	Statistically inconsistent
VCJ15	Beta288887	9590-9458	Organic silt	-11.32	-17.09	Index	Statistically inconsistent
VCJ15	Beta288888	8626-8425	Wood fragment within organic silt	-11.34	-14.89	Limiting	Base of basal peat Statistically inconsistent
VCJ15	Beta289590	9469-9262	Organic silt	-11.34	-14.89	Limiting	Base of basal peat Statistically inconsistent

2.4.2.4 Summary: Early Holocene

There are some data from the early Holocene in Wales to suggest a rapidly rising sea level, which correlates with the rate of global eustatic sea-level rise (Lambeck *et al.,* 2014). Evidence from the Menai Straits on the north coast of Wales indicates sea level rose at a rate of 6.72mm a^{-1} between 10654 to 8407 cal. BP (Roberts *et al.,* 2011). However, there is only one data point from the west coast at Aberayron (Wilks, 1977; Kitely, 1975). There is more data from the south coast, with SLIPs and limiting points indicating a rate of sea level rise of 3.28mm a^{-1} at Porlock Marsh (Jennings *et al.,* 1998), 1.82mm a^{-1} from Hinkley Point (Sturt et al. 2014, Griffith et al. 2015) and 5.3mm a^{-1} from the data published by Heyworth and Kidson (1982) for sites such as Stolford, Burnham, Highbridge, Wall Common and Port Talbot . The sea level data for the early Holocene appears to show a lower rate of sea-level rise on the south coast, which could be due to differential isostatic uplift. More data are needed, however, to determine if the data is influenced by a regional trend, rather than site-specific factors.

2.4.3 Mid Holocene (7500 to 5000 cal. BP)

2.4.3.1 North Wales

At Hendre Fawr on the Clwyd Plain (SH 96390 76586) Bedlington (1994) took 21 cores in a north-south transect, which revealed that unconsolidated sediments deepened towards the sea, with four phases of organic, terrestrial deposition separated by inorganic clay deposits. Biostratigraphic analysis was undertaken on two cores, which revealed the clays were dominated by marine and brackish diatoms (Bedlington, 1994). At the base of the core, silty clays were found with brackish and marine diatoms. Two peat intercalations were found at -2.50 and -1.00m OD which were eroded by sandy channel fill deposits, and in core HF29 the lower peat produced an early Holocene date (see section: early Holocene). Fine silty clays with intercalated peats were found between -1.00m and +2.00m OD. A peat layer at +1.25m OD produced a limiting point of 7250–6320 cal. BP (Hv 17812) with diatom evidence for a change from upper marsh to fen environments, with RSL estimated at

-2.74m OD (Bedlington, 1994). This correlates with the date of 7470–6910 cal. BP obtained from Morfa Penrhyn at -3.67m OD (Hv 17815) which was obtained from blue grey clay which contained evidence for a marine brackish environment, which progressed to an upper saltmarsh and then to a fen, with RSL estimated to be at -3.47m OD (Bedlington, 1994).

Table 2.11: Radiocarbon dates for the mid Holocene, north Wales (Bedlington, 1994; Heyworth and Kidson, 1982; Tooley, 1978)

Reference	Site Name	Lab code	Calibrated Date range (cal. BP)	Nature of contact	Altitude m OD	RSL (m)	Notes
Bedlington, 1994	Clwyd Plain Hendre Fawr (1)	HV17810	5432-4527	Regressive	1.87	-2.14	
Bedlington, 1994	Clwyd Plain Hendre Fawr (2)	HV17811	5739- 4872	Transgress-ive	1.70	-1.91	
Bedlington, 1994	Clwyd Plain Hendre Fawr (3)	HV17812	7248- 6350	Regressive	1.27	-2.74	
Bedlington, 1994	Clwyd Plain Hendre Fawr (5)	Hv 17813	7250–5490	Regressive		-	Rejected - too young
Bedlington, 1994	Morfa Penrhyn	HV17815	7460- 6965	-	0.28	-3.47	
Heyworth and Kidson, 1982; Tooley, 1978	Rhyl Beach	HV4348	5587- 5320	Regressive	2.43	-1.58	

At Hendre Fawr, the peat was overlain by silty clay containing marine-brackish diatoms and produced two dates of 5440-4520 cal. BP (+1.87m OD, Hv 17810) and 5730-4870 cal. BP (+1.7m OD, Hv 17811) (Bedlington, 1994). This correlates with the date of 5590–5310 cal. BP obtained from -1.58m OD at Rhyl beach (SJ 0310 8261) obtained by Tooley (1974, 1978) from a regressive overlap of peat underlying silty clay.

At Hendre Fawr organic sedimentation dominated between +2.00 m and +3.00 m OD separated into two horizons by a thin organic clay (Bedlington, 1994). The peats contained a high frequency of tree pollen suggesting the site may have been partially wooded. The fine grained, organic clay separating the two peat horizons contained brackish upper saltmarsh diatoms with saltmarsh pollen. Overlying the organic horizons was an iron-stained silty clay containing sand (Bedlington, 1994). The SLIPs from Hendre Fawr imply sea level rose at a rate of 0.3mm a^{-1} during the mid Holocene (Bedlington, 1994).

2.4.3.2 West Wales

On the west coast, there are a series of mid Holocene limiting points from Borth and Ynyslas (SN 60655 92807). Borth and Ynyslas have been investigated by many authors (Adam and Haynes, 1965; Godwin, 1943; Godwin and Newton, 1938; Godwin and Willis, 1961; Godwin and Willis, 1964; Haynes and Dobson, 1969; Prince, 1988; Taylor, 1973; Wilks, 1977, 1979). The stratigraphy reveals silty sands at the base of the sequence, overlain by blue-grey clay that Keeping (1878) found contained the marine shell *Scrobularia piperata*. This clay layer has been investigated by many authors and has been described as "*Scrobularia*" or "OD" clay as it occurs between 0 and -1m OD (Keeping, 1878; Prince, 1988; Godwin, 1948; Wilks, 1977, 1979). The OD clay was overlain by an extensive submerged forest peat which has been found to extend across the west coast of Wales (Heyworth *et al.,* 1985; Prince, 1988; Wilks, 1977). Wilks (1977, 1979) found evidence from the submerged forest peat for a pollen succession of *Alnus*, *Betula* and *Pinus* and a developing reed swamp. The forest peat was dated to 6179-5663 cal. BP (HAR1020, -0.25m OD) and 5586 to 5312 cal. BP (HAR 1019, +0.27m OD) (Figure 2.22) (Wilks, 1977, 1979).

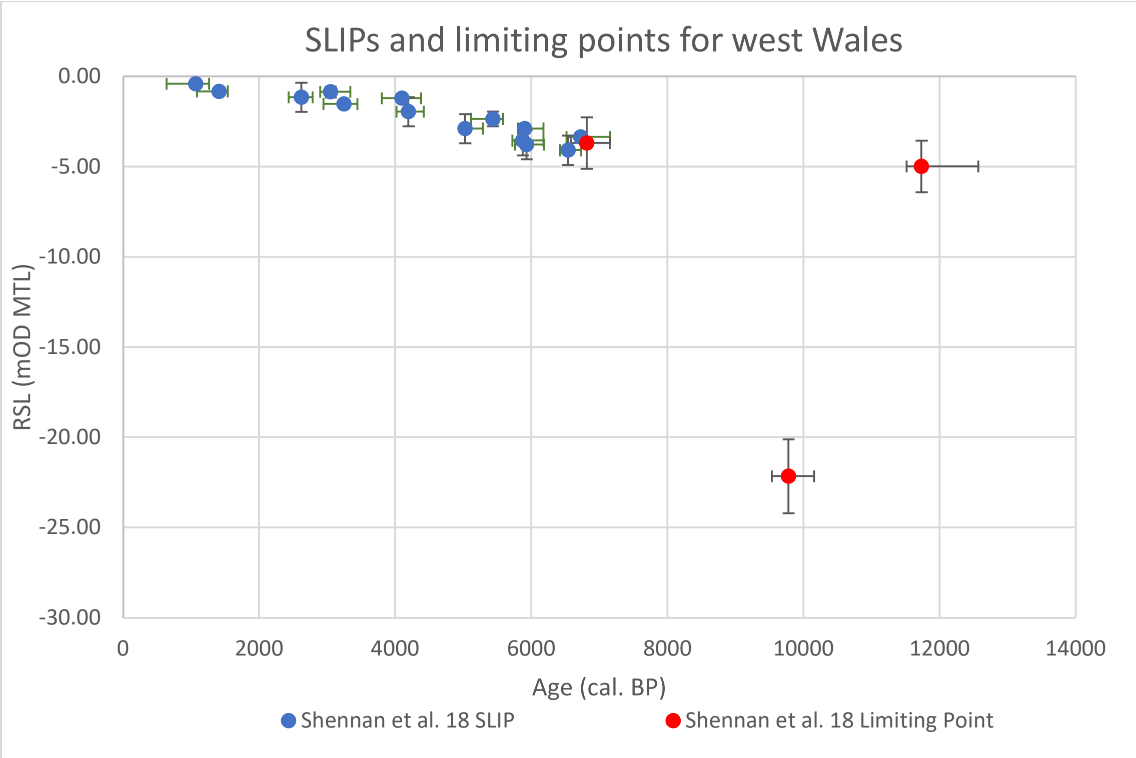

Figure 2.22: Sea-level curve for the west coast of Wales, with data from Shennan *et al.,* 2018 (blue: SLIP, red: limiting point).

Table 2.12: Radiocarbon dates for the mid Holocene, west Wales (Wilks, 1977, 1979; Heyworth *et al.,* 1985, Taylor, 1973; Heyworth and Kidson, 1982)

Reference	Site Name	Lab code	Calibrated Date range	Nature of contact	Altitude MSL	RSL (m)	Limiting or index point	Notes
Wilks, 1977, 1979	Ynyslas	HAR1019	5585 - 5312-		0.30	-2.36		
Wilks, 1977, 1979	Ynyslas	HAR1020	6178 - 5663		-0.23	-2.89		
Heyworth *et al.,* 1985	Clarach	HAR300	6185 - 5717		-1.22	-3.78		
Heyworth *et al.,* 1985	Clarach	HAR301	6731 - 6324		-1.53	-4.09		
Heyworth *et al.,* 1985	Clarach	HAR299	6178 - 5653		-1.00	-3.56		
Taylor, 1973	Clarach	NPL113	7152 - 6563		-1.95	-3.69		basal peat; top eroded
Heyworth *et al.,* 1985	Clarach	Har 1128	6785 - 6319	Beach peat	-4.56 (MHWST)		Limiting	Rejected: no link to sea level
Wilks, 1979	Clarach	Har 574	6290 - 5485	Woody fragments in gravel	-4.36 (MHWST)		Limiting	Rejected: no link to sea level
Heyworth and Kidson, 1982		Q382	7155 - 6405		-0.50	-3.36		

At Clarach (SN588 830) a deep borehole revealed glacial gravel at a depth of 18m and bedrock was encountered at 26m (Wilks, 1977; Heyworth *et al.,* 1985). The glacial gravel was overlain by a freshwater sequence of clays and silts which contained a typical late-glacial pollen spectrum and gave a radiocarbon date of 12511-10780 cal. BP (BIRM 229) (Taylor, 1973) (Heyworth *et al.,* 1985).

A series of boreholes behind the storm beach at Clarach revealed a peaty clay between -3.46 and -2.16m OD which was dated to between 6785-6319 cal. BP (HAR1128) and 6179-5654 cal. BP (HAR 299) and marked the onset of marine influence (Heyworth *et al.,* 1985; Wilks, 1977). The peat behind the storm beach was contemporary with the submerged forest on the foreshore, which was dated to 6733-6325 cal. BP (HAR 301, -3.99m OD) (Wilks, 1977). Taylor (1973) found pollen of *Pinus, Alnus, Corylus, Betula* and *Quercus* from the forest peat on the foreshore, and a date of 7154 to 6563 cal. BP (NPL113) was recorded from a pine stump which was found at -4.41m OD. The tree ring analysis from this pine stump implied it died prematurely at 70-80 years, which Taylor (1973) interpreted as being caused by rising sea level. Heyworth *et al.* (1985) found layers of silty clay in the forest peat on the foreshore containing freshwater diatoms indicating river flood events. Diatom evidence indicated a change from forest to marsh peat, which formed at -3.68m OD and dated to 6186-5719 cal. BP (HAR 300, Heyworth *et al.,* 1985).

Overlying the forest peat, a blue grey clay occurred between -1.5 and -0.5m OD, which Taylor (1973) interpreted as a marine clay similar to deposits found at Borth Bog and Ynyslas due to assumptions based on physical properties and similar altitude. Unlike the marine clay at Borth, however, the clay at Clarach contained coarse gravel and freshwater diatoms, indicating that it was not marine (Wilks, 1977). At +1.1m OD a woody peat containing pollen indicated a change from alder woodland to an open, disturbed habitat dominated by grass, sedges and other herbs such as *Plantago lanceolata* (Heyworth *et al.,* 1985). There was also evidence for an increase in marine diatoms indicating an intensification of flooding episodes, however there was no dating evidence for this event (Heyworth *et al.,* 1985).

2.4.3.3 South Wales

Most sea level data available for the mid Holocene are derived from the Severn Estuary and the Bristol Bay area. There has been a great deal of research into the deep sequences of estuarine silts and brackish/freshwater peats which developed along the edges of the Severn estuary in response to sea-level rise (Allen and Rae, 1987; Culver and Banner, 1978; Devoy, 1979; Evans and Thomspon, 1979; Hawkins, 1971; Heyworth and Kidson, 1982; Shennan, 1983).

The stratigraphy of the Severn Estuary is characterised by the Wentlooge formation, named after intertidal exposures on the Wentlooge level (ST23 77 to ST31 81), which consist of alluvial clay between layers of peat (Allen, 1987). Boreholes taken from the Wentlooge level revealed 3m of sand and gravel overlain by 12m of pale green estuarine silty clay containing siliceous marine sponge spicules as well as open-marine/estuarine diatoms and foraminifera (Anderson and Blundell, 1965;

Allen, 1987). This estuarine clay is known as the Lower Wentlooge level and was overlain by a thin peat layer with macrofossil evidence of a *Phragmites* swamp, which developed into a broad-leaved forest (Allen, 1987). The peat was overlain by the Upper Wentlooge Formation which comprised of 2.5m of pale green estuarine clay, containing the same estuarine diatom and foraminifera assemblage as the Lower Wentlooge formation (Allen, 1987).

The Wentlooge formation was described by Allen and Haslett (2007) where it was exposed at the site of Redwick. Five peat layers were found intercalated within pale-grey silts which ranged from structureless to laminated. Six transgressive-regressive cycles were identified over the last 8000 years and five SLIPs were produced for the mid Holocene, indicating that the rate of sea-level rise was between 1 to 2mm a^{-1}. This is much lower than the rate of 10mm a^{-1} for the early Holocene at Redwick (see section: early Holocene). Other sites in south Wales and the Bristol Bay area show a similarly slower rate of sea-level rise in the mid Holocene, such as the Gordano Valley with a rate of 0.91mm a^{-1} (Hill *et al.,* 2007). At Porlock marsh during the mid Holocene, sea level rose at a rate of 2mm a^{-1}, in contrast with 16mm a^{-1} during the early Holocene (Jennings *et al.,* 1998) (see early Holocene section).

Table 2.13: Radiocarbon dates from the mid Holocene at Redwick, Porlock and the Gordano valley (Allen and Haslett, 2007; Hill *et al.,* 2007; Jennings *et al.,* 1998) RWL = reference water level

Reference	Site Name	Region	Lab code	Calibrated Date	Altitude MSL	RSL (m)	Notes
Allen and Haslett 2007	Redwick	Bristol Channel	Beta113004	5887-5476	1.38	-5.35	
Allen and Haslett 2007	Redwick	Bristol Channel	Beta128779	6655-6299	-0.28	-6.37	
Allen and Haslett 2007	Redwick	Bristol Channel	Beta193623	6950-6679	-0.51	-7.24	
Allen and Haslett 2007	Redwick	Bristol Channel	Beta134641	7156-6675	-0.98	-7.91	
Allen and Haslett 2007	Redwick	Bristol Channel	Beta134640	7571-7324	-2.44	-9.37	
Hill *et al.,* 2007	Gordano Valley	Bristol Channel	Beta189682	5570 5305	2.04	-5.20	RWL defined by authors
Hill *et al.,* 2007	Gordano Valley	Bristol Channel	Beta189683	5715 5587	1.92	-5.31	RWL defined by authors
Hill *et al.,* 2007	Gordano Valley	Bristol Channel	Beta189684	6181 5935	1.77	-6.15	RWL defined by authors
Hill *et al.,* 2007	Gordano Valley	Bristol Channel	Beta189685	6882 6676	0.48	-6.64	RWL defined by authors
Jennings *et al.,* 1998	Porlock	Bristol Channel	Be81653	5921 5661	-1.46	-6.46	
Jennings *et al.,* 1998	Porlock	Bristol Channel	Be61543	6179 5657	-0.48	-5.28	
Jennings *et al.,* 1998	Porlock	Bristol Channel	Be61542	6396 5617	-0.67	-5.99	
Jennings *et al.,* 1998	Porlock	Bristol Channel	Be81654	7249 6887	-3.57	-8.57	

The Wentlooge formation has also been described at the site of Goldcliff (Smith and Morgan, 1989). The Lower Wentlooge estuarine clay was found, overlain by a submerged forest peat at approximately 0m OD. It is likely that as the rate of sea-level rise slowed during the mid Holocene, peat development became established. A similar estuarine clay was described as 'OD clay' by Kidson and Heyworth (1973), as it was consistently found near 0m OD. Similar clay with an upper surface around 0m OD has been found across the Somerset levels, overlain by submerged forest peat intercalated with fresh/brackish clay (Kidson and Heyworth, 1973). This also correlates with the '*Scrobularia'* or 'OD' clay described on the west coast at Borth (Wilks, 1977, 1979; Heyworth and Kidson, 1982).

The submerged forest peat is exposed over 30km in the intertidal zone along the Gwent coast, between layers of grey clay, often containing reeds and estuarine foraminifera (Smith and Morgan, 1989). At Goldcliff, the forest peat was intercalated by a thin layer of soft blue clay at +0.25m OD which contained estuarine diatoms, possibly reflecting the upper Wentlooge formation. SLIPs from the mid Holocene at Goldcliff indicate that there was a rate of sea-level rise of 1.25mm a^{-1}, similar to other sites across South Wales (Smith and Morgan, 1989). This implies that there was a reduction in the rate of sea-level rise across south Wales.

Table 2.14: Radiocarbon dates from the mid Holocene at Goldcliff (Smith and Morgan, 1989)

Reference	Site Name	Region	Lab code	Calibrated Date	Altitude MSL	RSL (m)
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR654	5992 5651	0.65	-6.08
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR656	6294 5944	0.45	-6.28
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR655	6397 6003	0.51	-6.22
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR776	6438 6017	1.06	-5.03
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR657	6530 6026	0.28	-5.81
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR778	6634 6303	0.97	-5.96
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR658	6878 6455	0.07	-6.66
Smith and Morgan 1989	Goldcliff	Bristol Channel	CAR659	7003 6567	0.01	-6.72

At Burnham on Sea two exposures in a cliff face revealed three thin layers of peat intercalated with silt (Druce, 2000). The clay at the base of the cliff was overlaid by peat, with pollen and foraminifera evidence for an intact regressive contact dated to 7426-7030 cal. BP at -3.10 OD (WK5298). Pollen from the peat reflected a brackish-freshwater environment near a fen of birch and willow (Druce, 2000). The peat was overlain by a second clay unit, which contained pollen and foraminifera evidence for a middle/high marsh environment, indicating higher sea level. At the top of the lower profile was a second peat layer, which was dated to 6532 to 6223 cal. BP at -2.86m OD (WK5297) with pollen and foraminifera indicating a brackish-freshwater reed swamp near MHWST (Druce, 2000). The clay at the base of the upper profile contained foraminifera and pollen from a low-middle saltmarsh. The regressive overlap into the third peat was dated to 6294 to 5955 cal. BP at 0.17m OD (WK5299) which contained pollen and foraminifera evidence of a *Phragmites* marsh above HAT (Druce, 2000).

Table 2.15: Radiocarbon dates from the mid Holocene at Burnham-on-sea (Druce, 2000; Heyworth and Kidson, 1982)

Reference	Site Name	Lab code	Calibrated Date	Nature of contact/material dated	Altitude (mOD)	RSL (m)
Druce, 2000	Burnham on Sea	WK5300	5647-5324	Peat	-0.02	-5.59
Druce, 2000	Burnham on Sea	WK5299	6294-5955	Regressive	0.17	-6.01
Druce, 2000	Burnham on Sea	WK5297	6532-6223	Transgressive	-2.89	-9.07
Druce, 2000	Burnham on Sea	WK5298	7426-7030	Regressive	-3.10	-9.28
Heyworth and Kidson 1982; Godwin *et al.,* 1958	Burnham on Sea	Q134	7428-6808	-	-4.60	-10.78

At the site of Caldicott Pill, basal gravel was overlain by pale grey silts with diatom evidence for marine conditions becoming progressively less saline (Scaife and Long, 1995). The overlying peat contained pollen evidence for reed swamp, which was replaced by mixed-oak hazel woodland, damp alder carr and freshwater ponds. Above the peat, there was a return to minerogenic sedimentation and marine conditions similar to the present (Scaife and Long, 1995). The regressive upper peat contact was dated at two locations: 7424-7166 cal. BP at -0.82m OD (B79887), and 7658 -7430 cal. BP at -2.48m OD (B79886). These data correlate with the SLIPs produced across South Wales and implies that there was widespread reduction in marine conditions along the north and south coast of the Bristol Channel, possibly in response to a regional reduction in the rate of sea-level rise (Scaife and Long, 1995).

Table 2.16: Radiocarbon dates for the mid Holocene at Caldicott Pill (Scaife and Long, 1995)

Reference	Site Name	Region	Lab code	Calibrated Date	Altitude MSL	RSL (m)
Scaife and Long 1995	Caldicot pill	Bristol Channel	B79887	7424-7166	-0.82	-7.52
Scaife and Long 1995	Caldicot pill	Bristol Channel	B79886	7658-7430	-2.48	-9.86

Heyworth and Kidson (1982) produced a synthesis of sea level data for sites across Wales, including Bridgwater Bay and Somerset. At the site of Stolford, Kidson and Heyworth (1973) recovered a sequence of clay with multiple intercalated peat layers behind a shingle ridge. These dates show a rise in sea level from -7.57m OD at 7421-6889 cal. BP (NPL148) to -3.47m OD at 5886-5081 cal. BP (I3395) at a rate of 1.75mm a^{-1} (Kidson and Heyworth, 1973; Heyworth and Kidson, 1982). At Shapwick Heath, there was a succession from saltmarsh to freshwater fen dated to 6555-5994 cal. BP at 0.3mOD (Q423). The data produced by Heyworth and Kidson (1982) reinforces that there was widespread peat development during the mid Holocene in South Wales. However, many sites do not have published palaeoenvironmental analysis, so the relationship to sea level is often unclear.

Table 2.17: Radiocarbon dates for the mid Holocene for south Wales and Bristol Bay published in Heyworth and Kidson, 1982

Reference	Site Name	Region	Lab code	Calibrated Date	Altitude MSL	RSL (m)
Heyworth and Kidson 1973, 1982	Stolford 6V	Bristol Channel	I3395	5886-5081	2.30	-3.47
Heyworth and Kidson 1973, 1982	Stolford 6F	Bristol Channel	I3396	6302-5664	0.80	-4.97
Heyworth and Kidson 1973, 1982	Stolford 5B	Bristol Channel	NPL147	6384-5931	-1.00	-6.77
Heyworth and Kidson 1973, 1982	Stolford 6F	Bristol Channel	I3397	6394-5769	-0.50	-6.88
Heyworth and Kidson 1973, 1982	Stolford 5C	Bristol Channel	NPL148	7412-6889	-2.00	-7.57
Heyworth and Kidson 1982	Weston-super-Mare	Bristol Channel	St 3297	5466-4871	0.50	-5.93
Heyworth and Kidson 1982	-	Bristol Channel	IGS35	6397-5909	0.23	-6.07
Heyworth and Kidson 1982, originally Godwin and Willis 1961	Shapwick Heath	Bristol Channel	Q423	6555-5994	0.30	-5.88
Heyworth and Kidson 1982	Kingston Seymour	Bristol Channel	I4844	6654-6192	0.15	-6.15
Heyworth and Kidson 1982, originally Godwin *et al.,* 1958	Tealham Moor	Bristol Channel	Q120	6444-5916	0.01	-6.17
Heyworth and Kidson 1982 originally Godwin *et al.,* 1958	Tealham Moor	Bristol Channel	Q126	6715-6184	0.01	-6.17
Heyworth and Kidson 1982	Clevedon Pill	Bristol Channel	St 3292	6897-6323	-0.90	-7.20
Heyworth and Kidson 1982	Freshwater West, Pembrokeshire	South Wales (Pembrokeshire)	Q530	7156-6501	-2.00	-5.29
Heyworth and Kidson 1982	Kenn Pier	Bristol Channel	St 3282	7257-6678	-1.50	-8.24
Heyworth and Kidson 1982, originally Godwin and Willis 1961	Margam	South Wales (Glamorgan)	Q274	6717-6031	-3.10	-10.10
Heyworth and Kidson 1982 originally Godwin and Willis 1961	Margam	South Wales (Glamorgan)	Q275	7416-6744	-3.20	-10.60

After the rapid sea-level rise seen during the early Holocene, the rate of sea-level rise was more reduced in the mid Holocene, between 1-2mm a^{-1}. Across South Wales and the Bristol Bay area there is a relatively uniform stratigraphy described as the Wentlooge formation, with two or more layers of estuarine clay intercalated with freshwater-brackish peat. The estuarine clay outcrops consistently near 0m OD across the Severn Estuary and the Somerset levels. There is evidence for widespread peat development that lasted between 7658-7430 cal. BP (B79886, -2.48m OD, Caldicott Pill, Scaife and Long 1995) and 5581-5071 cal. BP (Be142351, 0.43m OD, Oldbury on Severn, Haslett *et al.,* 2001). The uniform stratigraphy indicates a regional response to a lower rate of sea-level rise, leading to peat development and coastal expansion across the south coast of Wales and the Bristol Bay area.

2.4.3.4 Summary: Mid Holocene

There is limited data available for the north coast of Wales, but at the site of Hendre Fawr layers of estuarine clay intercalated with peat produced SLIPs which suggested a rate of sea level rise of 0.3mm a^{-1} (Bedlington, 1994). For the west coast, at Borth and Clarach an estuarine clay was found overlain by a submerged forest peat at around 0m OD. This submerged forest peat was dated to 6733-6325 cal. BP (HAR 301, -3.99m OD) at Clarach (Wilks, 1977) and produced dates ranging from 6,179-5,663 cal. BP (HAR1020, -0.25m OD) and 5586 to 5312 cal. BP (HAR 1019, +0.27m OD) at Ynyslas and Borth (Wilks, 1977, 1979). At Clarach, the SLIPs indicated a rate of sea-level rise of 0.31mm a^{-1}, which correlates with the data from Hendre Fawr.

A similar stratigraphy is found across the Severn estuary and the Somerset levels, with estuarine clay overlain by submerged forest peat around 0m OD. This submerged forest peat produced similar dates to those on the west coast, for example at Goldcliff the peat was dated to 6530- 6026 cal. BP (CAR657, 0.28m OD, Smith and Morgan, 1989).

The rate of sea-level rise on the south coast of Wales is slightly higher than the west coast, averaging 1-2mm a^{-1}. However, this is significantly lower than the rate of sea-level rise during the early

Holocene. More data are needed from the north and west coasts of Wales to confirm if the rate of sea-level rise was lower on the west coast, compared to the south, or if local or taphonomic factors have affected the results.

2.4.4 Late Holocene (4000 BP to present).

During the late Holocene there was a further decline in the rate of eustatic sea level rise as the influx of meltwater from the LGM stopped (Lambeck *et al.,* 2014). Sea level data for the late Holocene in Wales are limited, with no evidence from the north coast, and only limiting points from Ynyslas and Clarach on the west coast (Bedlington, 1994; Heyworth *et al.,* 1995; Prince, 1988; Roberts *et al.,* 2011; Wilks, 1977).

2.4.4.1 West Coast

At Borth and Ynyslas, basal peat was overlain by estuarine clay, with submerged forest and saltmarsh peat from 0m OD to the surface (see section: mid Holocene). A number of late Holocene limiting points were established from the peat (Wilks, 1977; Heyworth and Kidson, 1982). At Ynyslas, a date of 4381-3894 cal. BP (HAR 1018, 1.15m OD) was obtained from the middle of a saltmarsh peat containing marine-brackish diatoms (Wilks, 1977). Nearby, the top of a similar peat layer was dated to 3442-3040 cal. BP (HAR1017, 1.03m OD) and contained brackish diatoms and *Sphagnum* pollen, indicating a regeneration of bog growth (Wilks, 1977). At Ynyslas, the brackish peat was overlain by saltmarsh peaty clay which contained *Juncus maritimus*, high marsh foraminifera and freshwater diatoms as well as reed swamp pollen (Adam and Haynes, 1965; Godwin, 1943; Wilks, 1977). This freshwater saltmarsh peat was dated to 1538-1299 cal. BP (HAR 1016, 1.83m OD), with diatom evidence suggesting a change from brackish-limnic[23] to a weakly brackish environment (Wilks, 1977). These limiting points imply a rise in sea level of 0.12mm a^{-1} at Ynyslas during the late Holocene. However, care must be taken as many of the samples are from weakly brackish-freshwater peat and the relationship to sea level is ambiguous.

[23] low salt concentration

Table 2.18: Radiocarbon dates for the late Holocene for Clarach and Ynyslas (Heyworth and Kidson, 1982; Wilks, 1977, Wilks, 1979; Heyworth *et al.,* 1985).

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Nature of contact/ material dated	Altitude (m oD)	RSL (m)	Index or limiting point
Clarach	Heyworth and Kidson 1982, Heyworth *et al.,* 1985	HAR1572	1085-927	Peat	1.95	-0.41	Limiting
Clarach	Heyworth and Kidson 1982, Heyworth *et al.,* 1985	HAR1575	2782-2379	Peat	1.40	-1.16	Limiting
Clarach	Heyworth and Kidson 1982, Heyworth *et al.,* 1985	HAR1579	4418-3976	Peat	0.60	-1.96	Limiting
Ynyslas	Heyworth and Kidson 1982, Wilks 1977, 1979	HAR1016	1538-1299	Peat	1.83	-0.83	Limiting
Ynyslas	Heyworth and Kidson 1982, Wilks 1977, 1979	HAR1017	3441-3037	Peat	1.03	-1.53	Limiting
Ynyslas	Heyworth and Kidson 1982, Wilks 1977, 1979	HAR1018	4381-3892	Peat	1.15	-1.21	Limiting

At Clarach a similar stratigraphy to Ynyslas was recorded, with estuarine clay overlain by peat at 0m OD (see section: mid Holocene) (Wilks, 1977; Heyworth *et al.,* 1985; Heyworth and Kidson, 1982). Three dates were obtained from a terrestrial sedge peat which was dated to 4420-3975 cal. BP (HAR 1579, 0.6m OD), and contained freshwater diatoms. This was overlain by silty peat which was dated to 2784-2379 cal. BP (HAR 1575) at 1.4m OD and 1261-927 cal. BP (HAR 1572) at 1.95m OD and contained diatom evidence for freshwater conditions, with some marine benthic diatoms indicating occasional tidal flooding (Heyworth *et al.,* 1985). The limiting points from Clarach appear to indicate a rate of sea-level rise of 0.44mm a^{-1}, however care must be taken as the limiting points are from weakly brackish terrestrial peat.

2.4.4.2 South coast

The sediments of the Severn Estuary are characterised by the Wentlooge formation, which formed throughout the Holocene (see sections: early and mid Holocene). The transgressive contact which marks the end of the middle Wentlooge formation has been dated to between 4149-3780 cal. BP (CAR-1499) and 2751-2350 cal. BP (Beta-73059) (Bell and Neumann, 1997). The dating evidence suggests that the marine transgression was time transgressive (see Table 2.19). Bell and Neumann (1997) found a concentration of Bronze Age archaeology on the surface of the middle Wentlooge peat, such as roundhouses at Rumney and Chapel Tump, rectangular buildings at Redwick and charcoal scatters at Cold Harbour, which were sealed by the Upper Wentlooge marine clay. The Upper Wentlooge clay accumulated rapidly during rising sea level until reclamation and drainage occurred in the Romano British period (Bell and Neumann, 1997).

Table 2.19: Radiocarbon dates from the end of the middle Wentlooge formation, published in Bell and Neumann (1997).

Site	Reference	Lab code	Calibrated date	Material dated
Magor	Allen and Rippon 1997, Bell and Neuman 1997	Beta-73059	2751-2350	Peat
Vurlong Reen	Walker and James 1993, Bell and Neuman 1997	Beta-63590	2748-2379	Peat
Goldcliff	Bell 1993, Bell and Neuman 1997	CAR-1438	2847-2437	Peat
Goldcliff	Bell and Neuman 1997	SWAN-31	3056-2751	Peat
Goldcliff	Bell and Neuman 1997	CAR-1499	4149-3780	Peat

The middle Wentlooge formation was found at the site of Goldcliff, where Smith and Morgan (1989) found *Sphagnum* peat, containing *Calluna* pollen, which was dated to 3868-3479 cal. BP (CAR-645, -4.82m OD). This was overlain by reed swamp peat containing *Phragmites* dated to 3543-3116 cal.

BP (CAR-644, -4.53m OD). The reed swamp peat was overlain by blue-grey estuarine clay representing the Upper Wentlooge formation. Palaeonvironmental analysis suggested that marine conditions retreated after the mid Holocene, and during the late Holocene the site transitioned to a *Sphagnum* raised bog followed by reed swamp, until estuarine conditions returned in the present day (Smith and Morgan, 1989).

The data from Goldcliff correlates with other radiocarbon dates across the south coast of Wales. At Margam, Glamorgan, the upper surface of a submerged forest peat was dated to 3913-3397 cal. BP (-2.40m OD, Q265). This was overlain by grey sand and silty clays which were described as marine (Gordon and Willis, 1961). At the site at Llanwern, a transgressive boundary between the peat and clay containing marine diatoms was dated to 3056-2382 cal. BP (+2.98m OD, Q691) (Godwin and Willis, 1961; Heyworth and Kidson, 1982).

At the site of Redwick, the top of the middle Wentlooge formation was dated to 3826-3452 cal. BP (Beta144020) from a regressive contact at 3.08m OD, with foraminifera evidence for a transition from mid to high marsh (Allen and Haslett, 2007). The peat was overlain by 1.81m of estuarine pale green silts which graded upwards into a thin peat layer, with foraminifera transitioning from a mid to high marsh (Allen and Haslett 2007). The thin peat was dated to 3316-2948 cal. BP at 4.58m OD (Beta144021) (Allen and Haslett, 2007).

The transition from the Middle Wentlooge peat to marine clays of the Upper Wentlooge formation can be seen across the south coast of Wales, indicating a regional marine transgression. There is also evidence at sites such as Redwick for multiple transgressive and regressive episodes during the late Holocene (Allen and Haslett, 2007). The dates are asynchronous, however, indicating that the marine inundation was time transgressive.

Table 2.20: Radiocarbon dates from the late Holocene for the south coast of Wales (Godwin and Willis, 1961, 1964; Heyworth and Kidson, 1982; Smith and Morgan, 1989; Allen and Haslett, 2007).

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Material dated	Altitude (m oD)	RSL (m)	Index or limiting point
Margam, Glamorgan	Godwin and Willis 1961, Heyworth and Kidson 1982	Q265	3913-3397	Transgressive overlap	-2.40	-9.20	Index
Llanwern Monmouth	Godwin and Willis 1964, Heyworth and Kidson 1982	Q691	3056-2382	Transgressive overlap	2.98	-3.11	Limiting
Goldcliff	Smith and Morgan 1989	CAR644	3550-3160	Peat	1.56	-4.53	Limiting
Goldcliff	Smith and Morgan 1989	CAR645	3888-3509	Peat	1.47	-4.82	Limiting
Redwick	Allen and Haslett 2007	Beta144021	3316-2948	Thin silty peat	4.58	-2.35	Index
Redwick	Allen and Haslett 2007	Beta144020	3826-3452	Regressive contact	3.08	-3.01	Index

Edwards (2006) produced three late Holocene limiting points from the Loughor Estuary, Wales. Compact, grey clay was overlain by well humified peat, followed by iron-stained silts and clays (Edwards, 2006). A monospecific foraminifera assemblage of *J.macrescens* was recovered, which indicates the sediment was deposited approximately 40cm above MHWST (Edwards, 2006). Five SLIPs were produced in total with three from the late Holocene indicating very little change in sea level from 2760-2381 cal. BP (CAMS41309, 2.65m OD), to 762-563 cal. BP (CAMS41307, 4.10m OD). Furthermore, sediment compaction had a significant effect and was calculated to increase the rate of sea-level rise by between 0.7 to 1.0mm a^{-1} (Edwards, 2006).

Table 2.21: Radiocarbon dates for the late Holocene at the Loughor Estuary, south Wales (Edwards, 2006)

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Material dated	Altitude (m oD)	RSL (m)	Index or limiting point	Environment
Loughor Estuary, Glamorgan	Edwards 2006	CAMS41307	762-563		4.10	-0.08	Index	RWL defined by author
Loughor Estuary, Glamorgan	Edwards 2006	CAMS41310	2099-1824		3.30	-0.83	Index	RWL defined by author
Loughor Estuary, Glamorgan	Edwards 2006	CAMS41309	2760-2381		2.65	-1.47	Index	RWL defined by author

Hewlett and Birnie (1996) investigated three sites in the inner Severn estuary, which revealed a single peat layer up to 4.5m thick, overlain by grey silty clay. Diatoms in the clay were very sparse, but pollen evidence indicated a decline in alder and rise in sedge prior to clay deposition (Hewlett and Birnie, 1996). Particle size analysis indicated that the clay was deposited in still water, not fluvial conditions. The environment of the clay was interpreted as the 'perimarine' zone defined by Hageman (1969) in the Netherlands, where sedimentation is influenced directly by rising sea levels but is devoid of marine or brackish water indicators (Hewlett and Birnie, 1996). The regressive contact from peat to clay was dated at three sites. The altitudes of the regressive contact decrease inland as the dates become younger: 3445-3183 cal. BP at 5.1m OD (Be80696), 2701-2183 cal. BP at 4.65m OD (Be81686), and 2697-2156 cal. BP at 3.9m OD at (Be80693) (Hewlett and Birnie, 1996). This was interpreted as a series of inland bays which infilled with estuarine sediment as sea levels rose, leading to a time transgressive contact (Hewlett and Birnie, 1996). The data from the inner Severn estuary confirm there was a marine inundation during the late Holocene which reached inland (Hewlett and Birnie, 1996). However, only one transgressive episode was recorded, indicating later episodes of marine inundation seen at sites such as Redwick may be caused by local factors such as barrier development (Allen and Haslett, 2007).

Table 2.22: Radiocarbon dates for the late Holocene for the inner Severn Estuary, Hewlett and Birnie, 1996

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Material dated	Altitude (m DD)	RSL (m)	Index or limiting point	Notes
Longney	Hewlett and Birnie 1996	Be80693	2697-2156	Regressive contact	3.90	-3.48	Limiting	weak marine signal
Elmore	Hewlett and Birnie 1996	Be81686	2701-2183	Regressive contact	4.65	-2.73	Limiting	weak marine signal
Slimbridge	Hewlett and Birnie 1996	Be80696	3445-3183	Regressive contact	5.10	-2.28	Limiting	weak marine signal

There is also evidence for a late Holocene marine transgression in the Bristol Bay area. For example, at Nyland Hill, Haslett *et al.,* (1998) found a marine clay layer, between 2.42 to 4.52m OD, dated to between 3640-3330, 3715-3460 and 3725-3675 cal. BP. Haslett *et al.* (1998) concluded that relative sea level rose at a rate of 0.41-0.82mm a^{-1}. Similar dates have been obtained from other sites in the Bristol Bay area, such as the Steart Peninsula (3900-3720 cal. BP, -4m OD, Elliott, 2015) and the Gordano Valley (3841-3637cal. BP, 2.78m OD, Hill *et al.,* 2007). Furthermore, dates published in the synthesis by Heyworth and Kidson (1982) from the sites of Kenn Pier (St3294, St3295), Stolford (NPL146) and Avonmouth (St3257) occur between 4081-3485 cal. BP and 3554-3003 cal. BP (see Table 2.22) (Heyworth and Kidson, 1982). However, details of stratigraphy were not published, so the relationship to sea level is unclear.

Table 2.23: Radiocarbon dates from the late Holocene for the Bristol Bay (Heyworth and Kidson, 1982; Haslett *et al.,* 1998; Elliott, 2015; Hill *et al.,* 2007)

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Material dated	Altitude (m oD)	RSL (m)	Index or limiting point	Notes
Nyland Hill	Haslett *et al.,* 1998	Be101742	3687-3265		3.81	-1.99	Index	
Nyland Hill	Haslett *et al.,* 1998	Be101740	3822-3456		3.88	-1.92	Index	
Nyland Hill	Haslett *et al.,* 1998	Be101741	3825-3468		2.42	-3.38	Index	possibly eroded, no biostrat
Gordano Valley	Hill *et al.,* 2007	Beta189681	3841-3637		2.78	-5.17	Index	RWL defined by authors
Steart Peninsula	Elliott 2015		3900-3720		-4.00			
Avonmouth Bristol Channel	Heyworth and Kidson 1982	IGS27	3564-3038		3.96	-2.74	Index	
Stolford Somerset	Heyworth and Kidson 1982	NPL146	3967-3482		0.60	-5.17	Index	
Kenn Pier Somerset	Heyworth and Kidson 1982	IGS36	4084-3561		2.69	-3.61	Index	
Weston Super Mere Somerset	Heyworth and Kidson 1982	IGS40	4347-3715		1.60	-4.20	Index	
Kingston Seymour	Heyworth and Kidson 1982	I4846	4404-3723		1.30	-5.00	Index	

At Godney Moor on the Somerset levels, multiple transects revealed blue clay across the site, which was overlain by a complex stratigraphy of peat and lake deposits. Godwin (1941) analysed the basal blue clay at Shapwick station, which was found between 0 and 1.2m OD and contained marine-brackish diatoms and foraminifera. Houseley *et al.,* (1988) analysed a core from Batch Farm in Godney Moor and found a basal marine-estuarine clay which was overlain by *Phragmites* peat, dated to 3680-3378 cal. BP (3.1m OD, Q 2457). The radiocarbon date was taken from the middle of the peat layer with no evidence of sea level and was therefore rejected as a SLIP. The *Phragmites* peat was overlain by *Cladium mariscus* sedge fen peat and fen carr deposits (Houseley *et al.,* 1988). A fine detritus lake mud indicated rising water levels prior to the deposition of blue-grey clay which Godwin (1955) found to contain brackish-estuarine diatoms and foraminifera. Prior to dating evidence, Godwin (1941) interpreted this as Romano British. Evidence of a 'Romano British transgression' has also been described by Kidson and Heyworth (1973) who found 'floodplain clay' which they interpreted as a Romano British rise in sea level, although the clay was not dated. It has been questioned whether there was a transgression in Romano British times, due to lack of evidence (Hawkins, 1971; Murray and Hawkins, 1976; Smith and Morgan, 1998). At Godney Moor, the estuarine clay was dated to between 2759-2469 cal. BP (3.8m OD, Q2459) to 3155-2855 cal. BP (3.6m OD, Q2458), which is earlier than the Romano British period (Houseley, 1988). This second late Holocene marine transgression is similar to the sequences on the south coast of Wales at sites such as Redwick (Allen and Haslett, 2007) and may indicate local factors such as barrier development causing multiple transgressive episodes.

Houseley *et al.* (2007) investigated another site on Godney Moor which revealed *Sphagnum* peat between 2.16-0.5m OD overlain by grey clay between 3.8m OD to 1.66m OD. The grey clay was interpreted as the Upper Wentlooge formation, which graded upwards into humified peat between 4.14m OD to 3.8m OD (Houseley *et al.,* 2007). The regressive boundary at 3.8m OD was dated to 2764-2488 cal. BP (GU3247) and 2837-2491 cal. BP (GU3246) and contained pollen such as Poaceae, Cyperaceae and saltmarsh species (Houseley *et al.,* 2007).

Table 2.24 Radiocarbon dates from the late Holocene at Godney Moor, Somerset (Houseley 1988; Houseley *et al.,* 2007).,

Site Name	Reference	Lab code	Calibrated Date (cal. BP)	Material dated	Altitude (m oD)	RSL (m)	Index or limiting point	Notes
Godney Island Somerset Levels	Housley 1988	Q2459	2759-2469	Top of estuarine clay	3.80	-1.97	Index	
Godney Island Somerset Levels	Housley 1988	Q2458	3155-2855	Base of estuarine clay	3.60	-2.38	Index	
Godney Island Somerset Levels	Housley 1988	Q2457	3680 - 3378	Phragmites-woody peat	3.10	-	-	Rejected: no marine influence
Godney Moor, Somerset Levels	Housley *et al.,* 2007	GU3247	2764-2488	Regressive boundary	3.80	-2.38	Index	Pollen indicators
Godney Moor, Somerset Levels	Housley *et al.,* 2007	GU3246	2837-2491	Regressive boundary	3.80	-2.38	Index	Pollen indicator

Overlying the Wentlooge formation there are three terraced deposits beneath modern saltmarshes, known as the Rumney, Arwe and Northwick formations (Allen, 1987). Much of the estuary wetland was reclaimed in the Roman period and the modern saltmarsh is confined to the area seawards of the coastal defences. The Rumney, Awre and Northwick terraces formed due to coastal expansion after the embankment took place (Allen, 1987).

The Rumney formation was the highest terrace, with the surface marsh at 0.5m above MHWST, named after the cliff exposures on the Rumney Great Wharf (ST 22 77) and Penham Moors (ST 2276, 2277) (Allen, 1987). It began to form from the early medieval to early modern period (Allen, 1987). The Rumney formation forms a 2.7m sequence deposited on an erosional surface cut down almost to the Wentlooge peat, with poorly laminated pink silty clays containing *Scrobularia plana* and *Macoma balthica.* This clay was overlain by well laminated pink silty clays containing siliceous marine sponge spicules and open-marine to estuarine diatoms and foraminifera, which graded upward into dark grey silty-sandy clay at the marsh surface (Allen, 1987). At the site of Oldbury on Severn Haslett *et al.* (2001) recovered the top 1m of the Rumney formation, with foraminifera indicating a sequence of a progressively higher saltmarsh surface. Geochemical analysis of zinc in the saltmarsh sediments was used to date the core, and revealed multiple, uniform zinc peaks which have been dated by French (1996) for the Severn Estuary. Haslett *et al.* (2001) produced dates between AD 1937+7 (6.85-6.9m OD) and AD 1958±4 (7.00-7.05m OD). A SLIP was produced from a sample of saltmarsh peat which was dated to 1951±4AD at 6.95-6.9m OD. Biostratigraphic evidence suggested the peat formed at MHWST, which is at 7m OD. The SLIP indicated that sea level rose from 1950AD to the marsh surface, dated to 1995, at a rate of 1.04-2.50mm a^{-1} (Haslett *et al.,* 2001).

The Arwe formation forms the second terrace 0.4m below the Rumney formation and is named after cliff exposures near the village of Arwe (SO 72 08, 71 08, 71 09). It has been hypothesised to have formed in the nineteenth century (Allen, 1987). The sequence consists of up to 2m of well laminated silty sands containing sponge spicules, as well as estuarine diatoms and foraminifera (Allen, 1987). The lowest terrace is the Northwick formation, 0.25m below the Arwe formation. It is the youngest lithostratigraphic element, named after cliff exposures at Northwick Warth (ST 55 87, 55 88) and began formation in the late nineteenth and early twentieth centuries (Allen, 1987). It consists of 1.2m of grey well-laminated silty-sandy clays containing plant roots, siliceous marine sponge spicules and mixed open-marine to estuarine diatoms and foraminifera (Allen, 1987). It is unclear how these three terraces relate to sea level, but it has been suggested that they represent shoreline instability and channel movement over the last 2000 years (Allen, 1987).

2.5 Pollen analysis and vegetation changes in Wales

This section will outline the evidence for late Quaternary palynological research, primarily for northwest Wales. The review will focus on the nature and timing of vegetation and coastal changes, as well as the differences between upland and lowland pollen sites. See Figure 2.23 for sites mentioned in the text.

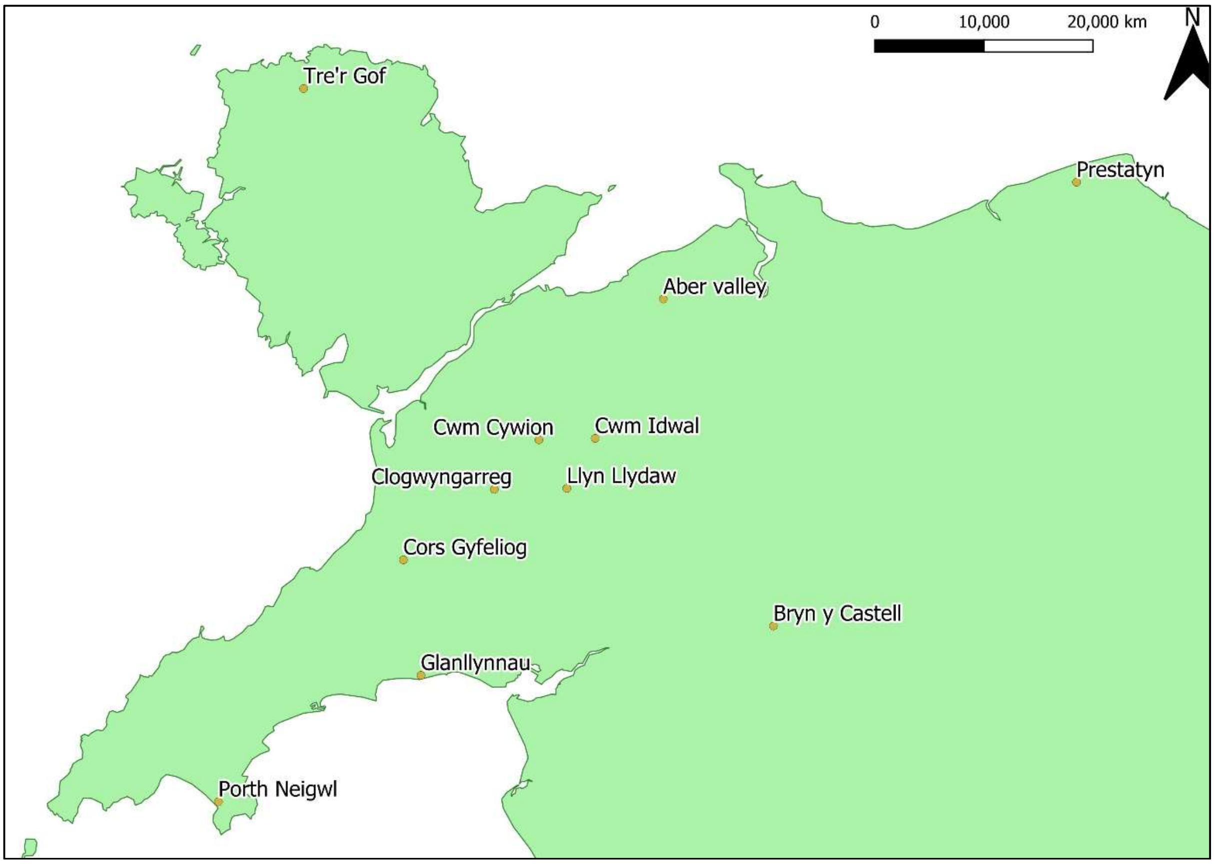

Figure 2.23: Map indicating pollen sites in northwest Wales, with sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252)

2.5.1 Post Late Glacial Maximum (18000 to 10500 cal. BP).

During the post glacial period in Wales the pollen record was characterised by *Rumex* (dock and sorrel) grasslands, with *Juniperus-Betula* (juniper-birch) scrub gradually expanding. By the onset of the Holocene pioneer *Betula* and *Pinus* (pine) woodland had become established. There were considerable variations in north Wales, as small-scale vegetation communities were often influenced by local environmental factors.

Two kettle-holes at Glanllynnau, Caernarvonshire (Figure 2.23), revealed a post glacial landscape dominated by Poaceae (grass), Cyperaceae (sedge), some *Artemisia* (mugworts) as well as *Pinus* (Figure 2.24) (Simpkins, 1974). By 12825-11300 cal. BP (GaK-1603) the grasslands were colonised by *Rumex acetosa* (sorrel)*, Betula*, *Salix* (willow) and *Juniperus*. *Salix* and *Juniperus* gave way to local fen communities, indicated by the appearance of herbs such as *Filipendula* (meadowsweet) (Simpkins, 1974). There was an increase in *Juniperus* dated to 10750-10250 cal. BP (GaK-1602). The onset of the Holocene led to the expansion of *Betula*, up to 70% Total Land Pollen (TLP), with the appearance of *Ulmus* (elm), *Quercus* (oak), *Alnus* (alder) and *Corylus* (hazel) as *Juniperus* declined (Simpkins, 1974).

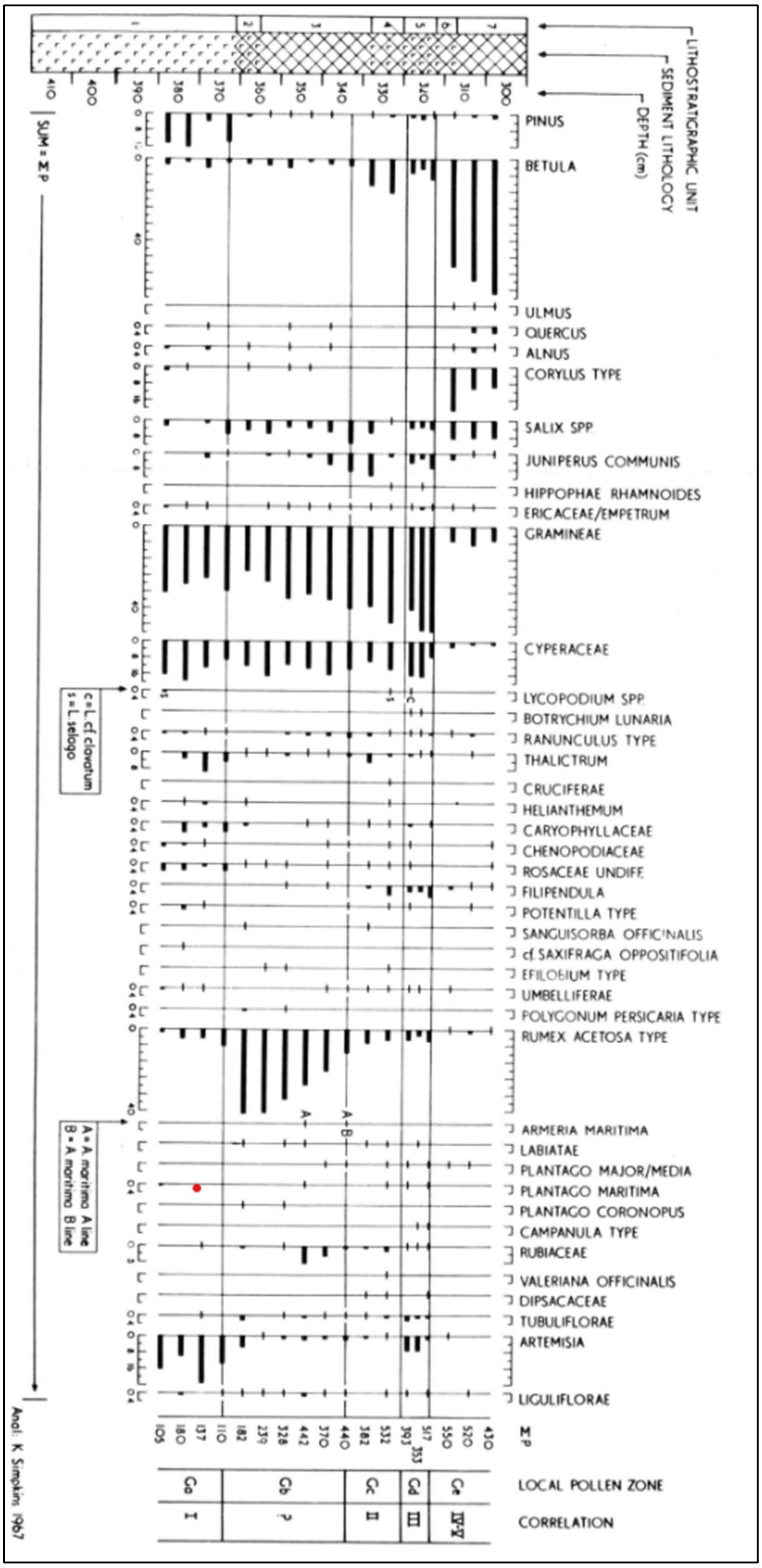

Figure 2.24: Glanllynnau Marsh percentage pollen diagram reproduced from Simpkins (1974). Horizontal dash indicates values <1%.

Ince (1983) found a similar pattern at two upland sites in Snowdonia, Llyn Llydaw and Cwm Cywion. At both sites, immediately after deglaciation, the uplands were colonised by herb-rich grassland communities dominated by *Rumex*. *Betula-Juniperus* scrub expanded into the grasslands, followed by *Betula-Corylus* woodland, finally giving way to a mixed woodland with *Betula, Pinus, Quercus, Ulmus* and *Alnus* (Ince, 1983).

At Cwm Idwal, North Wales, early arctic-alpine vegetation communities were present in the post glacial period alongside *Rumex*-Cyperaceae grasslands (Tipping, 1993). *Juniperus* expanded, until the appearance of *Betula* signalled the decline of both *Juniperus* and Poaceae. *Corylus*-type pollen colonised the catchment, preceding the expansion of a mixed woodland of *Ulmus* and *Quercus* by the Holocene (Tipping, 1993).

The lowland site of Clogwyngarreg, Snowdonia, was ice free by 17342-15767 cal. BP, as *Rumex* grasslands and open habitat herbs became established. The grassland was colonised by *Juniperus-Betula* scrub, followed by a resurgence of grassland during the late glacial interstadial (Ince, 1983). Environmental conditions deteriorated by 13085-12470 cal. BP leading to the spread of open ground, disturbed pollen taxa. By the onset of the Holocene, climate amelioration led to the establishment of early woodlands throughout the catchment (Ince, 1983).

At the lowland, lacustrine site at Tre'r Gof, northwest Wales, the pollen profile was indicative of a late glacial, regional tundra, or park tundra environment. Grassland colonised the area with *Rumex, Filipendula* and *Artemisia,* as well as many cold-climate taxa such as *Populus* (poplar)*, Juniperus*, *Salix herbacea* (dwarf willow) and tundra herbs such as *Thalictrum alpinum* (alpine meadow-rue) (Botterill, 1988). *Juniperus* continued as pioneer *Betula* and *Corylus* woodland spread. Tundra environments declined with the spread of woodland fringe herb species and herb taxa such as *Filipendula* indicating mire and wet woodland around the lake (Botterill, 1988).

In northwest Wales, there was evidence for open habitats through the post glacial period, particularly *Rumex* grasslands, which were colonised by *Betula-Juniperus* scrub. Climate amelioration at the onset of the Holocene was marked by the spread of pioneer woodland. However, there are very few dated pollen profiles for the post glacial period in Wales, so the timings of changes are unclear.

2.5.2 Early Holocene (10500 to 7000 cal. BP).

The early Holocene in north and west Wales was characterised by *Betula* and *Pinus* pioneer forests giving way to the expansion of mixed *Quercus-Ulmus* woodland. *Corylus* and *Alnus* colonised later, although the timing of changes varied across Wales.

At Bryn y Castell, Snowdonia, the early Holocene was characterised by *Betula, Corylus* and *Pinus* colonising meadow grassland (Figure 2.25) (Mighall and Chambers, 1995). *Juniperus* was conspicuously absent at Bryn-y-Castell, which is unusual as *Juniperus* was ubiquitous in late Devensian and early Flandrian pollen diagrams across north Wales (Handa and Moore, 1976; Hibburt and Swinstur, 1976; Ince, 1983). *Juniperus* declined in north Wales between 9700 and 9500 BP (Watkins, 1991), therefore Bryn-y-Castell either lacked suitable environments, or the pollen profile began after 9700 BP (Mighall and Chambers, 1995). At Bryn-y-Castell the *Betula-Corylus-Pinus* woodland continued to develop through the early Holocene, alongside the rapid expansion of *Quercus* and *Ulmus*. *Quercus* was established in Wales by approximately 9000 BP, but *Pinus* and *Ulmus* did not reach their national limits until 8000 BP or later (Mighall and Chambers, 1995).

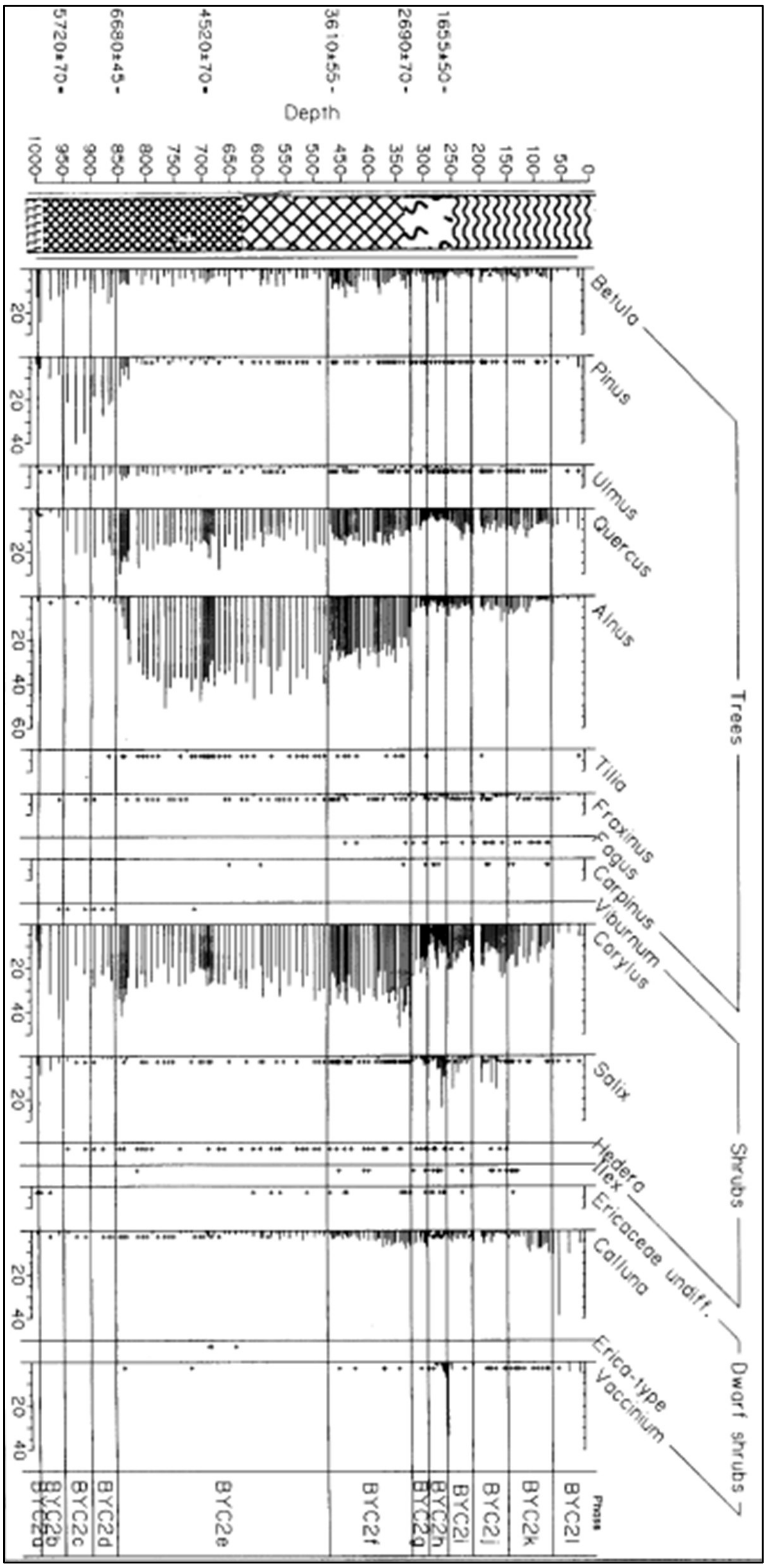

Figure 2.25: Percentage pollen diagram from Bryn y Castell showing trees, shrubs, and dwarf shrubs, (from Mighall and Chambers 1995).

The timing of vegetation changes varied at each site due to the local environment, such as pollen dispersal speed and mechanisms (Birks, 1989). *Betula* was present in Wales before 9500 BP (Birks, 1989), while *Corylus* was established between 9500 and 8800 BP (Hibburt and Swinstur, 1976). For example, at Cwm Cywion the arrival of *Corylus* was dated to 9893- 8728 cal. BP (Ince, 1983). It has been argued that there may have been *Corylus* refugia in central Wales, which may explain the early expansion of *Corylus* at Cors Gyfelog and Tre'r Gof (Botterill, 1988).

The expansion of *Alnus* across Britain was erratic, controlled by local conditions (Bennet and Birks 1990, Birks 1989, Chambers and Elliott 1989). *Alnus* prefers a damp habitat, such as swamp, fen carr, riverbanks, lake shores and valley bottoms (Smith, 1984; Bennett and Birks, 1990). Chambers and Price (1985) found *Alnus* in North Wales at 8465 BP at Moel y Gerddi which is earlier than the rest of the British Isles and argued that there was a possible *Alnus* glacial refugia in Ardudwy. However, there is considerable debate about the early spread of *Alnus* in north-west Wales (Tallantire, 1993). Birks (1989) suggested *Alnus* would first colonise sites at an elevation up to 400m by 7000-6000 BP (Mighall and Chambers, 1995). The early establishment of *Alnus* can also be seen at other sites, such as at Cors Gyfelog, where a rapid expansion of *Alnus* was dated to 8641-8400 cal. BP (Botterill, 1988). At the Graeanog ridge, *Alnus* was abundant by c. 7350 BP (Chambers 1998) and at Nant ffrancon the appearance of *Alnus* was dated to 9885- 9020 cal. BP, becoming widespread by 7933-7571 cal. BP (Hibburt and Swinstur, 1976). At Melynllyn, a mountain tarn in Snowdonia, the appearance of *Alnus* was dated to 8518-7869 cal. BP (SRR639) (Watkins 1978), while at Tregaron Bog, Dyfedd, *Alnus* was first seen at 8334-7622 cal. BP and expanded by 8160-7571 cal. BP (Hibburt and Swinstur, 1976). At Bryn-y-Castell *Alnus* was not widespread until 7656- 7432 cal. BP, which is similar to other sites in Wales (Mighall and Chambers, 1995).

The cause of the early spread of *Alnus* in Wales during the early Holocene has been widely debated, with explanations including Mesolithic human activity, natural fires, beavers, or the creation of suitable habitats through floodplain development (Huntley and Birks, 1983; Smith, 1984; Chambers and Price, 1985; Brown, 1988; Chambers and Elliott, 1989; Bennett and Birks 1990; Edwards, 1990; Tallantire, 1993; Mighall and Chambers, 1995).

The early Holocene in Wales was characterised by the recolonisation of tree species after the glacial period, initially as a mixed woodland of *Quercus* and *Ulmus*. The spread of *Alnus* and *Corylus* followed, however the timing has been widely debated, with some evidence of *Corylus* and *Alnus* earlier in the Holocene compared to the rest of the UK.

2.5.3 Mid Holocene (7500 to 5000 cal. BP).

The mid Holocene marked a period of dramatic coastal change triggered by rapid sea-level rise. Submerged forests are found along the Welsh coastline, dated to between 9000-5500 BP (Bell *et al.*, 2007). At Goldcliff, Severn estuary, the submerged forest, consisting predominantly of *Quercus* was dated to 7682-7576 cal. BP (OxA-12359) (Bell *et al.*, 2007). Marine inundation drowned the *Quercus* trees, with pollen indicating the development of a saltmarsh and reed swamp, and foraminifera providing evidence for sporadic marine inundation throughout the profile at Goldcliff (Bell *et al.,* 2007). Alongside the reed swamp, *Corlyus* woodland developed with *Pteridium* (ferns), *Quercus* and *Ulmus*. Saltmarsh developed by 7667-7513 cal. BP (OxA-12358), indicated by a rapid expansion in Chenopodiaceae (goosefoot) pollen (Bell *et al.*, 2007).

By 6632-6410 cal. BP (OxA-13934) at Goldcliff, estuarine silts were deposited in response to a further sea-level rise (Bell *et al.*, 2007). As the rate of sea-level rise decelerated, saltmarsh/estuarine pollen such as Chenopodiaceae and *Aster*-type (Michaelmas daises) expanded, indicating the development of first saltmarsh and then reed swamp (Bell *et al.,* 2007). By 5800 BP (OxA-12355) a rapid increase in *Salix* and *Alnus* pollen, alongside a decline in reed swamp species, indicated that the wetland was inundated by carr-woodland. The presence of herb pollen, including *Galium* (bedstraw) and *Filipendula*, indicated the presence of tall herb fen. By 5899-5745 cal. BP (OxA-12355) *Alnus* had declined (Bell *et al.,* 2007; Smith and Morgan, 1989; Caseldine, 2000). Sea-level rise across the Severn Estuary caused the coastline to move inland (e.g., Scaife and Long, 1995; Jennings *et al.*, 1998), which led to a rise in the water table across the wetland at Goldcliff, and further drowning of the forests (Bell *et al.*, 2007).

At the site of Bryn-y-Castell, Snowdonia, the mid Holocene pollen profile reflected a mixed deciduous woodland surrounding an infilling lake basin fringed by *Alnus* carr (Mighall and Chambers, 1995). *Ulmus* fell, likely representing the elm decline, seen across northwest Europe between 5500-4800 BP, caused by competitive exclusion, climate change, human activity, or disease (Ten Hove, 1968; Clark and Edwards, 2004). In Wales, evidence for the elm decline was widespread, and at Bryn-y-Castell the disappearance of *Ulmus* was accompanied by a small drop in other trees, the appearance of *Plantago lanceolata* (ribwort plantain) and large charcoal values suggest human activity around 5000 BP (Mighall and Chambers, 1995). Similar evidence of land use was seen across Britain, with the elm decline often correlated with the Mesolithic/Neolithic transition and the onset of agriculture around 5000 BP (Mighall and Chambers, 1995).

There was circumstantial evidence for human impact at Bryn-y-Castell before the elm decline (Mighall and Chambers, 1995). There is no clear evidence for agriculture before the elm decline in Wales (Groenman-van Waateringe, 1993; Edwards and Hirons, 1984; O'Connell, 1987). At Bryn-y-Castell, a possible cereal-type pollen was found in pollen zone BYC2e, accompanied by anthropogenic indicator herbs such as *Rumex* spp., Chenopodiaceae and *Artemisia* (Mighall and Chambers, 1995). Cereal pollen identification is often problematic, however, and it is likely that the single cereal grain at Bryn-y-Castell represents the wild-grass group of Andersen (1979), so the presence of pre-elm decline cultivation is highly unlikely. It is possible that there was Mesolithic ecological disturbance, which could also be seen through circumstantial pollen and charcoal evidence at other sites such as Moel y Gerddi (Chambers and Price 1985, Chambers *et al.,* 1988a) and in the Berwyn mountains (Bostock, 1980).

There is evidence for Mesolithic exploitation of the environment across Wales, for example at upland sites such as Waun-fignen-felen (Smith and Cloutman, 1988) as well as at Banc Wernwgan, Mynydd Du (Caseldine, 2013a), Clogwyngarreg (Grant 2012a) and Ffridd y Bwlch (Grant, 2012b; Caseldine *et al.,* 1990; Caseldine, 2017). At the site of Prestatyn, North Wales, Mesolithic shell middens were studied with pollen profiles that spanned before, during and immediately after the middens were in use (Bell *et al.*, 2007). The basal estuarine clay contained saltmarsh pollen, such as Chenopodiaceae and *Plantago maritima* (sea plantain) with a local *Quercus-Corylus* woodland. *Pinus* was likely overrepresented due to its good preservation and buoyancy, which causes artificially high concentrations of *Pinus* pollen in estuarine sediments (Hopkins, 1950; Long *et al.,* 1999). Through the blue-grey clay there was a continuation of *Pinus* and Chenopodiaceae pollen, which reflected continuation of estuarine or marine conditions. Pollen preservation was poor in the clay, however, due to calcareous inwash, or rapid sediment accumulation (Bell *et al.*, 2007). The transition from the blue grey to humic clay led to increased pollen concentrations, with Poaceae and *Typha latifola* (bullrush) replacing the saltmarsh pollen. From 5470 cal. BP (NHRA-3) *Alnus* encroached onto reed swamp to became dominant locally. This was followed by a rapid reduction in *Ulmus* pollen, reflecting the elm decline, which led to openings in the woodland canopy with the appearance of *Hedera helix* (ivy), *P. lanceolata* and other herbs (Bell *et al.*, 2007). The fluctuations in forest cover and corresponding spikes in herb species indicated possible woodland clearances. By approximately 3500 BP there was an open landscape as the woodland declined, with dominant *Carex* (sedge) and *Pteridium* spores and some indications of cultivation herbs, implying human presence in the landscape (Bell *et al.*, 2007).

At Porth Neigwl, Llŷn Peninsula, a Mesolithic and early Neolithic intertidal peat surface was dated to 7661-7432 cal. BP (SUERC 27554); 7655-7434 cal. BP (SUERC 27554) and 5744-5594 cal. BP (SUERC-27548). *Pinus-Corylus* woodland was present at the base of the profile, alongside low levels of Poaceae and *Alnus* (Caseldine *et al.*, 2016). *Pinus* and Poaceae declined, replaced with a *Quercus-Ulmus* woodland, followed by a sharp rise in *Alnus* indicating the spread of carr woodland. *Alnus* dominated the rest of the profile, until *Alnus* and *Corylus* dropped, replaced by Poaceae, and herbs such as *Filipendula*, *P. lanceolata, Urtica* (nettles), *Solanum nigrum* (black nightshade) and *Rumex*, which are often associated with disturbance and anthropogenic activity (Smith *et al.*, 2016).

Through the Holocene the site of Tre'r Gof was lacustrine until approximately 4000 BP (Botterill, 1988). During the mid Holocene, arboreal pollen was dominant, and wet meadow gave way to *Alnus* carr, and some sparse woodland edge taxa as well as low Cyperaceae and sporadic *Filipendula* (Botterill, 1988). Between 8373-8026 cal. BP and 7247-6794 cal. BP (GrN-15522) there was a major expansion of *Alnus* (Botterill, 1988). By 4406-3984 cal. BP (GrN-15521) there was nearly full woodland cover, with warm climate taxa such as *Tilia* (lime) and *Ilex* (holly). Some human activity was indicated by fluctuations in forest cover, and cultivation indicator herbs, such as *P. lanceolata* and *Rumex* were present throughout, with three large grass grains present, likely the reed swamp species *Glyceria* (sweetgrass). There was also an expansion of reed swamp and developing fen indicated by *Filipendula* and Cyperaceae (Botterill, 1988). There was a similar pattern at Cors Gyf, Llŷn peninsula. By 8176-7871 cal. BP (SRR 3325) there was complete woodland cover and high *Alnus* values. At 5288-4871 cal. BP (SRR 3324) there was a succession to mixed oak woodland with *Tilia*, *Ilex* and *Hedera* reflecting warm conditions (Botterill, 1988).

The mid Holocene in Wales marked a period of rapid sea-level rise and coastal change. However, the pollen evidence for anthropogenic impact was varied and largely short-lived, with very little permanent woodland clearance.

2.5.4 Late Holocene (4000 BP to Present).

The late Holocene in Wales was marked by anthropogenic impact on the landscape, with woodland clearance and agriculture transforming the Welsh landscape. Continuing rising sea level transformed the coastline, moving further inland.

By the end of the Bronze Age at Bryn-y-Castell large scale woodland clearance produced an open landscape dominated by mire vegetation, which persisted to the present day (Mighall and

Chambers, 1995). Neolithic and early Bronze Age arable and pastoral farming was suggested by temporary woodland clearances, reflected in a decline in tree pollen and the expansion of Poaceae along with anthropogenic indicator herbs *Rumex* sp., *P. lanceolata* and cereal type pollen (Mighall and Chambers, 1995). A permanent decline in woodland, affecting all tree species (*Alnus*, *Quercus*, *Corylus*, *Betula*), was dated to 2992-2711 cal. BP (GrN-17583); 2932-2750 cal. BP (GrN-17580) and 2992-2725 cal. BP (HAR-6107) (Mighall and Chambers, 1995). Archaeological evidence of local anthropogenic activity was limited to around 2700 BP, however a climate discontinuity across northwest Europe between 2800-2200 BP led to cooler and wetter conditions, potentially limiting the spread of trees (Wendland and Bryson, 1974; Barber, 1982; Blackford and Chambers, 1991). During the occupation of the Iron Age hillfort at Bryn-y-Castell there was no significant impact on the local woodlands with only small-scale clearance evident from the pollen record. After the occupation of the hillfort (c.1900 BP) there was no archaeological evidence of local settlement, however the pollen profile suggested agriculture continued throughout the historic period, with *P. lanceolata*, *Rumex* sp. and cereal pollen present (Mighall and Chambers, 1995). An increase in agriculture during the medieval period may have been caused by improving climate and increasing population pressure, leading to the spread of upland farming and settlement in north Wales (Chambers *et al.*, 1988a; Caseldine, 1990), with a transhumance farming system adopted until the enclosure of the early nineteenth century (Mighall and Chambers, 1995). Transhumance pastoral agriculture could have been present at Bryn-y-Castell, with species such as *Pteridium*, *Narthecium* (asphodel), *Erica*-type (heather), *Empetrum* (crowberry) and *Potentilla*-type indicating pastoral farming was prevalent (Mighall and Chambers, 1995).

At Tre'r Gof, after 4406-3984 cal. BP (GrN-15521) tree cover declined, particularly *Betula*, *Pinus* and *Ulmus* (Botterill, 1988). The sharp decline in woodland correlated with onset of the Bronze Age, however taphonomy must be taken into account as the lake became infilled, and the subsequent change in the pollen catchment area would affect the pollen profile. However, increased loss on ignition and the expansion of Poaceae, and arable or ruderal herbs and cereal pollen implied deforestation for agriculture (Botterill, 1988).

At Cors Gyfeliog, Llŷn Peninusla, during the late Holocene the lake also infilled, with a corresponding return to fen conditions indicated by the presence of *Filipendula* as well as other wetland taxa, such and *Thalictrum* (meadow-rues), *Pedicularis* (lousewort), Caryophyllaceae (carnation) and *Valeriana* (valerian). Woodland clearance could be seen by the decline in tree pollen alongside the expansion of diverse herbs such as Cyperaceae and *Plantago* sp. (plantain) (Botterill, 1988). The onset of the Bronze Age was marked by the recolonisation of *Alnus*, although regional woodland was unchanged. By the end of the Bronze Age, widespread agriculture was evident, with significant fluctuations in local *Alnus*, large-scale woodland clearance and the expansion of cereal pollen as well as arable and ruderal herbs. After the Roman period, the catchment was open with pastoral and arable herbs dominant, including both cereal and Cannabaceae (cannabis) pollen (Botterill, 1988).

In the Aber valley on the north coast, woodland persisted into the late Holocene. For example, at site Av08, dated to between 3690-3467 cal. BP (BETA-246693) and 897-673 cal. BP (BETA-246691), *Alnus-Betula* woodland was present with *Corylus*, and open ground taxa such as Poaceae and *Calluna vulagris* (heather) (Woodbridge *et al.*, 2014) (Figure 2.26). As *Alnus* declined, *Betula* and *Quercus* expanded alongside more diverse herbs. Woodland clearance was marked by a decline in tree pollen and subsequent expansion of Poaceae and diverse herbs, in particularly *Potentilla* (cinquefoils), *Lactuceae* (cichorieae), *P. lanceolata* and Asteraceae (daisy), with consistent, low levels of microcharcoal. (Woodbridge *et al.*, 2014).

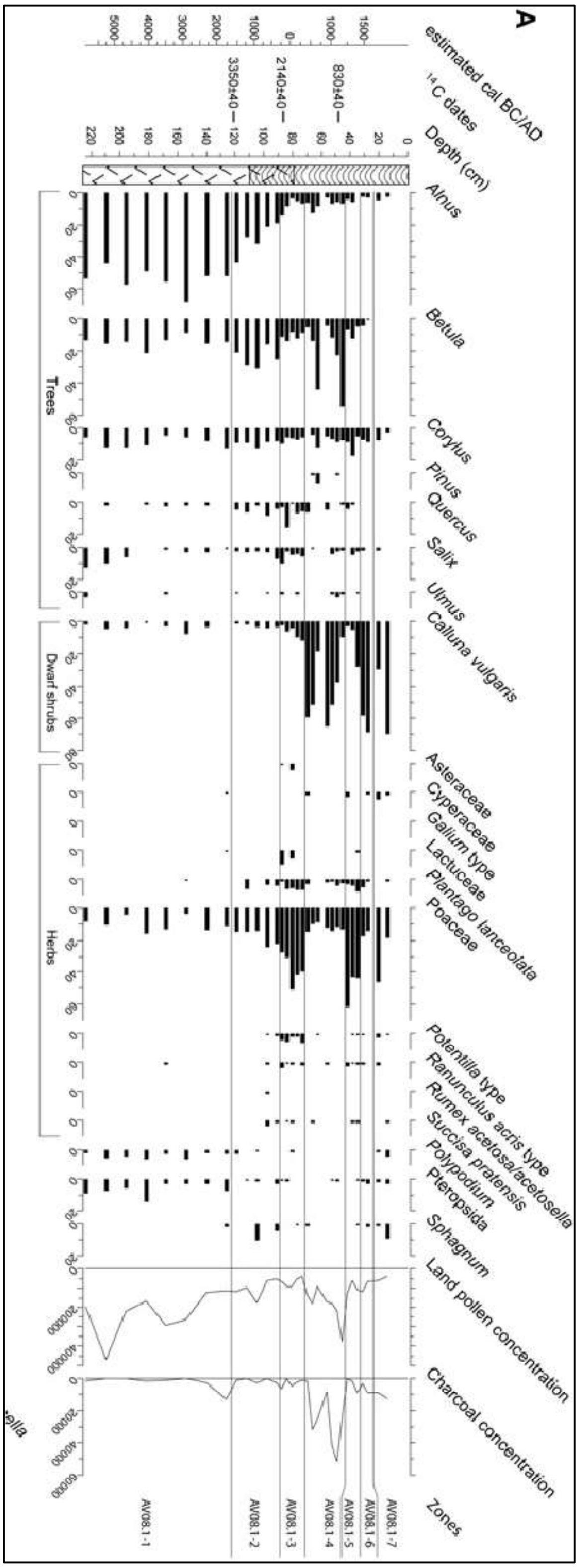

Figure 2.26: Percentage based (% TLP) pollen diagrams from Aber Valley core AV08·1, Aber valley, showing selected taxa. All taxa not grouped in trees, shrubs and herbs are outside the TLP sum (reproduced from Woodbridge *et al.,* 2014)

At Goldcliff, Severn estuary, the reed swamp gave way to raised mire following sea-level rise, dated to 5595- 5315 cal. BP (CAR-651), 5287- 4840 cal. BP (CAR-773) and 5443- 4884 cal. BP (Swan-223) (Smith and Morgan, 1989; Bell *et al.*, 2007; Caseldine, 2000). Poaceae, *Aster*-type and Chenopodiaceae pollen were dominant, indicating high saltmarsh on the wetland. Tall herb fen was indicated by the presence of *Filipendula* and *Lotus*-type pollen. An expansion of *C. vulgaris* and *Sphagnum* (peat moss) was seen, as well as *Drosera rotundiflora* (round-leaf sundew) indicative of raised mire (Bell *et al.*, 2007), which continued until sea-level rise submerged the wetland, dated to 3482-3165 cal. BP (CAR-644) and 2848-2428 (CAR-1438) (Smith and Morgan, 1989; Caseldine, 2000).

The late Holocene marks the onset of large-scale anthropogenic impact on the Welsh landscape. Woodland clearance and continued coastal change led to a dramatic transformation, although the timing and pattern of differences varied across Wales.

2.5.5. Summary: Pollen and vegetation changes in Wales

Following the LGM, Wales was largely an open landscape dominated by dock grasslands and juniper scrub. Pioneer birch-pine woodland developed, which was later colonised by oak-elm woodland. The spread of alder and hazel occurred later, although the timing varied across Wales. Some anthropogenic activity was evident from the Mesolithic period although it was varied and short-lived. Pollen evidence for large-scale anthropogenic impact became common in Wales during the later Holocene, with woodland clearance and coastal change leading to a transformation of the landscape, although timing varied considerably.

2.7 Conclusions

There are many factors which influence sea-level change on both a local and regional scale, including eustasy, isostasy, tectonics, and sediment compaction. Taphonomic processes such as sediment compaction have a significant effect on the records of sea-level change. In the UK, the Devensian glaciation has had the principal impact on sea-level change during the Holocene.

During the Devensian the British and Irish ice sheet developed over Scotland, Ireland, Wales and the Lake District. An ice stream was established in the Irish sea, which met the Welsh ice cap in the Cheshire-Shropshire Lowlands, in Anglesey and the Llŷn Peninsula, as well as the west coast of Wales. The uplands of Snowdonia were covered in cold-based ice, which drained to the west, leading to glacial over-deepened valleys such as the Tal-y-llyn, Dysynni and Mawddach. The ice limit reached to the south coast of Wales, but the glacial history is less well understood due to a lack of geomorphological evidence.

During deglaciation, the Irish Sea ice stream retreated leading to many glacial lakes forming across the west coast, such as Late Teifi as well as small lakes in the Dysynni and Tonfannau valleys. Multiple advance and re-advance episodes are recorded in areas such as the Llŷn peninsula, where local factors led to complex interactions as the Welsh ice cap decoupled from the Irish Sea ice stream. The BIIS decoupled from the Welsh ice cap which drained through valleys to the west and south.

During the Loch Lomond Stadial, glacial conditions returned, and ice sheets formed in areas such as Scotland. There is evidence for cirque glaciation in the uplands of Snowdonia which is attributed to the Loch Lomond glaciation. The chronology of glaciation is unclear, however, due to a lack of dating evidence

The British and Irish ice sheet was relatively small compared to larger ice sheets in Europe, but the unequal north-south distribution led to significant isostatic rebound. Areas of Scotland, where the ice was thickest, continue to rebound while the south coast of the UK sinks. GIA models are essential to predict future sea level change; however, many models predict a mid Holocene high stand in north Wales, which is not seen in the evidence. More recent GIA models such as Shennan *et al.,* (2018) do not predict a high stand which provides a better fit for the available data.

After deglaciation, global sea level rose dramatically due to the influx of meltwater. There is very little pre-Holocene sea level data available for Wales, but evidence from Menai Straits and the south coast indicate sea level rose between 2-3mm a^{-1}. During the early Holocene in Wales, sea level continued to rise rapidly up to 10mm a^{-1}. There are very little data from the west coast, however, and more data is needed.

Following the LGM, the pollen history of Wales shows an open landscape dominated by dock grasslands and juniper scrub. Pioneer birch-pine woodland developed, which was later colonised by oak-elm woodland. The spread of alder and hazel occurred later, although the timing varied across Wales.

During the mid Holocene, the rate of sea-level rise slowed, as the influence of meltwater receded. Across Wales there is a change from marine clay to terrestrial peat development as the rate of sedimentation overtook sea-level rise, leading to coastal expansion. There appears to be a slightly higher rate of sea-level rise on the south coast, with 1-2mm a^{-1} compared to 0.3mma $^{-1}$ on the west coast. This could be due to differential uplift, however more data are needed to confirm this pattern, particularly from the west coast where data are limited. During the mid Holocene of Wales, the submerged forests developed which can be seen at low tide across areas of Wales.

During the late Holocene there is evidence for a marine transgression across south Wales and the Bristol Bay area, which was time transgressive, between 4404-3723 and 762-563 cal. BP. There is no evidence of sea-level change on the north coast during the late Holocene. On the west coast, limiting points from fresh-brackish peat indicate continuing rising sea levels, however there is no evidence of a second marine transgression during the late Holocene. More data are needed from the west coast to confirm if there was a second marine transgression similar to the south coast. Pollen evidence for large-scale anthropogenic impact became common in Wales during the late Holocene, with woodland clearance and coastal change leading to a transformation of the landscape, although timing varied considerably.

Chapter 3: Methodology and techniques of analysis

This chapter outlines the methods and techniques used in this research in order to reconstruct sea-level and environmental changes. The chapter is divided into field and laboratory methods and the statistical techniques used for analyses. The project's research aims are as follows:

1. To reconstruct the Holocene sea-level changes in the Dysynni Valley.
2. To recreate the environmental and vegetation changes in the Dysynni Valley in order to fully understand past changes in the coastal environment.
3. To examine the impacts that the changing coastal environment had on human settlement in the area.

Fieldwork, including gouge survey, will establish the sediments and identify appropriate sampling sites through the valley. Biostratigraphic analyses including loss on ignition, particle size and microfossil analyses will establish the environmental change through each sample core. Specifically, diatom and foraminifera analyses will establish the salinity and marine influence, while pollen will establish coastal and vegetation changes. Radiocarbon dates will be used to establish chronology. The results of the diatom and foraminifera analyses will be used to apply transfer functions and create sea-level index points which will establish the sea-level change in the Dysynni Valley. The results of the vegetation, coastal and sea-level change will be compared to the known archaeological record in order to analyse the reactions of past populations and inform on future work.

3.1 Fieldwork

The research aims to establish both the vertical and horizontal marine inundation of the Dysynni Valley in order to fully reconstruct past changes in the coastal environment. Understanding the changing coastal configuration is vital to our interpretation of human-environment interactions within the Dysynni Valley. Suitable sampling sites were chosen with the potential to preserve sea level records across the Dysynni Valley, from the coast to the 10m elevation at the base of Craig yr Aderyn, as a marine transgression would not have exceeded 10m during the Holocene (Lambeck *et al.,* 2014). Site selection was initially guided by analysis of published data, including archaeological assessments (e.g. Smith, 2004), British Geological Survey (BGS) reports and borehole records (e.g. Pratt *et al.,* 1995; Leng and Pratt, 1987) and Ordnance Survey (OS) maps. Fieldwork and detailed litho-stratigraphical analysis were undertaken in transects across each potential site with a gouge corer (5cm wide, 1m long) in order to determine the nature of stratigraphy. Sediments were described in the field using the Tröels-Smith system (1955). Each borehole location was surveyed using differential Global Navigation Satellite System (GNSS) with a horizontal and vertical positional accuracy of <0.01m.

At each site, the location with the most complete sedimentary sequence was sampled. A sample core was obtained using a Russian corer (Jowsey, 1966) (5cm wide, 50cm long) from overlapping boreholes to provide a full stratigraphic sequence. The core was then placed in PVC pipe and wrapped in cling film and foil and stored in the cold store (<5°C).

3.2 Laboratory analysis

The cores were taken to the laboratory for further analysis. The sediments were described, and subsamples taken for loss on ignition, particle size, diatom and foraminifera analyses. At Penllyn, samples were also taken for pollen analysis.

3.2.1 Sediment Description

Sediment description followed the Tröels-Smith system (1955). Each stratigraphic horizon was analysed for darkness (*nigror*), degree of stratification (*stratificatio*), elasticity (*elasticitas*), degree of dryness (*siccitas*), colour, structure (*structura*), the sharpness of the lower boundary (*limes*), humicity (*humositas*) and composition.

Each characteristic was attributed a number between 0-4, where 0 indicates absence and 4 indicates the characteristic is fully present. For example, in terms of dryness, 0 represents a saturated sample while 4 reflects a desiccated layer. Composition was determined based on texture and visible components within the samples, with a ratio out of 4 indicating approximate percentages. The presence of minor components was indicated by a + or a ++ (Table 3.25). The sharpness of the lower boundary refers to the width of transition (Table 3.26). The determinations are subjective and cannot be compared to work produced by other authors, however, it provides an indication of transitions within the core.

Table 3.25: Sediment composition acronyms and definitions (Tröels- Smith, 1955).

Code	Component Type	Definition
As	*Argilla steatoides*	Clay – mineral particles, <0.002mm
Dg	*Detritus granosus*	Small woody and herbaceous fragments and unidentifiable fossils or fragments, measuring >0.1mm and < 2mm in length
Dh	*Detritus herbosus*	Herbaceous plant fragments >2mm in length
Sh	*Substantia humosa*	Humous substance – completely disintegrated or decomposed organic substances/humic acids[24]. Black and structureless) *i.e.* fully humified peat
Th	*Turfa bryophytica*	Peat composed of moss, with macroscopic structure (not fully humified, may be fresh/unhumified).

Table 3.26: The numerical score and specific depth for attribute 8 *limes* (Tröels-Smith, 1955).

Lim. 0:	-	Boundary zone 10mm or more
Lim. 1	diffusas	Boundary zone 2-10mm
Lim. 2	conspicuus	Boundary zone 1-2mm
Lim. 3	manifestus	Boundary zone 0.5-1mm
Lim. 4	acutus	Boundary zone less than 0.5mm

[24] Alkali soluble organic compounds that are components of humus, the major organic fraction of soil and peat.

3.2.2 Loss on ignition

Loss on Ignition is used to estimate the organic and carbonate content of sediment (Beaudoin, 2003). Organic matter begins combusting at 200°C and is completely burned to ash at 550°C (Santisteban *et al.,* 2004). Carbonate minerals are destroyed at higher temperatures, for example calcite combusts at 800-850°C and dolomite at 700-750°C, so temperatures of 950°C are used to estimate the inorganic carbonate content (Santisteban *et al.,* 2004). Loss on ignition is widely used as it is easy, inexpensive and produces reasonably reliable results (Beaudoin, 2003; Heiri *et al.,* 2001; Stein, 1984).

Care must be taken as there are many sources of error that can affect the results (Heiri *et al.,* 2001; Mook and Hoskin; 1982; Stein; 1984). It is important to be consistent in the method used as the sample size, the exposure time and the position of samples in the furnace will influence the results (Heiri *et al.,* 2001). Loss on ignition at 550°C does not directly measure the organic carbon content, but includes other organic matter, so the organic carbon content makes up approximately half the weight lost at 550°C (Dean, 1974; Stein, 1982). At 950°C the mass loss results from the amount of carbon dioxide lost, and therefore does not differentiate between carbonate minerals. The weight loss must therefore be converted to a carbonate percentage (Stein, 1984). The majority of carbonate minerals break down at temperatures over 650°C but there is a degree of carbonate combustion which overlaps with the breakdown of organic carbon (Neumann, 1967; Boyle, 2004), such as siderite, magnesite, rhodochrosite and dolomite, which combust between 425-520°C (Heiri *et al.,* 2001; Santisteban *et al.,* 2004). It is therefore important to use consistent temperatures to ensure reliable results are produced.

One of the most important sources of error in loss on ignition is the loss of structural water from clay (Heiri *et al.,* 2001; Mook and Hoskin, 1982; Stein, 1984). Lattice water in clay minerals begins to combust at 400°C and the water is not completely removed until 1000°C (Beaudoin, 2003; Dean, 1974; Stein, 1983), which can lead to up to a 5% loss of mass (Santisteban *et al.,* 2004). This is particularly evident in samples low in carbonates but high in clay, such as estuarine sediments (Stein, 1984). Although it is possible to use lower temperatures to avoid loss of water, this results in incomplete combustion of organic matter (Beaudoin *et al.,* 2004; Boyle, 2004). As clay-rich estuarine sediments are susceptible to loss of water at high temperatures, it reinforces the importance of using consistent temperatures when firing, and mass loss less than 5% at high temperatures should be treated with caution.

The method used followed the procedures detailed by Heiri *et al.,* (2001), Gale and Hoare (1991) and Stein (1984). Crucibles were weighed and samples of between 2-4g were added and weighed before being placed in a drying oven at 105°C for 12-14 hours. They were then left to cool to room temperature and reweighed. The samples were then placed in a preheated muffle furnace at 550°C for four hours, and then placed in a desiccator to cool to room temperature. The samples were then weighed and the loss on ignition values were calculated for each sample with the following equation (Heiri *et al.,* 2001):

$$LOI_{550} = \frac{DW_{105} - DW_{550}}{DW_{105}} \times 100$$

Where LOI_{550} represents the Loss on Ignition (LOI) for organic matter as a percentage, DW_{105} represents the dry weight of the sample and DW_{550} is the dry weight of the sample after combustion at 550°C (Heiri *et al.,* 2001). The dry weight was calculated by subtracting the weight of the crucible from the combined weight of the crucible and sample.

Finally, the crucibles were placed in a preheated muffle furnace at 950°C for two hours, after which they were then placed in a desiccator and allowed to cool to room temperature. Samples were weighed a final time and the loss on ignition for inorganic carbon was calculated with the following equation:

$$LOI_{950} = \frac{DW_{550} - DW_{950}}{DW_{105}} \times 100$$

LOI_{950} represents the percentage loss on ignition for inorganic carbon. The loss on ignition results were then plotted against sample depth (Heiri *et al.,* 2001)

3.2.3 Particle Size Analysis

Particle size analysis allows the accurate description of sediments in order to aid understanding of the depositional environment (Lowe and Walker, 2014). When assessing microfossil preservation an understanding of grain size and organic content can be helpful (e.g. Horton and Edwards, 2005, 2006; Zong and Horton, 1999).

In the past, the standard method for particle size determination was the sieve-pipette method, where wet sieving was used to determine particle size of the fraction larger than 63µm, and pipette sampling was used for clay and silt fractions (Beuselink *et al.,* 1998). The pipette settling method is based on Stoke's law, which assumes that particles are rigid, smooth spheres, not plate-like, such as clay (Gee and Bauder, 1986). The sample is suspended in a column of water and the sediment concentration is determined by removing pipette samples at specific time intervals, which correlate to the specific particle sizes 32, 16, 8 and 2µm. This produces the mass percentage for the defined size class (Beuselink *et al.,* 1998). However, this process is time consuming, heavily dependent on technique (Syvitski *et al.,* 1991) and a large quantity of sediment is required (Beuselink *et al.,* 1998).

The development of laser diffraction methods allows large quantities of small samples to be analysed efficiently (Blott and Pyre, 2006). Laser diffraction analysis is based on the concept that particles diffract light at certain angles depending on size, with larger particles diffracting at smaller angles (Beuselink *et al.,* 1998). However, particle morphology has a significant impact on the results, with the clay and silt fraction often under-estimated compared to settling methods (Beuselink *et al.,* 1998; Scott-Jackson and Walkington, 2005; Fisher *et al.,* 2017). Direct comparison of the two methods requires calibration (Scott-Jackson and Walkington, 2005; Fisher *et al.,* 2017). It is important to use consistent preparation methods for all samples when using laser diffraction, as inconsistencies can influence the results, for example organic or carbonate particles could cause errors (Fisher *et al.,* 2017).

Laser diffraction was chosen as it is a fast, efficient method for processing small samples. A Malvern Mastersizer 2000 laser granulometer was used. Samples which had more than 80% organic content were not analysed for particle size as there would not be sufficient material to process after the digestion of organic material. Prior to analysis, samples were sieved to 2mm where necessary, as grains larger than 2mm would damage the analyser. The fraction larger than 2mm was weighed as a whole, but only one sample had greater than 2mm fraction (Chapter 5.3.3). Organic content was removed by adding 20ml of H_2O_2, and the initial, violent reaction allowed to finish, before heating on a hotplate until all organics were removed (Allen and Thornley, 2004). A few drops of HCl were added to remove any carbonates and any undigested H_2O_2. Samples were washed in distilled water using a centrifuge until all chemicals were removed. A few millilitres of 0.4% sodium pyrophosphate solution were added to deflocculate the clay, and samples were stirred (Allen and Thornley, 2004).

The laser analyser was cleaned overnight with washing up liquid and between each sample the analyser was washed twice with tap water and twice with deionised water. The analyser was calibrated using a sand standard (0.152-0.422mm) to ensure consistency, and each sample was analysed three times. The results were separated into particle size fractions based on the Wentworth (1922) scale.

3.2.4 Diatom analysis

Diatoms are single-celled algae that exist in all aquatic environments from freshwater to marine. They have two valves formed of silica, which allows them to be preserved well in sediments. Variations in the shape and patterning of valves allows taxonomic identification to species level (Battarbee *et al.,* 2001; Round *et al.,* 1990; Stoermer and Smol, 1990).

Diatom assemblages are influenced by salinity, the nature of the substrate, the food supply, pH and temperature. Diatoms inhabit the entire range from hyper saline to freshwater, and salinity has been widely recognised as an important forcing factor in diatom communities (Denys and de Wolfe, 1999; Lowe and Walker, 2015; Zong and Horton, 1999). The halobian system classifies diatom species based on salinity tolerances: polyhalobian[25], mesohalobian[26], oligohalobian-halophilous or oligohalobian-indifferent[27], and halophobous[28] (Hustedt, 1953, 1957; Palmer and Abbott, 1986; Ridgway *et al.,* 2000).

Diatoms can also be classified based on their life history. Planktonic and tychoplanktonic taxa exist in the water column, while epiphytic species live attached to vascular plants and algae. Epipelic species live in rocky areas and epipsammic diatoms are attached to sand grains. Aerophile taxa exist in the subaerial and supratidal zone (Cooper, 1999; Vos and de Wolf, 1993). Through the identification and reconstruction of the salinity tolerances and life history of the diatom species that make up an assemblage, the environment of deposition can be reconstructed.

It is possible for diatoms to be transported from one environment and deposited within a different ecological zone: for example, when freshwater diatoms are transported by fluvial action and deposited in a saltmarsh environment. Transported diatoms within an assemblage can cause errors of interpretation and can give misleading ecological information. In order to distinguish between *in situ* (autochthonous) and transported (allochthonous) species a number of methods have been proposed (e.g., Beyens, 1982; Vos and De Wolf, 1993); however, there is no single accepted method. In estuarine environments as there are other considerations such as taphonomy, which is affected by both mechanical and chemical processes, leading to a preservation bias. The fragile species fragment and dissolve while the more robust and highly silicified species survive, thereby skewing the data (Cooper; 1999; Sherrod *et al.,* 1989). It is therefore important to consider potential taphonomic processes, as well as the transported components when analysing diatom assemblages (Sawai, 2001; Sherrod *et al.,* 1989; Vos and de Wolf, 1993).

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[25] Marine, salinity greater than 30‰
[26] Brackish, salinity between 2-30‰
[27] salinity up to 2‰
[28] freshwater

Diatom preparation followed the standard procedure by Palmer and Abbott (1986). Initially, subsamples for diatom analysis were taken at low resolution in order to assess preservation and establish overall changes in salinity using the halobian classification system. The results of the initial diatom analysis determined higher resolution sampling where changes in salinity occurred.

Subsamples of $1cm^3$ were weighed into suitable beakers. Approximately 20ml of H_2O_2 was added to remove any organics. Samples were left for an hour until the initial reaction had taken place, then the samples were placed on a hotplate at 90°C for 4-6 hours or left off the hotplate overnight. Once the reaction had completed, a few drops of HCl was added to remove carbonates and any unreacted H_2O_2. Samples were transferred to tubes, centrifuged, and decanted. At least 4 washes were undertaken with distilled water, with 1% ammonia added in the last wash to remove clay from the samples. In order to make slides, the samples were diluted and pipetted onto cover slips and left to dry overnight where they would not be disturbed. Once dried, glass slides were placed on a hotplate at 130°C, and the coverslips applied with Naphrax.

Due to poor preservation in some cores, a second preparation method was used to rule out breakages caused by the preparation method. Centrifuging has been reported to cause breakages to diatom valves (Abrantes *et al.,* 2005; Battarbee *et al.,* 2001), and prolonged digestion is time consuming, and could potentially cause differential dissolution (Vermeulen *et al.,* 2012; Hart, 1957; Hendey, 1964; Flower, 1993). The method suggested by Abrantes *et al.,* (2005) was adapted, which did not use centrifuging. Samples of $1cm^3$ were weighed into beakers and left soaking overnight in sodium pyrophosphate to disaggregate the samples. Organics were removed by adding 20ml of H_2O_2, which resulted in a faster digestion process due to disaggregation. Instead of centrifuging, samples were left to settle overnight, and a vacuum pump was used to remove excess liquid. The samples were resuspended in distilled water and left to settle again, and this was repeated at least 4 times. This preparation method removed any potential sources of damage to the diatom valves. Having processed the samples with the new method, preservation remained low, with broken diatom valves. Every precaution was taken to ensure the preparation method did not affect preservation, and it is possible that taphonomic factors influenced the poor preservation.

Diatoms were counted using a microscope with phase contrast at x1000 magnification under immersion oil. 250 valves were counted where possible, using systematic traverses. Two hundred and fifty valves represent a reasonable proportion of the entire sample prepared given time limitations, and is widely used (Best 2016, Long *et al.,* 1998; Ridgway *et al.,* 2000; Watcham *et al.,* 2013). Taxa were identified using Hartley *et al.,* (1996), Hendey (1964) and Krammer and Lange-Bertalot (2010). Percentage diagrams were created in Tilia (Grimm 2004). Zones were stablished using Constrained Incremental Sum of Squares (CONISS) cluster analysis using Tilia v.2.0.2 and TG View v.2.0.2 (Grimm 2004).

3.2.5 Foraminifera

Foraminifera are single celled organisms that live in marine environments, from deep oceans up to the shallow intertidal zone and are abundantly preserved in saltmarsh sediments making them useful for analysis of sea-level change (Gehrels, 2007; Kemp *et al.,* 2012). The distribution of foraminifera within a saltmarsh is determined by the duration and frequency of intertidal exposure (Horton *et al.,* 1999; Horton and Edwards, 2006; Kemp *et al.,* 2012; Scott and Medioli, 1978). Foraminifera assemblages can therefore be used to determine the saltmarsh elevation relative to the tidal frame and establish the indicative meaning (Horton and Edwards, 2006; Scott and Medioli, 1978). For example, there is a pronounced decline of foraminifera at the upper limit of marine influence, which can be used to calculate the highest high water to ±5cm (Horton *et al.,* 1999).

Surveys of modern foraminifera distribution in the UK have found that foraminifera can be divided into agglutinated species[29] that live on the vegetated marsh, and highly diverse calcareous species which occupy the mudflats and sand flats of the intertidal zone (Horton and Edwards, 2006; Horton *et al.,* 1999). The highest and transitional zones of a saltmarsh produce the most accurate

[29] Foraminifera where the tests formed from foreign particles stuck together with a variety of cements.

reconstructions of tidal level (Edwards and Horton, 2000; Gehrels *et al.,* 2001; Kemp *et al.,* 2012). However, there is considerable variation between sites, so regional trends may not be applicable (Horton and Edwards, 2006).

Foraminifera preparation followed the standard method outlined by Scott and Medioli (1980). Two to five cm^3 of wet sediment was wet sieved through 500 and 63 µm mesh sieves, with the 63µm fraction used for analysis. Floating organic material was removed by allowing the sample to settle in water for c. 1 minute then decanted. The whole sample was counted where possible, but where high numbers of foraminifera tests prevented analysis, a wet splitter was used to divide the sample into 8 equal portions. Samples were counted wet on a spiral tray using a microscope with between 32x and 120x magnification. Foraminifera were picked and placed on a pre-glued slide to assist identification (Scott and Medioli, 1980). Ideally 200 tests were counted, which is suitable as foraminifera form low diversity assemblages (Gehrels *et al.,* 2001), but in many samples counts were too low to achieve this. Foraminifera were identified following Murray (1971, 2000) and Horton and Edwards (2006). Concentration calculations were made in Tilia (Grimm, 2004) based on the total volume counted, and diagrams were created in Tilia v.2.0.2 and TG View v.2.0.2 (Grimm, 2004).

3.2.6 Pollen

Pollen analysis is an essential tool when investigating Quaternary palaeoenvironments, particularly when reconstructing human impact on the landscape (Birks and Birks, 1980; Berglund, 2003; Galliard, 2013; Leipe *et al.,* 2019). The presence of saltmarsh and coastal vegetation can inform sea-level reconstructions alongside diatom and foraminifera analyses.

Pollen preparation involves digestion in very strong acids and bases, and commonly hydrofluoric acid (HF) is used to remove silicates (Faegri and Inversen, 1989; Faegri *et al.,* 2000; Moore *et al.,* 1991), however there are significant safety risks when using HF (Leipe *et al.,* 2019; Campbell *et al.,* 2016). Digestion with strong acids can also lead to corrosion of the pollen, and HF is unsuitable for clay rich sediment (Allen *et al.,* 2009; Caffrey and Horn, 2013; Campbell *et al.,* 2016; Leipe *et al.,* 2019).

An alternative to HF is dense media separation, which takes advantage of the different densities of pollen exines and silica to remove pollen from the sediment matrix (Campbell *et al.,* 2016; Leipe *et al.,* 2019). Dense media separation has been commonly used for clay rich or silty sediment (Allen *et al.,* 2009; Caffrey and Horn, 2013; Campbell *et al.,* 2016; Fletcher and Hughes, 2016), however it can also be used for organic sediment (Campbell *et al.,* 2016; Nakagawa *et al.,* 1998). Dense media separation is more expensive and time consuming than HF, however it is safer and produces much cleaner slides making counting much more efficient (Allen *et al.,* 2009; Campbell *et al.,* 2016; Forster and Flenley, 1993; Leipe *et al.,* 2019; Nakagawa *et al.,* 1998; Philips and Playford, 1984).

In the past, toxic dense media have been used such as Thoulets solution (a mix of mercury iodide (Hg_2I_2), potassium iodide (KI) and water), bromoform ($CHBr_3$); a solution of bromoform and acetone (C_3H_6) or ethanol (C_2H_6O); zinc bromide ($ZnBr_2$); zinc chloride ($ZnCl_2$); caesium chloride (CsCl) and a solution of potassium iodide and cadmium iodide (CdI_2) (Caffrey and Horn, 2013; Campbell *et al.,* 2016; Leipe *et al.,* 2019; Nakagawa *et al.,* 1998; van Geel *et al.,* 2011; Wood *et al.,* 1996). Since the 1980s non-toxic, water-soluble salts have become available, such as sodium polytungstate (SPT), lithium metatungstate (LMT) and lithium heteropolytungstate (LST) (Caffrey and Horn, 2013; Campbell *et al.,* 2016; Leipe *et al.,* 2019; Munsterman and Kersholt, 1996).

Dense media separation has been compared to HF and found to be a suitable alternative (Björck *et al.,* 1978; Caffrey and Horn, 2013; Campbell *et al.,* 2016; Leipe *et al.,* 2019; Nakagawa *et al.,* 1998). Nakagawa *et al.,* (1998) found no statistical difference between HF and dense media separation, which was confirmed for SPT by Campbell *et al.,* (2016). Leipe *et al.,* (2019) found that SPT was statistically similar to HF but there were slight differences in pollen taxa when LST was used.

Samples for pollen analysis were taken at 8cm intervals from the core at Penllyn. Samples of $1cm^3$ were disaggregated before digestion in KOH to remove humic acid, followed by acetolysis to remove cellulose. Heavy liquid separation was undertaken using LST Fastfloat at a density of $1.95g\ cm^{-3}$. A known quantity of *Lycopodium* exotic markers was added in tablet form.

A minimum of 250 pollen grains were counted per slide at x400 magnification. Identifications were made using Moore *et al.*'s pollen key (Moore *et al.,* 1991) and the University of Sheffield's reference collection with nomenclature following Bennett (1994).

Percentages were calculated in Tilia v.2.0.2 (Grimm 2004). The following sum was used to calculate concentration (Stockmarr, 1971):

$$\left(\frac{1}{volume\ of\ sample}\right) \times \left(\left(\frac{grains\ counted}{number\ of\ Lycopodium}\right) \times (grains\ per\ tablet\ \times number\ of\ tablets)\right)$$

Pollen diagrams were created in Tilia and TG View v.2.0.2 (Grimm 2004) and CONISS analysis was used to establish zones.

3.2.7 Radiocarbon dating

A robust chronology is essential for environmental archaeology and sea-level reconstructions. Radiocarbon dating is widely used and measures the rate of decay of the unstable ^{14}C isotope. Radiometric dating measures beta decay, which requires a large sample size, while accelerated mass spectrometry (AMS) dating measures the ^{14}C concentration through high-energy accelerators. AMS dating therefore requires a much smaller sample size and allows more high-resolution dating (Muller, 1977; Törnqvist *et al.,* 2015).

Radiocarbon dates from bulk peat samples contain organic carbon from a number of different sources, such as fluvial input of reworked carbon and bioturbation (Törnqvist *et al.,* 1992, 2015; Nilsson *et al.,* 2001). Nilsson *et al.,* (2001) found that bulk peat dates varied between 365 and 1000 years compared to *Sphagnum* macrofossils, and Hu (2010) calculated an error or ±100 years to account for uncertainties in bulk peat dates.

Macrofossils are widely accepted as a reliable source of dating in peat as they are thought to be unaffected by the same problems as bulk samples (Björck *et al.,* 1998; Nilsson *et al.,* 2001; Törnqvist *et al.,* 1992, 2015). Care must be taken when selecting macrofossils for dating, as transported macrofossils are a significant problem in fluvial environments, for example Turney *et al.,* (2000) found that dates from *Carex* seeds were inverted and likely redeposited. Fragile terrestrial macrofossils, such as *Salix* leaves, are unlikely to survive long transport and should provide more accurate dates (Björck *et al.,* 1998; Hatte and Jull, 2007; Turney *et al.,* 2000). Bioturbation, caused by roots, animals or insects, could also affect macrofossils, for example *Phragmites* can grow up to 2m into the sediment (Hatte and Jull, 2007; Howard *et al.,* 2009). Furthermore, it is essential the macrofossils are short-lived, for example the heartwood of a tree may be centuries old by its death (Hatte and Jull, 2007) but a small twig or roundwood, such as a branch, may be suitable, as they represent only a few years of growth (Hatte and Jull, 2007). It is also possible for old wood to be redeposited meaning timber could be a thousand years older than expected (Hatte and Jull, 2007).

There are also factors which affect the ^{14}C content of organisms, such as the reservoir effect, which occurs when waterbodies such as lakes, estuaries or the open ocean exchange carbon with the atmosphere at a slower rate, leading to organisms with a lower $^{14}C/^{12}C$ ratio (Törnqvist *et al.,* 2015). Furthermore, estuaries are influenced by both marine and freshwater inputs, with ^{14}C depleted CO_2 from deep ocean water mixing with atmospheric CO_2, leading to a reduction in ^{14}C (Hatte and Jull, 2007; Törnqvist *et al.,* 2015). In hardwater catchments, fluvial input of $CaCO_3$ can cause incomplete mixing of CO_2 with the atmosphere, leading to a differential $^{14}C/^{12}C$ ratio (Törnqvist *et al.,* 2015). There is also isotopic fractionation, which occurs when carbon is transferred through processes such as photosynthesis or crystallisation, leading to discrimination between lighter or heavier carbon isotopes, which can be corrected based on $^{\delta 13}C$ measurements (Törnqvist *et al.,* 2015). It is important, therefore, to use only well-identified, local, short-lived, terrestrial macrofossils for radiocarbon dating (Björck *et al.,* 1998; Nilsson *et al.,* 2001; Törnqvist *et al.,* 1992, 2015).

Calculations must be made before radiocarbon dates can be used. Underpinning radiocarbon methodology is the assumption that the amount of ^{14}C has remained the same over time, which is not the case and must be corrected (Törnqvist *et al.,* 2015). Radiocarbon dating became widespread before the half-life of carbon was accepted (5730 ^{14}C a, Godwin, 1962), so instead the "Libby half-life" is used (5568 ^{14}C a, Libby 1952), which must be corrected (Törnqvist *et al.,* 2015). Radiocarbon ages do not correlate directly with calendar years, and must be calibrated (De Vries *et al.,* 1958; Suess, 1970; Törnqvist *et al.,* 2015). Calibration curves have been created from tree ring chronologies, ^{235}U-^{90}Th dating (Bard, 1990) and terrestrial macrofossils from varved sediments (Lolter, 1999).

Samples were selected based on litho- and bio-stratigraphic changes in the core in order to produce sea-level index points and produce a robust age-depth model for the core. Care was taken to avoid contamination, including using sterilised equipment. No suitable macrofossils were found, such as moss or leaves. Wood fragments were discounted due to the risk of transportation, and the age of the wood at the time of the death of the tree cannot be determined. *Phragmites* reeds were not used as they can grow up to 2m into sediment. Due to a lack of datable macrofossils, AMS bulk dates from peat were used. Bulk dates may be less accurate as they are a mixture of sediment from different sources but are preferable to non-local or unidentifiable macros of uncertain age ranges. Where an age reversal was evident, both humic acid (Alkali soluble) and humin (acid- and alkali-insoluble) fractions were dated in order to determine the most representative date (Shore *et al.,* 1995; Brock *et al.,* 2011; Marshall *et al.,* 2016). If the humic and humin dates were similar the date was considered accurate, however where the humic acid was younger than the humin, the sample may have been contaminated by inwash of old carbon. Inversely, if the humin was younger, there may have been contamination by modern plant rootlets (Shore *et al.,* 1995; Brock *et al.,* 2011; Marshall *et al.,* 2016). Radiocarbon dates were obtained from DirectAMS (USA) and they were calibrated to cal. years BP using Oxcal v4.4 (Ramsey, 2001, 2005; Ramsey and Lee, 2013) with the IntCal13 atmospheric curve (Reimer *et al.,* 2013).

3.3 Statistical analysis

3.3.1 Sea-level index points

A sea-level index point (SLIP) defines the position of former sea level in space and time, by quantifying litho- and bio-stratigraphic information of a sediment sample in relation to a reference water level (Engelhart and Horton, 2012; Edwards and Horton, 2000; Horton *et al.,* 2013; Shennan, 1982, 1986). A SLIP consists of the age, altitude, tendency and indicative meaning of a sediment sample (Barlow *et al.,* 2013; Edwards and Horton, 2000; Engelhart and Horton, 2012; Horton *et al.,* 2013; Shennan, 1982). The tendency of a SLIP defines whether marine influence is increasing (positive) or decreasing (negative) (Barlow *et al.,* 2013). The indicative meaning of a sample represents the relationship between the environment of deposition and the former tide level, and includes both a vertical range (the indicative range) and a mid-point (the reference water level, RWL) (Edwards and Horton, 2000; Horton *et al.,* 2013; Engelhart and Horton, 2012). Relative sea level (RSL) is calculated by the following equation:

RSL = A-RWL

Where relative sea level is calculated by subtracting the refence water level (RWL) from the altitude of the sample (A) (Engelhart and Horton, 2012; Horton *et al.,* 2013; Shennan *et al.,* 2000). For a modern surface sample, both A and RWL will be 0, therefore RSL is expressed as below present (Engelhart and Horton, 2012).

The errors must be accounted for in calculating a SLIP, which include surveying errors (±0.01 with high precision equipment), measuring sample depth, compaction during coring (±0.01m), the angle of coring (±1% of sediment overburden) and the sample thickness (± half the sample thickness) (Barlow *et al.,* 2013; Engelhart and Horton, 2012; Shennan, 1986; Törnqvist *et al.,* 2008). Sediment compaction must also be considered, as it results in a lowered elevation and underestimation of sea-level change (Edwards, 2006; Engelhart and Horton, 2012; see Chapter 2.2.2.2). There is currently no accepted method for accounting for compaction (Barlow *et al.,* 2013; Brain *et al.,* 2012, 2015, 2016, 2017; Edwards, 2006; Hill *et al.,* 2007; Long *et al.,* 2006, 2010; Massey *et al.,* 2006a, b; Shennan *et al.,* 2000; Streif, 1979; Törnqvist *et al.,* 2008; van de Plassche, 1982). By differentiating between basal SLIPs, which overlie a non-compressible substrate, and intercalated SLIPs it is possible to estimate the effects of compaction (Barlow *et al.,* 2013; Edwards, 2006; Engelhart and Horton, 2012).

Changes in tidal range must also be considered, as changes in sediment supply, land level, coastal geography and land use can lead to changes in tidal regime (Barlow *et al.,* 2013). If the tidal range was greater in the past, the RWL would be greater, leading to an underestimation of RSL change (Barlow *et al.,* 2013; Horton *et al.,* 2013). However, changes in tidal regimes are hard to quantify and models of palaeotides are limited (Austin 1991; Barlow *et al.,* 2013; Gehrels *et al.,* 1995; Hill *et al.,* 2007; Hinton 1996, Shennan *et al.,* 2003, Uehara *et al.,* 2006).

Errors for SLIP are calculated with the following equation:

$$E = \sqrt{(e_1^2 + e_2^2 + e_3^2 \ldots + e_n^2)}$$

Where e_1 to e_n represent the individual error sources for each sample (Barlow *et al.,* 2013; Engelhart and Horton, 2012; Shennan and Horton, 2002; Preuss, 1979).

3.3.2 Transfer functions

A transfer function produces a quantitative reconstruction of relative sea level by expressing the value of an environmental variable (e.g., elevation) as a function of biological data (e.g., diatom or foraminifera abundance) (Birks, 1995, 2010; Barlow *et al.,* 2013; Kemp and Telford, 2015). Transfer functions have been widely used by sea-level researchers in order to expand the sea-level reconstructions possible, from a single depositional horizon provided by a SLIP to a complete record of sea-level change available from the whole core (Barlow *et al.,* 2013; Kemp and Telford, 2015).

Care must be taken, however, to represent the full vertical error produced by the model, as it is possible for a very accurate model to have poor error values (Barlow et al. 2013).

The first step is developing the training set in order to quantify the relationship between the modern taxa and the environment. This is done using regression calculations[30] which produce an optimal elevation and vertical range for each species (Woodroffe and Long, 2009). This relationship is then used to calibrate the biological assemblages from the fossil core in order to infer the past elevation of the sample from the number of each taxon present (Edwards *et al.,* 2004). Table 3.27 details the stages required to construct a transfer function model, and these will be discussed further in the following sections.

Table 3.27: Steps involved in transfer function development.

1	Choosing or developing training sets, comprising species abundance and environmental variables
2	Converting elevation (m OD) to Standardised Water Level Index (SWLI)
3	Choosing model based on unimodal or linear response
4	Assessing model performance, choosing best component and pruning datasets
5	Applying modern analogues to fossil cores in order to evaluate the suitability of the training set
6	Reconstructing environmental conditions from the fossil cores using model

Transfer functions provide a quantitative reconstruction of sea level, with a wider range of information than a sea level index point, which is restricted to stratigraphic boundaries (Barlow *et al.,* 2013). However, there is a limitation. The transfer function method is underpinned by the theory of uniformitarianism, that the present is the key to the past (Hutton, 1795). The relationship between species and the environment may have changed over time, for example, the modern

[30] set of statistical processes for estimating the relationships between a dependent variable

species tolerance of salinity may have changed in the past, making reconstructions inaccurate (Barlow *et al.,* 2013).

3.4.2.1 Choosing the training sets

It was not possible to construct a local training set as part of this research due to constraints of time. The regional UK training sets for foraminifera (Horton and Edwards, 2006) and diatoms (Zong and Horton, 1999) were initially chosen, however there are no sites in Wales in either training set. The foraminifera training set developed by Rushby *et al.,* (2019) for the Maltreath marsh, Anglesey, was included as it is located in Wales, as well as the diatom training set for the Severn Estuary developed by Hill *et al.,* (2007).

First, a modern training set must be established which consists of biological taxa (such as diatoms and foraminifera) and environmental variables (such as elevation or pH) (Wilson and Lamb, 2012; Kemp and Telford, 2015). Modern samples are taken from a marsh at regular intervals along one or more transects perpendicular to the direction of tidal inundation, or alternatively sampling can be determined by vegetation or sedimentary successions (Barlow *et al.,* 2013). The method of sampling varies depending on the proxy being studied, with the top centimetre sampled for foraminifera (Wright *et al.,* 2011) or a few millimetres, removing surface sediment, for diatoms (Hamilton and Shennan, 2005). Environmental variables such as elevation, particle size, organic content and pH are also recorded (Barlow *et al.,* 2013; Kemp and Telford 2015).

Care must be taken, as the sampling strategy may cause bias in the transfer function models. Typically, samples are taken in transects across the marsh in order to build training sets. The regular sampling strategy leads to spatial autocorrelation, which causes nearby sites to falsely resemble each other (Mills *et al.,* 2013). Spatial autocorrelation can cause bias when statistical techniques assume independence of residuals (Avnaim-Katav *et al.,* 2017; Barlow *et al.,* 2013; Telford and Birks, 2009, 2005). This can lead to exaggerated predictive power, with over-estimated accuracy and precision. Spatial autocorrelation can be mitigated through random sampling of the training set, i.e., choosing random samples of the entire dataset (Leorri *et al.,* 2008) or by sampling multiple sites and multiple transects (Mills *et al.,* 2013).

A local training set, which is in close proximity to the fossil core, is likely to have similar environmental conditions, such as hydrology, leading to a more precise transfer function (Woodroffe and Long, 2010; Kemp *et al.,* 2009). Local training sets are not always available due to anthropogenic influence restricting the modern coastal environments represented, particularly when compared to the early Holocene (Wilson and Lamb, 2012).

A regional training set is constructed from multiple sites over a large area, covering a wider range of species and environments. There are different definitions of a regional training set, ranging from one estuary to a whole country (Edwards and Horton, 2000; Horton and Edwards, 2006; Gehrels *et al.,* 2001; Kemp *et al.,* 2009; Leorri *et al.,* 2008; Szkornik *et al.,* 2006; Watcham *et al.,* 2013; Zong and Horton, 1999; Barlow *et al.,* 2013). Using multiple sites from a range of environments increases the predictive power but reduces the precision of the training set (Edwards *et al.,* 2004; Gehrels, 2000; Wilson and Lamb, 2012). Barlow *et al.,* (2013) compared the use of regional and local training sets. They found that although the local training set produced a more precise model, it lacked good modern analogues and so was less reliable compared to the regional training set which had much better analogues.

It is possible to use either linear[31] or unimodal[32] methods when constructing a training set (Barlow *et al.,* 2013). Detrended canonical correspondence analysis (DCCA) is used to establish whether linear or unimodal methods are suitable: if there is a gradient length greater than two standard deviation units, a unimodal response is appropriate (Barlow *et al.,* 2013). Within ecology, all species-environment responses are non-linear with an environmental optimum and range. If only a limited part of the environmental curve is sampled, however, the response can appear linear, and therefore linear methods would be appropriate (Birks, 1995, 2010).

[31] used to predict the value of a variable based on the value of another variable.
[32] In statistics, a unimodal probability distribution is a probability distribution which has a single peak

Research has shown that multiproxy transfer functions produce better results than using a single proxy (Elliot, 2016; Kemp *et al.,* 2009; Horton and Edwards 2000; Gehrels *et al.,* 2001; Best, 2016). Kemp *et al.,* (2009) compared diatom, foraminifera and multiproxy training sets from North Carolina, USA, and found that the multiproxy transfer function had improved predictive power. Gehrels *et al.,* (2001) compared diatom, foraminifera, testate amoebae and multiproxy training sets from the Taf Estuary, Wales, and concluded that the multiproxy training set produced the best results.

3.4.2.2 Development of a transfer function

Differences in the tidal range must be accounted for in order to compare multiple sites. Many standardisation methods have been used in the past (e.g. Gehrels, 1999; Horton *et al.,* 1999). It is common practice to convert the elevation (m OD) to Standardised Water Level Index (SWLI) units using an equation (Kemp and Telford, 2015; Barlow *et al.,* 2013; Horton and Edwards, 2006). The equation uses the upper and lower limits of the tidal frame, most commonly mean higher high water (MHHW) or highest astronomical tide (HAT) and mean tide level (MTL) or mean lower low water (MLLW) (Kemp and Telford, 2015). Ideally, a tide gauge would be available at each site, and it is possible to take field measurements, but if this is not possible the nearest tide gauge is used (UK Hydrographic Office, 2020). There are a number of equations available, but the following equation from Zong and Horton (1999) was chosen as it produced a positive value for all sites, unlike other equations from Horton and Edwards (2006) or Barlow *et al.,* (2013).

$$SWLI = \left(\left(\frac{(Altitude - MTL)}{(MHWST - MTL)}\right) x\ 100\right) + 200$$

In order to combine multiple datasets, it is important to use the same SWLI calculation throughout as well as using the same nomenclature (Barlow *et al.,* 2013). When combining a dataset with two different proxies, if percentages are calculated with both datasets at the same time, it will create false links in the data, for example, an increase in one foraminifera species could cause a false decrease in a diatom species. It is therefore important to calculate the percentage for each proxy separately, so that the total percentage of a dataset with two proxies will equal 200 (Elliott, 2016).

Many researchers prune the training set prior to analysis, removing samples or species with low counts, which are likely to introduce significant bias in species tolerance and optima (Horton *et al.,* 1999, Edwards *et al.,* 2004). There is no accepted threshold, however (Elliott, 2016; Gehrels *et al.,* 2001; Rushby *et al.,* 2019) and some researchers do not remove low count samples, assuming that the spread in the dataset better defines the diversity of the sampled modern environments (Barlow *et al.,* 2013; Zong and Horton, 1999; Hill *et al.,* 2007). Many researchers remove low count species from the training set, especially those that contribute to less than 2% or 3% in at least one sample (Shi 1993, Skzornik *et al.,* 2006; Best, 2016). Care must be taken, however, as many robust marine species, such as *Actinoptychus senarius* or *Coscinodiscus nitidus* will not be abundant in coastal samples (Denys, 1991/2), and removal of these key species will be detrimental to the interpretation of the diatom assemblage as a whole.

It is important to preserve a robust training set in terms of sampling size when combining two different proxies. Elliott (2016) suggests that when one proxy is too low it should be reduced to 0, leaving the other proxy in the sample intact. The alternative is to remove a sample from the training set entirely when one proxy is poorly represented which can result in a very limited training set (Elliott, 2016).

For this research, samples with fewer than 30 foraminifera were removed, the only exception being the low count, monospecific samples of *Jadammina macrescens* which have a well-defined relationship to tide level (Gehrels *et al.,* 2001). Due to preservation, diatom samples with less than 100 valves were removed, rather than the usual 200, as removing samples with under 200 valves resulted in a very limited fossil dataset. Species were removed if they did not occur in at least one sample at 3% or higher and following Elliott (2016) where one proxy was removed the other was left if it was appropriately represented.

There are many different methods available when constructing a transfer function and it is important to select the method that is the most appropriate for the datasets. A unimodal response has been established for the training sets used in this research (Horton and Edwards, 2006; Zong and Horton, 1999; Rushby *et al.,* 2019; Hill *et al.,* 2007). A number of unimodal methods are available for transfer function development, including weighted averaging (WA), weighted averaging partial least squares (WA-PLS) and tolerance downweighted weighted averaging (WA-Tol) (Birks, 1995, 2010; Barlow *et al.,* 2013).

The methods can be separated into classic or inverse regression. Classic methods use regression of the taxa abundances on the environmental variables to estimate a calibration function, which is then reversed to estimate the unknown variable in the fossil sample (Birks, 1995, 2010). Inverse methods estimate the calibration function directly from the modern training set by regressing the environmental variables on the taxa abundance. The environmental variable is then estimated from the fossil sample directly using the modern regression (Birks, 1995, 2010). Inverse regression is most effective when fossil samples are from the centre of the distribution of the modern training set, while classic regression is more effective at the extremes (Ter Braak, 1995; Sundberg 1985; Birks *et al.,* 1990a; Gasse *et al.,* 1995).

Maximum likelihood (ML) regression and calibration model the relationship between taxa abundance and the environmental variables using an environmental response curve and associated errors (Kemp and Telford, 2015; Birks, 1995, 2010). The curve is fitted to the modern training set by non-linear regression producing a model of responses used to calculate the probability that the environmental variable may occur, with the highest probable value the maximum likelihood estimate (Birks, 1995, 2010). It is the most statistically robust, classic regression; however, it is very complex and computer intensive (Kemp and Telford, 2015).

Weighted averaging (WA) regression is a simpler method which produces similar results to maximum likelihood regression (Birks, 1995, 2010). WA considers that each taxon inhabits a unique ecological, unimodal niche with an environmental gradient consisting of the species optima and a tolerance range (ter Braak and Juggins, 1993; Woodroffe and Long 2009; Kemp and Telford, 2015). In WA regression, the most abundant taxa present are used to determine the optimum for the environmental variable reflected. The environmental optimum is calculated from the average of the total range of the environmental variable weighted to the taxon abundance (Birks, 1995, 2010). Due to averages being taken twice, once for regression and again at calibration, 'shrinkage' occurs, leading to a reduction in the inferred value for the environmental variable (Barlow *et al.,* 2013). De-shrinking is a relatively simple process and can be done with classic or inverse methods depending on the part of the gradient analysed, with the inverse method most effective in the centre and the classic method at the extremes (Kemp and Telford, 2015). Weighted averaging is also sensitive to the distribution of sites along an environmental gradient which can cause bias. WA treats every environmental variable separately and it does not consider the residual correlations in the biological data (Birks, 1995, 2010).

In contrast, weighted averaging-partial least squares (WA-PLS) utilises the residual correlations in order to improve the fit between the environmental variables and biological data in the modern training set (Ter Braak and Juggins, 1993; Birks, 1995, 2010). WA-PLS produces a number of components, with the first chosen to maximise the covariance between the weighted averages and the environmental variable. Subsequent components are chosen independently for the same reason and have no direct relationship with the other components. If the first component is used, the model will be equivalent to the WA method, while using subsequent components will enhance performance up to 60% in some datasets (Birks, 1995, 2010; Hill *et al.,* 2007; Kemp and Telford, 2015). Adding an increasing number of components appears to increase the success of the model, but it also enhances statistical complexity, adding noise which masks the relationship between the species and the environmental variable (Elliott, 2016; Barlow *et al.,* 2013). It is important to choose the smallest number of components to produce an acceptable model, using no more than three (Barlow *et al.,* 2013; Kemp and Telford, 2015; Scheder *et al.,* 2019). WA-PLS produces species coefficients and standard errors for each species which are used to choose the optimum model (Woodroffe and Long, 2009; Birks, 1995, 2010).

Some species are more sensitive to changes in sea level and therefore have smaller elevation ranges, and such narrow tolerance species can therefore provide more precise information about changes in elevation (Horton and Edwards, 2000). Another variant of WA regression, known as tolerance downweighted regression (WA-TOL) provides greater weight to species with narrower tolerances (Hill *et al.,* 2007). A number of researchers have compared WA, WA-PLS and WA-TOL methods, and WA and WA-PLS have been found to be more effective than WA-TOL (Massey *et al.,* 2006; Elliott, 2016).

3.4.2.3 Assessing model performance.

It is important to evaluate the model reconstruction using a range of methods, including simple scatter plots, the root mean squared error of prediction (RMSEP) and the coefficient of determination (R^2) (Birks, 1995; Kemp and Telford, 2015; Barlow *et al.,* 2013). It is also important to assess the relationship between the training set and fossil samples using the modern analogue technique (MAT) (Birks, 1995, 2010; Barlow *et al.,* 2013).

The transfer function produces a standard error, or root mean squared error (RMSE) which describes the predictive power of the transfer function and the R^2 value and measures the strength of the relationship between the observed and inferred values. (Birks, 1995, 2010). When based on the transfer function alone, however, RMSE and R^2 will overestimate the performance of the model. Cross validation or split sampling presents a more accurate estimation of any errors (Birks, 1995, 2010).

Split sampling partitions a portion of the training set and uses it to estimate the environmental variable for a separate test set (Birks, 1995; Kemp and Telford, 2015). Split sampling provides a realistic assessment of the RMSEP and R^2 by comparing the observed and predicted values. The RMSEP estimates the predictive ability of the training set but does not give sample specific errors for fossil samples, as the observed environmental variable is not known (Kemp and Telford, 2015). Split sampling requires large training sets, however, so it is not always applicable (Birks, 1995, 2010).

A number of cross validation techniques are available, such as jacknifing, also known as 'leave one out' (ter Braak and Juggins 1993). A reconstruction is repeated, with a single sample removed each time, producing a predicted value (Milker *et al.,* 2017). The error is calculated by subtracting the predicted from the observed value (Birks, 1995, 2010). Jacknifing produces RMSEP (jack) from the sum of all the errors, as well as an $R^2_{(jack)}$ value (Kemp and Telford, 2015).

While the RMSEP evaluates the random or standard error, it is also important to review the systematic errors. The relative bias represents the systematic differences in the predictions, and therefore the mean error (Birks, 1995, 2010). In order to calculate the maximum bias, the sampling interval of the environmental variable is separated into equal parts, and the mean bias of each interval is calculated. The largest value of mean error for an interval is the maximum bias (Juggins and Birks, 2012).

Bootstrapping repeats the model runs in repeated cycles, usually 1000, removing a random subset of the training set each cycle. In order to replicate the size of the original, the removed samples are replaced by samples from the remaining training set to produce a test set (Kemp and Telford 2015). The test set is then used to infer the environmental variable for the samples that were removed (Birks, 1995, 2010).

Bootstrapping produces individual RMSEP error values for fossil samples as well as the modern training set (Milker *et al.,* 2017; Barlow *et al.,* 2013). Bootstrapping also produces an R^2 value as well as an estimated SWLI estimate and associated residual for each training set sample (Elliott, 2016). The residual is the difference between the observed and estimated value. Sample specific errors and residuals can give an indication of where in the intertidal zone the training set is strongest, and whether it is likely to under- or over-estimate past elevations (Elliott, 2016). Bootstrapping is often used due to the improved precision, reduced bias, and the ability to produce individual error values for individual samples (Birks, 1995; Hill, 2006; Kemp *et al.,* 2009; Leorri *et al.,* 2008, 2011).

When using WA or WA-PLS regression, the RMSEP and R^2 values are used to select the appropriate model component out of the many produced. A model is chosen with a high $R^2_{(boot)}$ and low RMSEP. No more than three components should be chosen, with each subsequent component selected if there an improvement of at least 5% RMSEP (Barlow *et al.,* 2013).

Samples with a poor relationship to elevation can be identified using sample specific error values. There is no accepted method, but some researchers remove samples with residuals greater than the standard deviation of the range of SWLI represented in the training set (Edwards *et al.,* 2004; Gehrels *et al.,* 2005; Horton and Edwards, 2006). Woodroffe and Long (2010) removed samples if the absolute residual were a quarter of the total range of the elevation gradient. Removing these outliers can improve the predictive power of the training set (Edwards *et al.,* 2004; Gasse *et al.,* 1995; Jones and Juggins, 1995).

Analyses such as RMSEP and $R^2_{(boot)}$ give an indication of model performance, but they do not determine how realistic the reconstructions may be (Horton and Edwards, 2006). It is important to assess the similarity between the fossil and modern samples, as if they are too dissimilar, the transfer function must extrapolate. In this way, a very precise transfer function may be useless if there are no good modern analogues (Woodroffe and Long, 2009). This is especially true for fossil samples from the distant past, such as the early Holocene, as there is greater chance of changing environments through time (Barlow *et al.,* 2013). A transfer function is most reliable with good modern analogues from the training set (Birks, 1995; Hamilton and Shennan, 2005; Edwards and Horton, 2000; Barlow *et al.,* 2013).

The MAT uses a suitable dissimilarity measure to compare the fossil sample to the modern training set which produces a minimum dissimilarity coefficient (minDC) for each sample (Barlow *et al.,* 2013; Woodroffe, 2006, 2009). A larger value indicates greater dissimilarity (Jackson and Williams, 2004; Barlow *et al.,* 2013). Squared chord distance is a widely used dissimilarity measure, as it maximises the signal to noise ratio when used with percentage data (Birks, 1995; Edwards and Horton; 2000; Overpeck *et al.,* 1985).

It is important to note that a low dissimilarity value does not necessarily represent a good analogue, and there is no fixed rule in determining what defines a good, close or poor analogue (Barlow *et al.,* 2013; Elliott, 2016; Wright *et al.,* 2011). Typically, the range of the dissimilarity in the modern training set is used, with the extreme 5-10% used as a margin for a good analogue (Birks, 1995). Watcham *et al.,* (2013) used the 20th percentile as a cut off between close and poor, with samples below the 5th percentile defined as a good analogue. Kemp *et al.,* (2009) also used the 20th percentile to differentiate between poor and close, however the 10th percentile was used as a cut off between close and good analogues. Many other researchers have followed the definitions produced by Kemp *et al.,* (2009) (Horton and Edwards, 2006; Edwards and Horton, 2000; Shennan *et al.,* 2015).

When using a transfer function, typically the sample specific RMSEP value is used without any additional errors (Barlow *et al.,* 2013). For a microtidal environment, the errors are often underestimated, so other sources of error must be taken into account, such as the angle of coring. It is often unnecessary when considering a single core, but it is essential to determine sources of error when comparing different sites (Barlow *et al.,* 2013) (see Chapter 3.3.1 for in depth discussion of errors).

3.4.2.4 Reconstruction

The transfer function produces an estimate of the palaeomarsh surface elevation (PMSE), also known as the indicative meaning, which represents the surface elevation of the fossil sample at the time of deposition (Barlow *et al.,* 2013; Elliott, 2016; Woodroffe and Long, 2009). Relative sea level (RSL) is then calculated by subtracting the PMSE from the present elevation (Barlow *et al.,* 2013; Gehrels *et al.,* 2001). Each reconstructed value has an associated sample specific error, produced by bootstrapping cross validation (Birks, 1995, 2010; Barlow *et al.,* 2013; Elliott, 2016).

Samples from the upper part of the intertidal zone provide the most accurate sea-level reconstructions, as there are many species with very narrow tolerance ranges (Allen, 1990; Gehrels *et al.,* 2001). Sediment from the mid to low intertidal zone have the highest error ranges as there are much wider environmental tolerances and poorly defined lower elevation ranges, of taxa such as calcareous foraminifera (Barlow *et al.,* 2013; Elliott, 2016).

The most realistic reconstruction does not necessarily equate to the model with the smallest errors. A number of considerations must be taken into account, including the number of good analogues. The reconstruction should be compared to models with different lines of evidence, and a reliable model should agree closely with other evidence. The reconstruction should also agree with the environmental and ecological interpretations based on biostratigraphical analyses. For instance, if the diatom or foraminifera evidence implies a rising sea level, but this is not reflected in the numerical model, there may be problems with the transfer function analysis. It is also important to use independent measures to assess the reliability of the reconstruction, (such as tide gauge data) (Barlow *et al.,* 2013).

3.4 Conclusions

Three sites were sampled in the Dysynni Valley at Penllyn, Perfeddnant and Gesail. A stratigraphic survey was undertaken at each site and a sample core was obtained which represented the most complete sequence. Sediments were described following Tröels-Smith (1955). Loss on ignition and particle size analysis was undertaken at all sites and samples taken for radiocarbon dates at Penllyn and Perfeddnant. Cores were sampled for diatoms and foraminifera from all three sites, with pollen samples taken from Penllyn. Statistical analyses were undertaken on the sediments from Penllyn and Perfeddnant including sea level index points and transfer functions.

A results chapter for each site will discuss the results of the stratigraphic survey, and review the biostratigraphic analyses including loss on ignition, particle size, diatom, foraminifera and pollen, where appropriate. Radiocarbon dates will be presented. Sea level index points and transfer functions will be described followed by a review of the environmental change at each site.

Section B: Presentation of Results

Chapter 4: Penllyn

4.1 Introduction

Penllyn farm is located to the south of Tywyn (SN583998); which is situated on a drumlin; a hill of glacial till predominantly composed of poorly sorted shale gravel (Pratt *et al.,* 1995) (Figure 4.27). The southern edge of the drumlin can be seen bisecting the farmland at Penllyn and marks the change from well drained pasture to marsh (Figure 4.28). The drumlin ridge of Tywyn gives way to the broad, flat basin of Penllyn, constrained by a storm beach barrier to the west, the drumlin to the north, and the uplands to the east (Leng and Pratt, 1985; Smith *et al.,* 2002)

Figure 4.27: Photograph of Penllyn farm, looking northeast towards Tywyn.

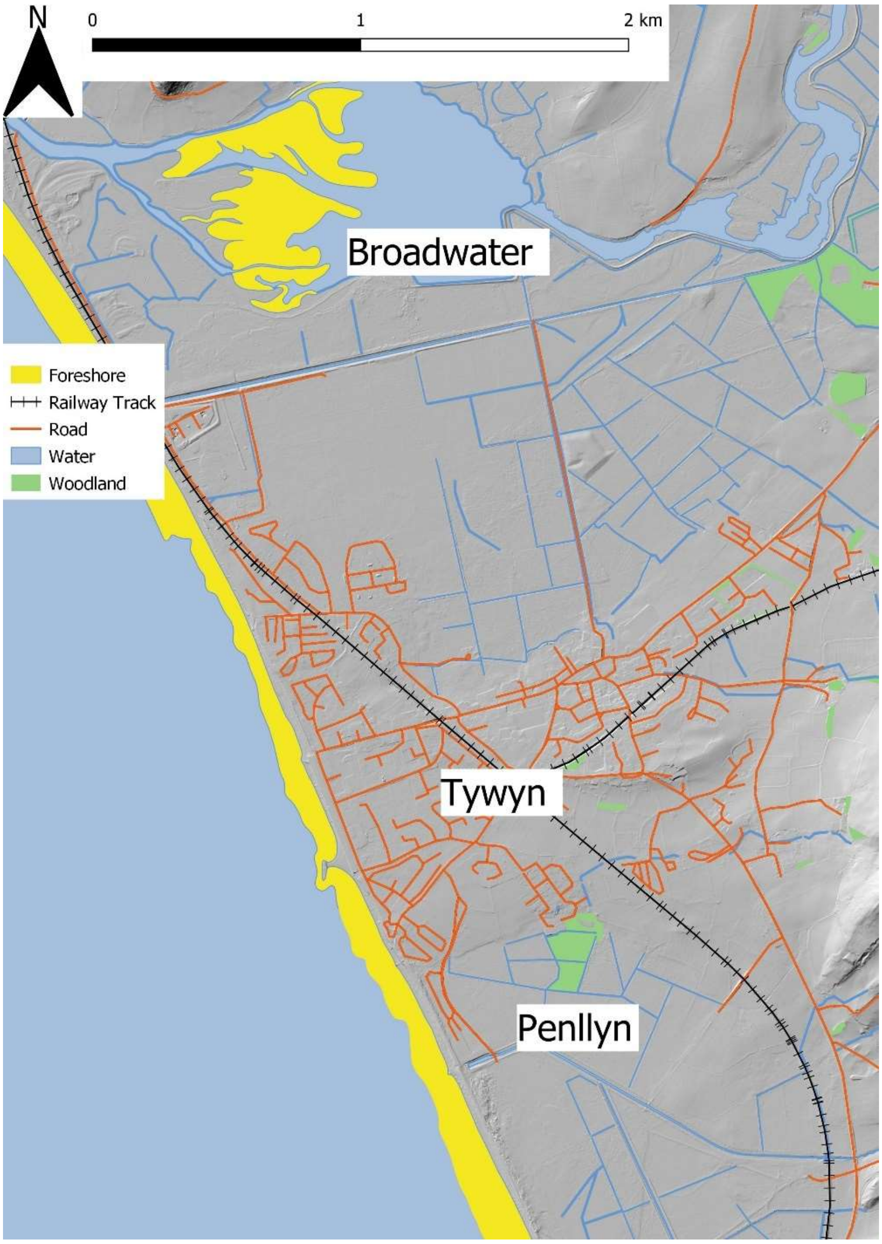


Figure 4.28: Location map showing Penllyn, and sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved.

The gravel of the drumlin produces dry grassland to the north of the site with some clover, dandelion and creeping buttercup. The northernmost field is very flat, giving way to a gentle slope with a concentration of thistles in an area of wet ground. South of the drumlin ridge, there is wet pasture and marshland with tall grass and reeds, as well as docks, sorrel, clover, nettles, thistles, creeping buttercup, ribwort plantain and mother-of-thousands. The ditches contain tall reeds and grass. The woodland to the east of the marsh is predominantly conifers with hazel shrubs. At the bank of the culvert to the south (points A and B, Figure 4.29) there is diverse vegetation, including thistles, dandelions, hay cap mushroom, mother-of-thousands, petty spurge, white clover, willowherb, and diverse grasses. The surrounding fields, including south of the culvert, reflect a similar wet grassland, although these fields were not surveyed. Between the beach and the fields there is a gravel storm beach that gives way to dunes, with a gravel beach south of the culvert. The field to the south is mowed wet grassland with some thistles and stumps of spiky grass as well as clover, creeping buttercup, and mother-of-thousands (Point D, Figure 4.29). There is a patch of un-mowed grass (Point C, Figure 4.29) which reflected the same species as the mowed grass, as well as the taller meadow buttercup. The ditches to the south of the site contain bullrushes, grass and dock, with the disturbed edges of the fields dominated by ribwort plantain, grass and thistles as well as clover and meadow buttercup. The vegetation on the gravel bank and dunes include grass, daisies, holly, sorrel, docks, dandelions, ferns and mother-of-thousands.

Figure 4.29: Map showing results of the vegetation survey with vegetation zones and survey locations. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved.

Leng and Pratt (1987) surveyed the area south of Penllyn farm (see Figure 4.30) and identified blue-grey clay overlain by a submerged forest peat, behind a storm-beach barrier. At low tide there was evidence for extensive buried peats with *in situ* logs, which have been recorded from the south of Tywyn to Borth (Smith *et al.*, 2004). Leng and Pratt (1987) concluded that there was a storm barrier out to sea during the early Holocene, which moved inland during rapid sea level rise. A similar sequence has been found nearby at Borth and Ynyslas, with blue-grey clay overlain by submerged forest peat which was dated to 6179-5663 cal. BP (HAR1020) and 5586-5312 cal. BP (HAR 1019) (Wilks, 1977, 1979; Chapter 2). Caseldine (in: Smith *et al.*, 2004) undertook a preliminary assessment of peat samples from the submerged forest peat at Penllyn and found macrofossil evidence for woodland giving way to reed swamp. Some remains of charcoal were found, either caused by human occupation or natural fires.

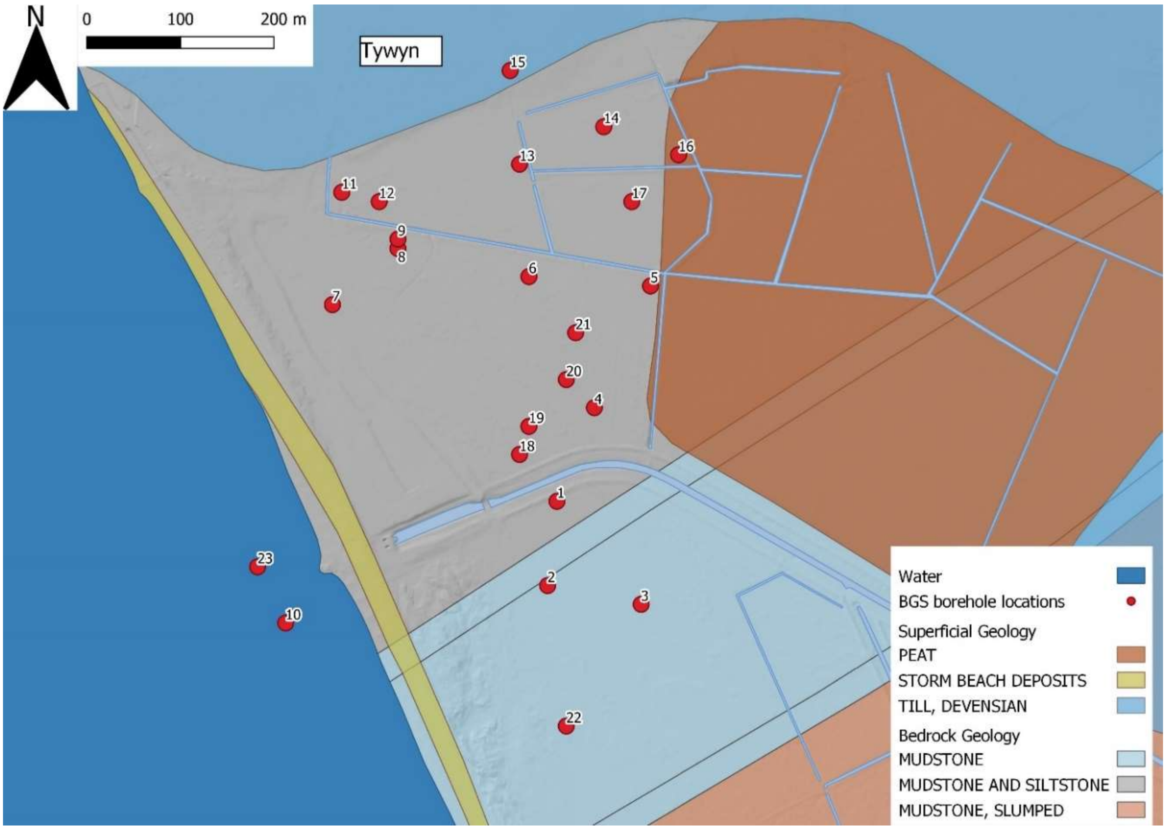


Figure 4.30: Map showing British Geological Survey borehole locations at Penllyn (Leng and Pratt 1987) with bedrock and surface geology. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved. Geological Map Data BGS © UKRI 2022

4.1.1 Archaeological Evidence

Despite the extensive intertidal peats prevalent at Tywyn, no Mesolithic finds have been reported (Smith *et al.*, 2002). Similar submerged forest peats to the south at Borth in the Dyfi estuary produced Mesolithic finds including a Mesolithic flint pick, flint flakes, an antler tool and a hearth (Houlder, 1994; Sambrook and Williams, 1996; Smith *et al.*, 2002).

At Aberystwyth, the site of Penyranchor was excavated in 1922, revealing evidence for extensive Mesolithic flint production as well as charcoal, fragmentary bone and burned flint from a low headland at the foot of Pen Dinas (Houlder, 1994). The flint industry at Penyranchor comprised several thousand pieces including cores, waste flakes, spalls and the characteristic microlith points

(Houlder, 1994). Nearby at Plas Goegerddan, amongst Neolithic and early Bronze Age flint tools revealed from fieldwalking, there were some lithics of possible Mesolithic origin (Smith *et al.*, 2002).

A number of Neolithic stone axes have been found around Cardigan Bay, with two confirmed Neolithic stone axes found at Celmi farm in the Dysynni valley (see Figure 4.31) (Bowen and Gresham, 1967); one was made of rock from Mynydd Rhiw, on the Llŷn Peninsula, the other made of quartz-diorite, possibly from northern Pembrokeshire, indicating the importance of coastal movement (Bowen and Gresham, 1967).

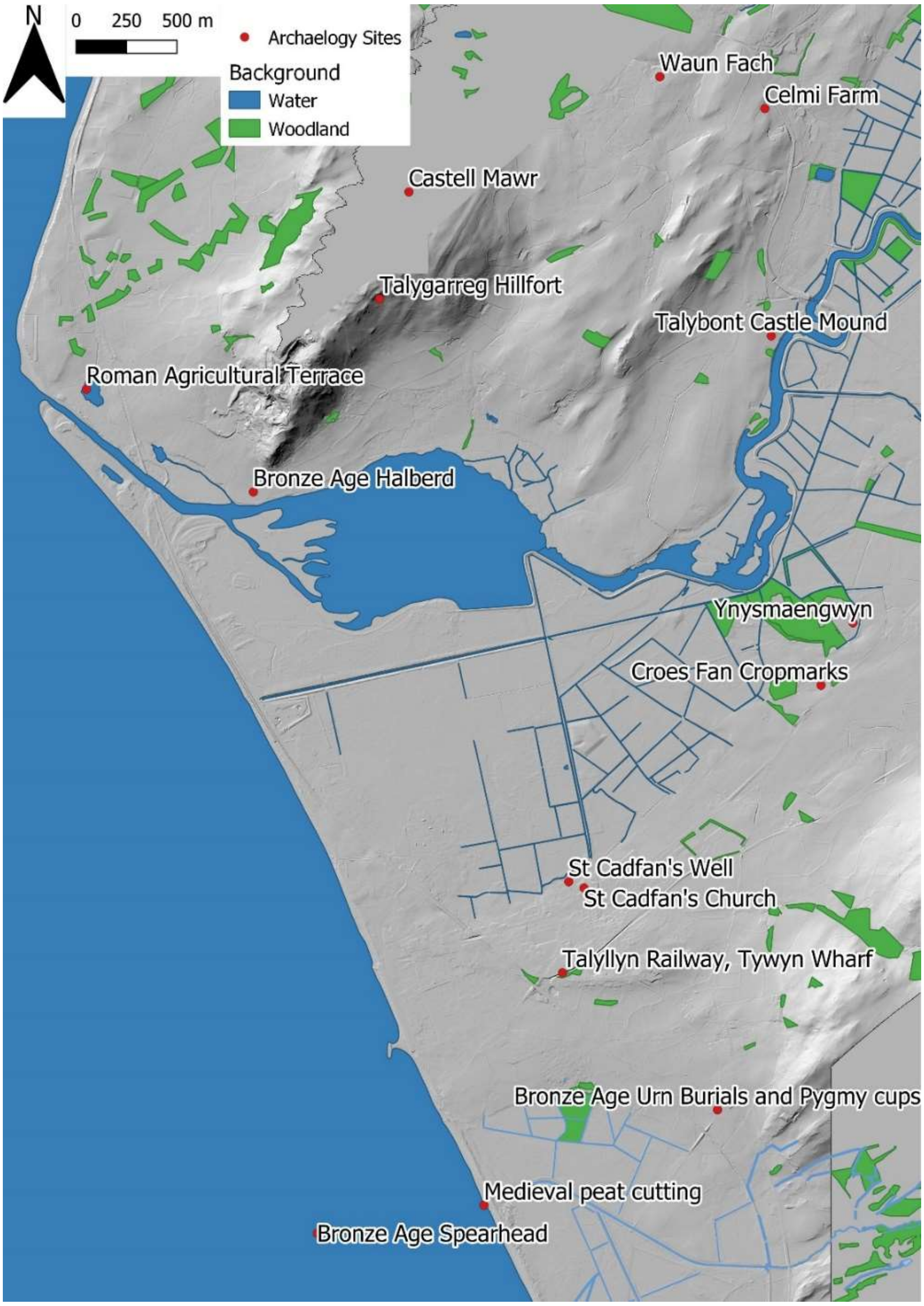


Figure 4.31: Archaeological evidence near Penllyn, indicating sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales and Archwilio HER information © Natural Resources Wales and Database Right. All rights Reserved

Several Bronze Age artefacts have been found within the Dysynni valley. There are a number of standing stones in the hills surrounding the valley, including many cairns as well as the Waun Fach stone near Llanegrin (Smith, 2001). A Late Bronze Age spearhead was found near Tywyn (Gwynedd

Archaeological Trust PRN4813, Briggs, 1979; Smith *et al.*, 2004, 2005) as well as a group of early Bronze Age burials and burial urns near Pantyneuadd (Gwynedd Archaeological Trust PRN4806), likely indicating the location of Bronze Age settlement nearby (Smith *et al.*, 2004, 2005). There are also many Bronze Age finds from the Dysynni valley, including a halberd found at the quarry of Tonfannau and a number of pygmy cups found in Pant y Neuadd, Tywyn (Frost, 2012).

There are a number of small Iron Age hillforts along the west coast of Wales which are largely coastal, possibly built as beach heads built by settlers arriving by sea (Garrard and Dobson, 1974). In the Dysynni valley two such hillforts can be seen at Castell Mawr and Castell y Gaer (Garrard and Dobson, 1974). The hillfort Castell y Gaer is on the outskirts of Llwyngwril between Tywyn and Barmouth, formed of earth and stone walls (Garrard and Dobson, 1974). Castell Mawr is located on a rocky outcrop extending to Mynydd Talgarreg, with a large bank forming an oval enclosure and a second bank to the north indicating a second period of occupation (Garrard and Dobson, 1974). The small, fortified round hut Talygarreg was found above Broadwater, however it was similar to the small medieval castle Carreg y llam on the Llŷn peninsula, which was dated to the eighth to ninth centuries AD (Garrard and Dobson, 1974).

There is very little Roman archaeology on the west coast of Wales, with the closest fort to the Dysynni to the south at Cefn Caer, Pennal, near Machynlleth which commanded the ford of the river Dyfi (Garrard and Dobson, 1974). The Roman Road, Sarn Helen, reached from Aberconwy to Carmarthen, with a stretch of road documented between Pennal and Llanio Roman military settlements (NPRN300159, NPRN303530) (Davies, 1994). There were two Roman agricultural terraces north and south of the Afon Gwril on the slopes overlooking the village of Llwyngwril, although little was preserved (Garrard and Dobson 1974). There was evidence for similar terraces at Tonfannau at the north of the Dysynni river mouth near the farm Cefncamberth (Garrard and Dobson, 1974).

The early medieval settlement at Tywyn was focussed on St. Cadfan's church (NPRN 43861) and the nearby St Cadfan's Well (Ffynnon Gadfan, NPRN 32397). There are records of four early medieval inscribed stones, including the earliest known Welsh inscription (NPRN 302689). The settlement at Tywyn was centred on a drumlin, which would have represented a dry island of gravel within an estuary which likely extended inland. There is evidence for medieval peat cutting at the coast near Tywyn, which is exposed during storm events (NPRN 411870, Smith *et al.*, 2004). The peat cutting beds consist of square and rectangular trenches and although no specific records relate to the peat cutting at Tywyn, there are records which document the exploitation of peat in the area in the eighteenth century, which in other parts of north Wales continued into the mid-nineteenth century (Smith *et al.*, 2004).

Smith's (2004, 2005) reviews of the archival sources and historic maps provide a summary of the recent landscape changes at Penllyn. Prior to enclosure, the area was dominated by marshes and lagoons. The earliest detailed map of Penllyn was produced in 1748 and shows the Penllyn area as "low marshes". The earliest OS map in 1837 shows a large lake named Llyn y borth, with several small pools behind the shingle bank and sand dunes, with an area of marsh to the south named Gwerglodd. The name Penllyn translates to the 'head or 'top of the lake', which is likely a reference to Llyn y borth. The 1837 map shows that the Afron Dyffryn Gwyn had a more winding channel, which drained into Llyn y borth and through the shingle bank. The 1841 tithe map indicates a smaller lake named Penllyn Pool. By 1891, the map shows that after drainage, the Afron Dyffryn Gwyn was channelled into a culvert under the shingle bank, and the lake was drained, marked only as a marsh liable to flooding.

4.2 Fieldwork

4.2.1 Site stratigraphy

The site Penllyn farm, south of Tywyn, was chosen for investigation due to the presence of blue grey clay recorded by Leng and Pratt (1987), potentially of estuarine or marine origin. A north-south transect of boreholes was undertaken in order to confirm the presence and extent of the blue-grey clay (Figure 4.32). An east-west transect was undertaken to establish the depth of the blue grey clay across the site and to establish the extent of the sand found by Leng and Pratt (1987).

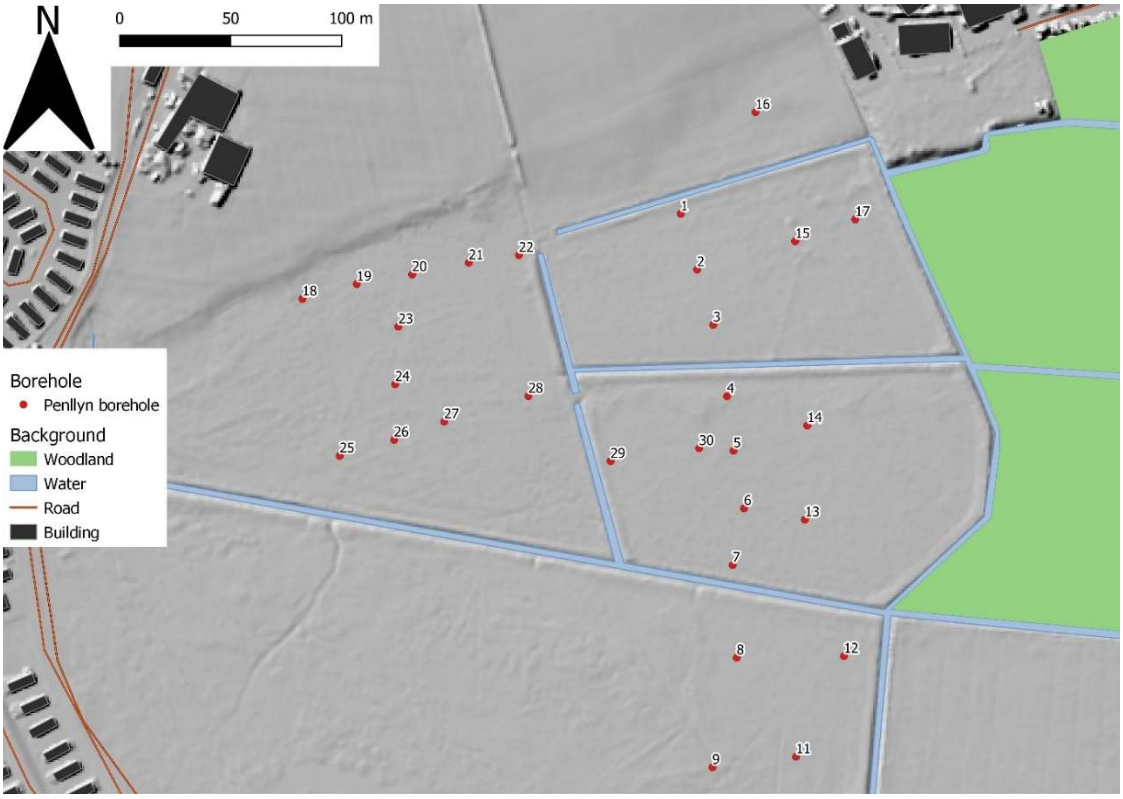


Figure 4.32: Map showing gouge survey locations for Penllyn. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved

In the base of some cores a coarse angular gravel was found in a dark blue clay matrix, predominantly shale fragments (2-10mm long) with occasional quartz (Figure 4.33). The gravel matched the description of boulder clay by Leng and Pratt (1987). To the north of the transect, the surface elevation rose abruptly and was too hard to core with a gouge auger, likely due to the underlying drumlin upon which Tywyn is situated.

In the north-south transect, the basal gravel was overlain by a thin layer of black, well humified peat (Figure 4.33). The basal peat was overlain by a soft blue-grey silt with reeds and black inclusions. In places, orange mottling was observed, possibly indicating an iron pan. The surface altitude of the blue-grey silt was consistent across the site between 0 and -1m OD. The blue silt at Penllyn was overlain by a woody peat layer. There was a concentration of wood at the base of the peat layer, with larger obstructions at similar depths preventing coring in some locations and possibly representing logs or tree stumps.

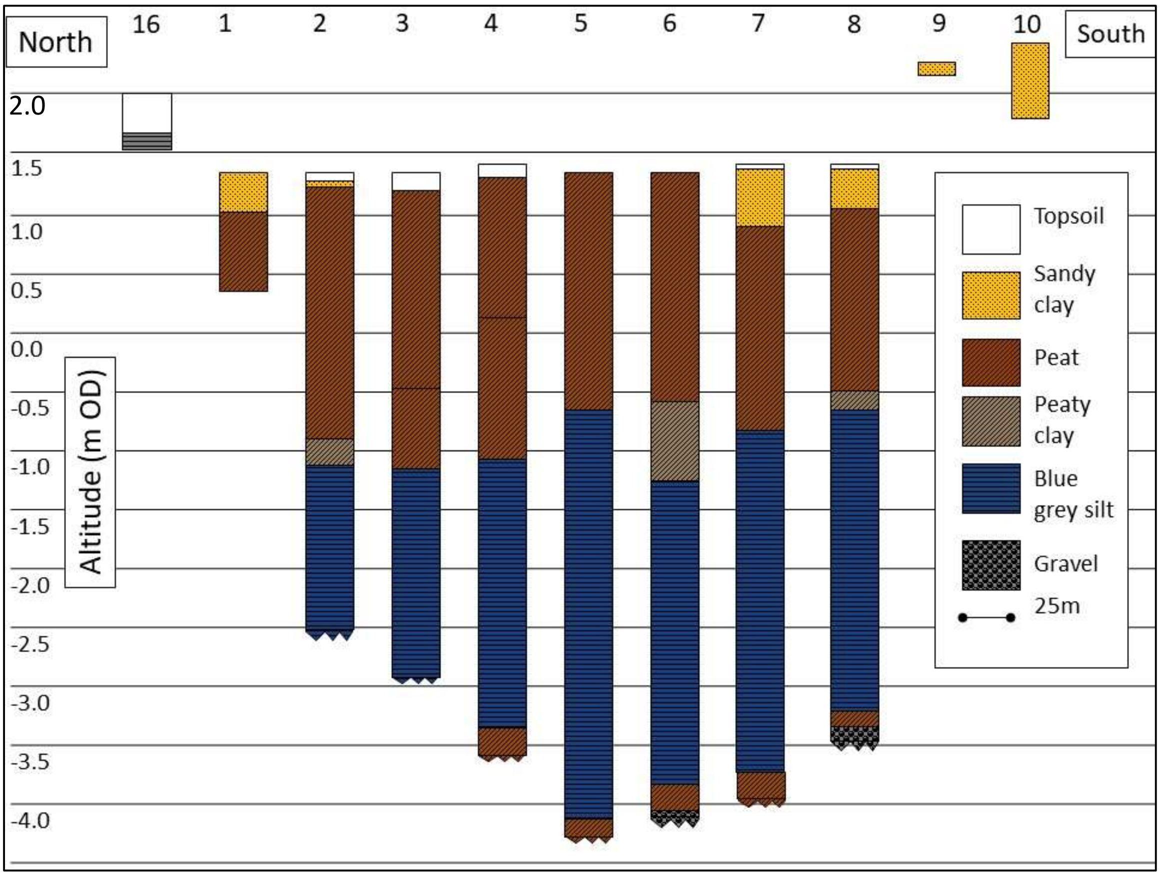


Figure 4.33: Stratigraphy at Penllyn for the north-south transect, boreholes 16, 1 to 10

Two transects were extended to the west to test the stratigraphy and establish the deepest sediments. Both transects revealed a thin layer of sandy clay at the surface towards the west, possibly indicating blown sand from the coast. The transect for boreholes 18 to 22 terminated in sand between -1 and 0m OD (Figure 4.34). The southern west-east transect recovered a sequence of blue-grey silt overlain by woody peat, similar to the boreholes in the north-south transect, with some sand at the surface. However, very little basal peat was recovered, only a thin layer of peaty clay (Figure 4.34 and 4.35)

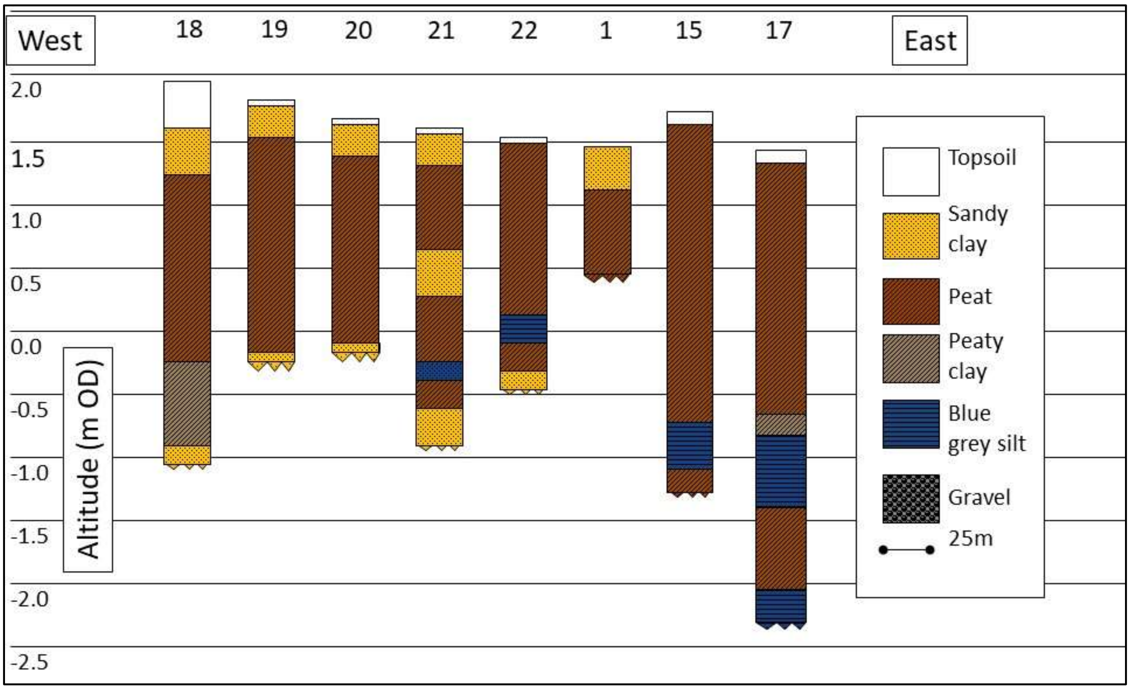


Figure 4.34: Stratigraphy at Penllyn for the northern west-east transect, boreholes 18 to 22, 1, 15, 17

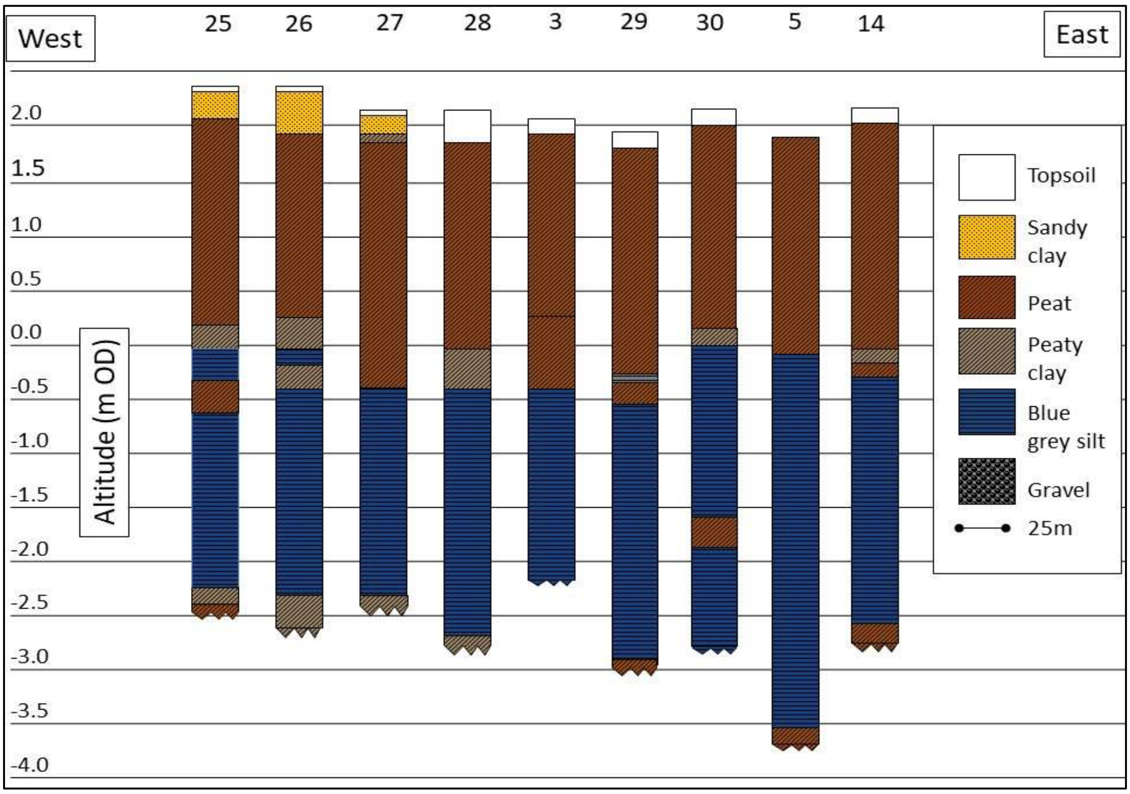


Figure 4.35: Stratigraphy at Penllyn for the southern west-east transect, boreholes 25 to 28, 3, 29, 30, 5 and 14

Preliminary analysis of samples from boreholes 4, 5 and 6 revealed well preserved diatoms. Borehole 5 was chosen for further analysis as it represented the deepest and most complete sequence of sediments. A core was obtained using a Russian-style corer, which recovered 5.15m of sediment including basal peat, blue grey silt, and overlying woody peat. The core was placed in guttering and wrapped in cling film and foil and taken to the cold store.

4.3 Laboratory work

In the laboratory, the stratigraphy of core 5 was described and the core was subsampled for loss on ignition, particle size and diatom analyses.

4.3.1 Stratigraphy

A well humified, black basal peat was recovered between 515-500cm (-3.39 to -3.24 m OD), overlain by soft blue grey clay with reeds to 232cm (-0.56m OD). A wood fragment was recovered between

232-230cm (-0.56 to -0.54m OD), overlain by peaty clay to 214cm (-0.38m OD). This was overlain by woody peat up to 15cm (1.61m OD), with some variations in colour, texture and inclusions.

Table 4.28: Sediment description of Penllyn stratigraphy following Tröels Smith (1955). Nig=Nigro, Strf- Stratificatio, Elas = Elasticitas, Sicc = Siccitas, Strf = Structura, lim = Limes, Hum = Humositas. See methods section.

Depth (cm)	Altitude (m OD)	Description	Nig	Strf	Elas	Sicc	Colour	Stru	Lim	Hum	Composition
0-15	1.76 to 1.61	Topsoil with roots	2	0	0	4	Light brown	Fibrous	0	0	Tb3 Th1
15-30	1.61 to 1.46	Black peat with roots	3	0	0	3	Black-brown	Fibrous	0	1	Tb3 Th 1
30-50	1.46 to 1.26	Black brown peat with reeds and roots	3	0	2	3	Black-brown	Homogeneous	0	2	Tb4+reeds and roots
50-58	1.26 to 1.18	Black humified woody peat with reeds and roots	4	0	3	2	Black	Homogeneous	0	3	Tb4 +reeds, roots and wood.
58	1.18	Large fragment of wood									
58-170	1.18 to 0.06	Black humified woody peat with reeds and roots	4	0	3	2	Black	Homogeneous	0	3	Tb4 +reeds, roots and wood.
170-200	0.06 to -0.24	Black humified woody peat with reeds	4	0	3	2	Black	Homogeneous	0	3	Tb4 +reeds, and wood.
200-214	-0.24 to -0.38	Black peat with reeds and roots	4	0	4	3	Black	Homogeneous	0	4	Tb4 + reeds +roots
214-230	-0.38 to -0.54	Grey-brown peaty clay	2	0	4	4	Grey-brown	Homogeneous	1	3	Tb 2, As2 plus roots and reeds
230-232	-0.54 to -0.56	Wood fragment									
232-500	-0.56 to -3.24	Blue grey silt	1	0	3	3	Blue-grey	Homogeneous	2	0	As4 plus reeds
340	-1.64	From about 340cm there is an increase in black and orange mottling									
500-515	-3.24 to -3.39	Well humified black peat	4	0	2	2	Black	Homogeneous	-	4	As1, Tb3

4.3.2 Loss on Ignition

Subsamples were taken for loss on ignition at 4cm intervals to establish changes in organic and carbonate content (Figure 4.36). The method used followed Heiri *et al.,* (2001, Chapter 3.3.2). The basal peat had high organic content, up to 74%. Throughout the blue grey clay both the carbonate and the organic content were low, between 0 and 10% with no evidence of disturbance. Between 280 and 218cm (-1.04 to -0.42m OD) the organic content rose to between 10-30%, which reflected the transition from clastic to organic sediment as the estuarine channel became infilled. There was a high organic content in the peat, between approximately 80 and 95%, with a low carbonate content.

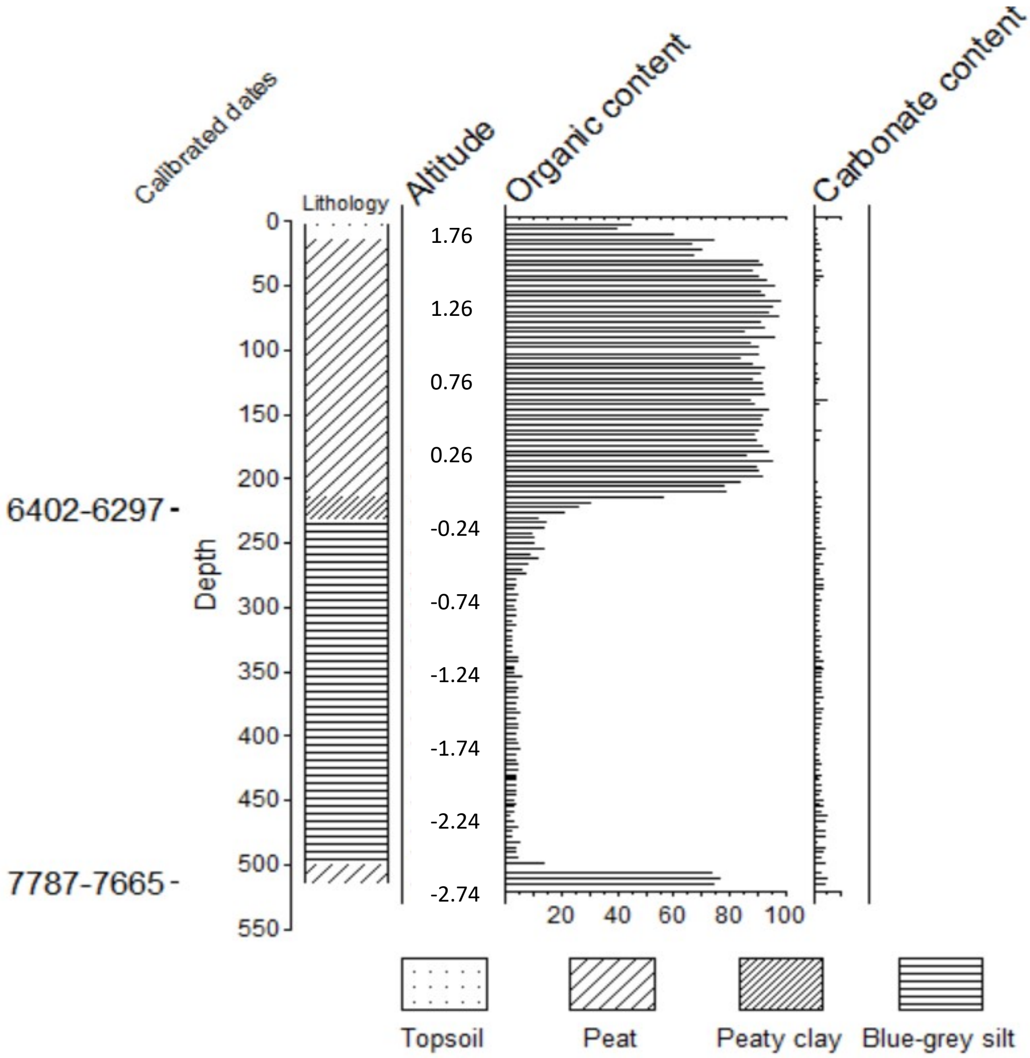


Figure 4.36: Loss on Ignition results for Penllyn

4.3.3 Particle Size

Subsamples for particle size analysis were taken at 24cm intervals and pre-treated with hydrogen peroxide to remove organic content (Figure 4.37). Due to the high organic content, only depths of 496 to 243 (-3.20 to -0.67 m OD) were processed. A Malvern laser granulometer was used to determine particle size (Chapter 3.3.3).

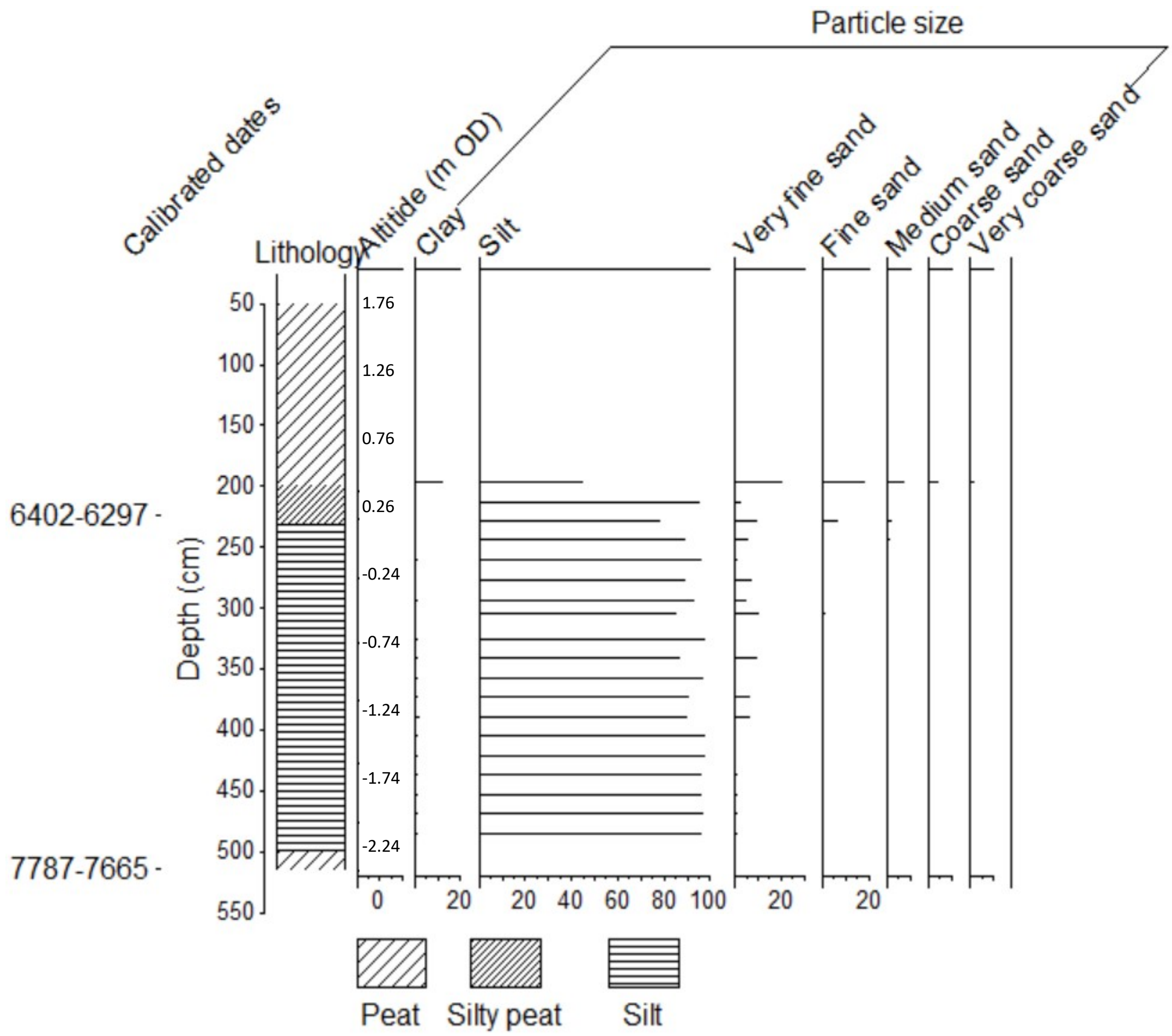


Figure 4.37: Graph showing the particle size data for Penllyn

The blue grey silt was dominated by the silt fraction, with some clay, which indicates a still water environment. Low levels of the very fine sand fraction were present throughout, which increased to c. 10% at 400cm (-2.24m OD) with low levels of fine sand appearing, indicating either an increase in fluvial activity or the presence of in-blown sand from the coast. There was a peak in the very fine,

fine, medium and coarse sand fractions at 340cm (-1.64m OD). At 260cm (-0.84m OD) medium and coarse sand appeared and at 200cm (-0.24m OD) there was a peak in all sand fractions. The peaks in sand fractions could possibly indicate inwash of fluvial or coastal material

4.3.4 Radiocarbon dates

Two samples for radiocarbon determination were submitted from the upper and lower boundaries of the blue-grey silt, where the diatom concentrations decline. Bulk samples were submitted due to the absence of *in situ* macrofossils, as only large wood fragments and degraded reeds were present.

Table 4.29: Radiocarbon dates for Penllyn. Results are presented in units of percentage modern carbon (pMC), the uncalibrated radiocarbon age before present (BP) and the calibrated age before present (cal. BP). All results have been corrected for isotopic fractionation with an δ13C value measured on the prepared carbon by the accelerator. These δ13C values provide the most accurate radiocarbon ages, but cannot be used to investigate environmental conditions, nor for trophic and nutritional interpretations. The pMC reported requires no further correction for fractionation.

DirectAMS code	Sample depth (cm)	Sample type	δ (^{13}C)	Fraction of modern carbon		Radiocarbon age		Calibrated age (cal. BP)
				pMC	1σ error	BP	1 σ error	95.4% probability
D-AMS 028035	224-225	Sediment (peat)	-34.8	50.06	0.19	5558	30	6402-6297
D-AMS 028034	514-515	Sediment (peat)	-25.8	42.46	0.15	6881	28	7787-7665

4.3.5 Foraminifera analysis

Subsamples for foraminifera were taken at 4cm intervals throughout the core and 8 species were identified in total. No foraminifera were found from 221cm to the surface 9-0.44 to 1.76m OD). Concentrations were calculated based on the total volume counted, and seven subzones were established using CONISS (Figure 4.38, Table 4.38).

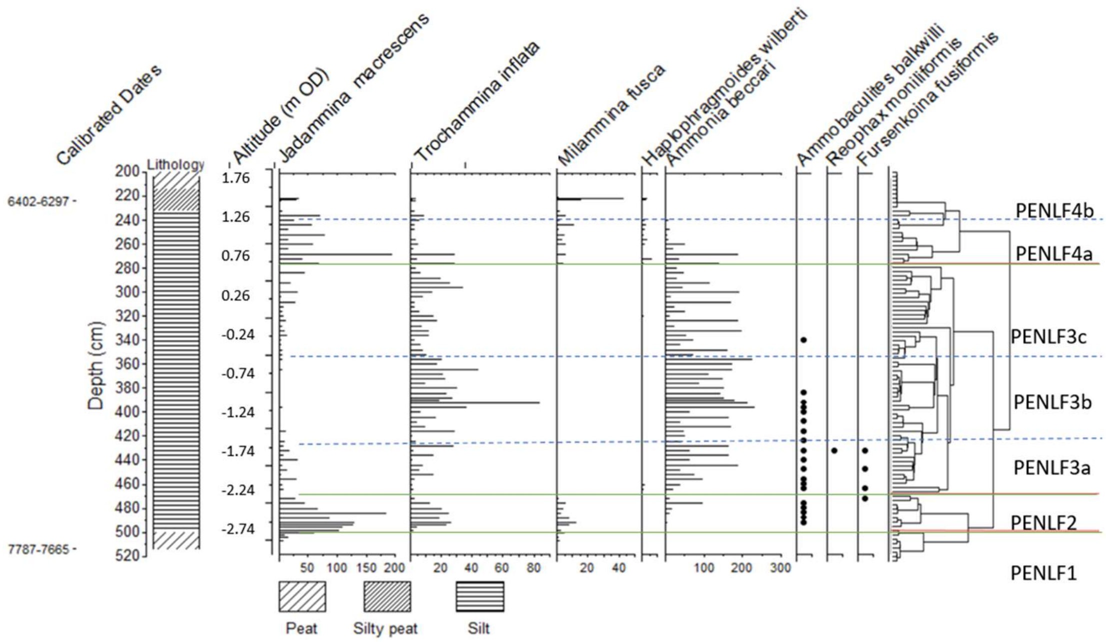


Figure 4.38: Foraminifera concentration diagram from Penllyn with stratigraphy, radiocarbon dates and zones. No foraminifera were found from 221cm to surface 9-0.44 to 1.76m OD).

Table 4.30: Summary foraminifera zones from Penllyn

Zone	Depth (cm)	Altitude (m OD)	Description	Environmental conditions
PENLF1	515-502	-3.39 to -3.26	The basal peat contained very low counts of *Jadammina macrescens* (less than 10) and single tests of *Millamina fusca* appearing at the top of the zone.	There was a high marsh environment on the very edge of marine influence
PENLF2	502-470	-3.26 to -2.94	*J. macrescens* was dominant, with high levels of *Trochammina inflata* and *M. fusca*. Low levels of *Ammonia* sp. and *Ammobaculities balkwilli* appeared towards the top of the zone	A more diverse high marsh, agglutinated assemblage
PENLF3a	470-398	-2.94 to -2.22	*Ammonia* sp. was dominant, while *J. macrescens* and *T. inflata* declined. *M. fusca* all but disappeared. *A. balkwilli* was present in low amounts as well as occasional tests of *Reophax moniliformis* and *Fursenkoina fusiformis.*	A less diverse, calcareous assemblage indicating a low marsh or intertidal environment
PENLF3b	398-326	-2.22 to -1.5	*Ammonia* sp. continued as the dominant species, *T. inflata* increasing. *J. macrescens* almost disappeared. *A. balkwilli* was present in small amounts.	intertidal assemblage with tidal action bringing inner shelf foraminifera onto the marsh
PENLF3c	326-270	-1.5 to -0.94	*Ammonia* sp. and *T. inflata* remained the dominant species, with levels of *J. macrescens* increasing. *M. fusca* and *Haplophragmoides* sp. appeared towards the top of the zone	Intertidal low marsh with gradual transition to high marsh
PENLF4a	270-221	-0.94 to -0.44	*J. macrescens* became the dominant species, with low levels *of T. inflata, M. fusca* and *Haplophragmoides* sp. present as well as very small amounts of *Ammonia* sp.	High marsh
PENLF4b	221 to surface	-0.44 to 1.76	Barren of foraminifera above 221cm	No marine influence

Zone PENLF1 (515-502cm, -3.39 to -3.26m OD) represented the basal peat, dated to 7787-7665 cal. BP (D-AMS 028034). There was a high marsh environment on the very edge of marine influence; indicated by very low counts of *Jadammina macrescens*, with less than 10 tests per sample. The brackish *J. macrescens* is ubiquitous to saltmarshes worldwide (Murray, 1979) and a low concentration, monospecific assemblage of *J. macrescens* has been found at the edge of marine influence at many sites (Scott and Medioli, 1978; Wright *et al.,* 2011; Horton and Edwards, 2006). Towards the top of the zone the euryhaline *Millamina fusca* appeared, which is a sediment dwelling species found in brackish lagoons, estuaries, and tidal marshes (Murray, 1979).

Across the transition to the blue-grey silt, Zone PENLF2 (502-470cm, -3.26 to -2.94m OD) reflected a shift towards a more diverse high marsh, agglutinated assemblage. *J. macrescens* rose to dominate the profile, with high levels of *M. fusca* and the sediment dwelling, high marsh *Trochammina inflata* (Murray, 1979). The agglutinated assemblage of *J. macrescens*, *T. inflata* and *M. fusca* has been found worldwide in high marsh environments (Horton and Edwards, 2006; Wright *et al.,* 2011; Kemp *et al.,* 2015). At the top of the zone, low levels of *Ammobaculities balkwilli* and the calcareous species *Ammonia* sp. appeared. *A. balkwilli* is a sediment dwelling species found commonly in the nearby Dyfi estuary (Murray, 1979). Horton and Edwards (2006) found very low counts of *A. balkwili* on the south coast of England, at both Bury Farm, Southampton Water, in the mid-low marsh, and at Arne peninsula, Poole Harbour, near the boundary between low marsh and tidal flat. *Ammonia* sp. is a calcareous, intertidal species common in the mid to low marsh (Murray 1979, Horton and Edwards, 2006; Gehrels *et al.,* 2001; Horton *et al.,* 1999).

In zone PENLF3a (470-398cm, -2.94 to -2.22m OD) there was a change from the high marsh, agglutinated assemblage towards a less diverse, calcareous assemblage, with *J. macrescens, T. inflata* and *M. fusca* replaced by *Ammonia* sp. as the dominant species. This reflects a shift towards a low marsh or intertidal environment (Horton and Edwards, 2006). *Ammonia* sp. is a diverse, marginal marine genus. Those with an umbilical boss can be found in marine, inner shelf environments, however at Penllyn the majority of *Ammonia* sp. were found without the umbilical boss, which is a variation of the species *Ammonia beccari* typical of estuaries and lagoons (Murray, 1979).

Occasional tests of *A. balkwilli* were found, as well as the marine, inner shelf species *Reophax moniliformis* and *Fursenkoina fusiformis*, which are sometimes transported into estuaries and are likely be allochthonous[33] (Murray, 1979). Care must be taken, however, as calcareous foraminifera are fragile, and likely to dissolve, leading to an underestimation of calcareous species in the profile (Gehrels *et al.,* 2001; Horton and Edwards, 2006). Zone PENLF3 may have been influenced by differential preservation, as the zone is dominated by *Ammonia* sp. with few calcareous species often found in low marsh and tidal flats in the UK, such as *Haynesia germanica* or *Elphidium* sp. (Horton and Edwards, 2006).

In subzone PENLF3b (398-326cm, -2.22 to -1.50m OD) the calcareous, intertidal assemblage continued as *Ammonia* sp. remained dominant, with low levels of *A. balkwilli*. Most agglutinated species had disappeared, except *T. inflata*, which became more prominent. Both inner shelf species *F. fusiformis* and *R. reophax* were absent.

In subzone PENLF3c (326-270cm, -1.50 to -0.94m OD), the agglutinated foraminifera *J. macrescens* began to increase, with high levels of *T. inflata* implying a reduction in marine conditions. High levels of *Ammonia* sp. indicated continued low marsh environments. Other agglutinated species such as *M. fusca* and *Haplophragmoides* sp. emerged at the top of the zone. *Haplophragmoides* sp. is a sediment dwelling, high marsh species (Murray, 1979), found across the saltmarsh, but is more prevalent above MHWST (Horton and Edwards, 2006). The increase in agglutinated, high marsh foraminifera towards the top of PENLF3 indicates a gradual transition to a high marsh environment.

Across the transition into peaty clay, subzone PENLF4a (270-221cm, -0.94 to -0.44m OD) represented a shift towards high marsh environments, as the low marsh, calcareous *Ammonia* sp. was replaced by an agglutinated assemblage of *J. macrescens* with low levels of *T. inflata, M. fusca* and *Haplophragmoides* sp. These agglutinated species are common to the high marsh and indicate marginal marine conditions (Murray, 1979; Horton and Edwards, 2006). At 224cm (-0.48m OD), the peat was dated to 6402-6297 cal. BP (D-AMS 028035). From 221cm to the surface (Zone PENLF4b,

[33] transported

-0.44 to 1.76m OD) no foraminifera were found, implying the conditions were freshwater through the peat.

4.3.6 Diatom analysis

Subsamples for diatom analysis were taken at 8cm intervals throughout the silt, with samples taken at 4 and 2cm over stratigraphic boundaries. Where possible, 250 valves were counted, however poor preservation made this difficult. No diatoms were found from 223 to the surface (-0.48 to 1.75m OD). Some samples were only counted to 100 valves, at 16cm intervals, particularly between 490 and 350 cm (-3.14 to -1.75m OD) (Figure 4.39, 4.40, Table 4.39). A total of 22 diatom species were identified, however many diatoms could not be identified to species level due to poor preservation, and 13 were identified only to genus level. Two preparation methods were used (Chapter 3.3.4) which confirmed that the poor preservation was not caused by the preparation procedure used.

Table 4.31: Summary diatom zones from Penllyn

Zone	Depth (cm)	Altitude (m OD)	Description	Environmental conditions
PENLD1	515-492	-3.40 to -3.17	Dominated by the aerophilic *Diploneis interrupta* and *Cosmioneis pusilla* as well the epipelic, mesohalobous species *Diploneis didyma*, *Navicula peregrina*, *Tryblionella navicularis*, *Caloneis westii*, *Cyclotella littoralis* and *Scoliopleura turmida.* Some of the polyhalobous species *Diploneis smithii*, *Actinoptychus senarius, Rhaphoneis amphiceros, Rhabdonema arcuatum* and *Amphitetras antediluviana* found as well as the freshwater species *Stauroneis phoenicentron* and *Pinnularia* spp.	An environment near the mean high-water level, where aerophilic species existed in back barrier marshes, with epipelic species in pools which periodically dry out
PENLD2	492-380	-3.17 to -2.05	Lower diversity. *T. navicularis, D. didyma, C. westii, C. littoralis* are present, as well as *D. smithii*, *Petroneis humerosa*, *Campylodiscus neofastuosus* and *A. senarius.* Some freshwater *Eunotia praerupta* and *S. phoenicentron* found	a reduction in saltmarsh conditions with intertidal mudflats, sand banks and a subtidal channel nearby.
PENLD3	380-325	-2.05 to -1.50	Preservation improved. *C. westii, C. littoralis, T. navicularis* and *D. interrupta* present as well as N. peregrina. There was a spike in *D. smithii,* with low amounts of marine taxa *A. senarius, C. neofastuosus, S. turmida, R. arcuatum* and *Auliscus sculptus*.	Continuing marine-brackish intertidal conditions, possibly with a higher salinity
PENLD4	325-276	-1.50 to -1.01	Poor preservation, dominated by broken *T. navilcularis*, with some *C. westii*, *C. littoralis, D. didyma* and *D. smithii*.	Marine influence may have decreased but care must be taken due to poor preservation and low concentrations.
PENLD5	276-223	-1.01 to -0.48	Dominated by *D. interrupta*, with some *C. westii, T. navicularis* and *D. smithii*.	A brackish marsh, with calm conditions. But care must be taken due to few samples.
PENLD6	223 to the surface	-0.48 to +1.75	No diatoms found	

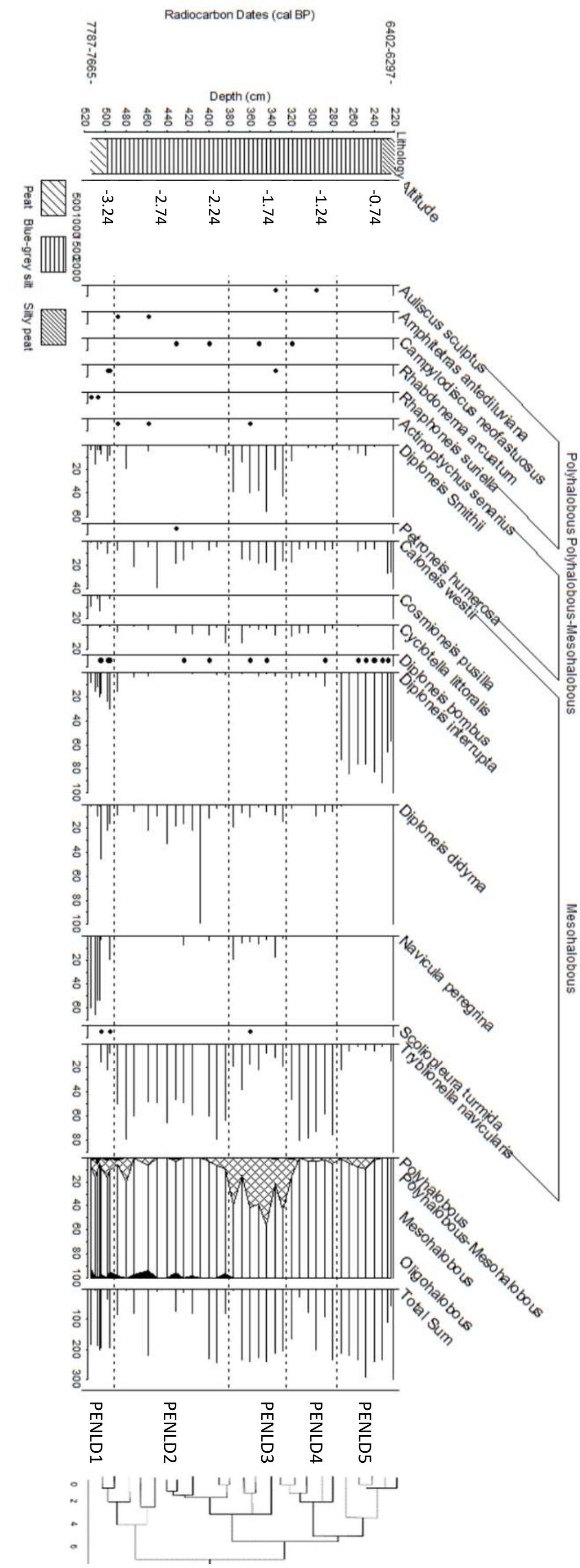


Figure 4.39: Diatom percentage frequency plot for Penllyn showing marine and brackish species, with stratigraphy and radiocarbon dates. Dots indicate presence below 5%. No diatoms were found from 223 to the surface (-0.48 to 1.75m OD).

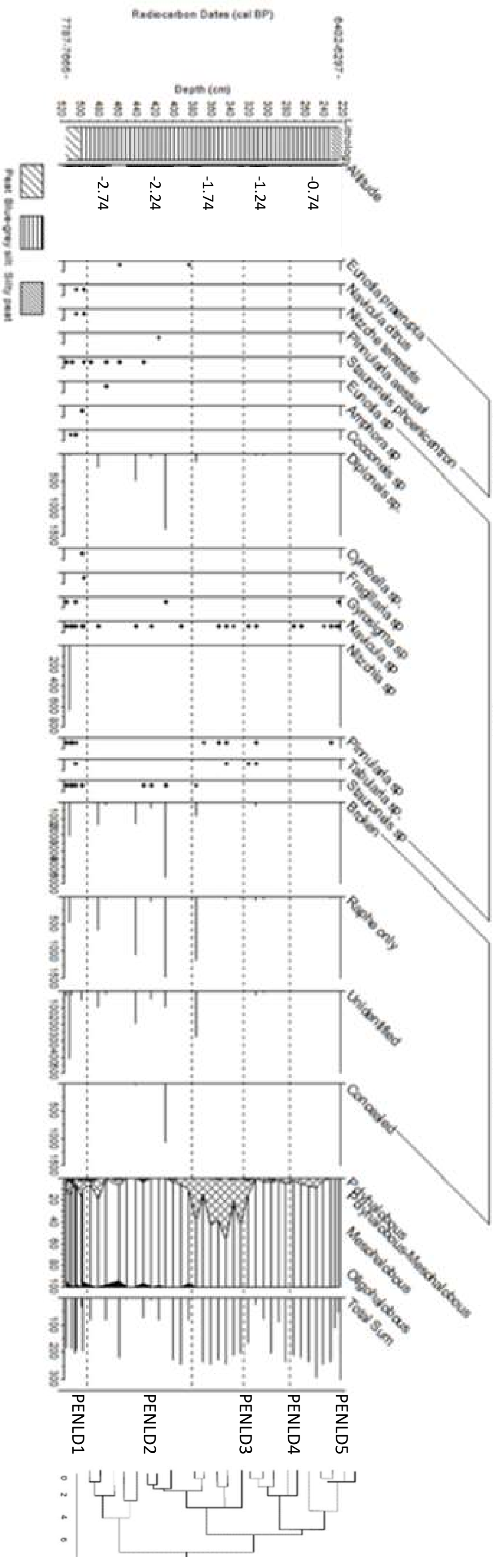


Figure 4.40: Diatom percentage frequency plot for Penllyn showing freshwater species and preservation, with stratigraphy and radiocarbon dates. Dots indicate presence below 5%. No diatoms were found from 223 to the surface (-0.48 to 1.75m OD).

The basal peat, zone PENLD1 (515-492cm, -3.40 to -3.17m OD), dated to 7787-7665 cal. BP (D-AMS 028034), had poor preservation, with many broken valves only identifiable to genus level. PENLD1 contained many robust species, such as *Diploneis interrupta*, and it is possible that some of the more fragile diatoms were lost. *D. interrupta* was present in low amounts throughout the zone, a mesohalobous[34], aerophilic[35] species with a very wide-ranging salinity tolerance (between 500-18,000 mg Cl L^{-1}) common to supratidal[36], salt marsh environments (Denys, 1991/2; Van dam *et al.,* 1994; Vos and de Wolf, 1988). There were also low concentrations of *Cosmioneis pusilla* present in this zone, which is an oligohalobous[37], aerophilic species common to salt marshes, with a salinity tolerance between 0-1000mg Cl L^{-1} (Denys, 1991/2; Van dam *et al.,* 1994; Vos and de Wolf, 1988). Vos and de Wolf (1993) describes the combination of *C. pusilla* and *D. interrupta* as indicative of a supratidal environment, such as a tidal levee or back barrier marsh.

However, many epipelic[38], mesohalobous species were also present in this zone, with high levels of *Diploneis didyma* and *Navicula peregrina,* some *Tryblionella navicularis*, and a single *Scoliopleura turmida* (Denys, 1991/2; Van dam *et al.,* 1994). These species form the *Navicula digitoradiata var minima* group of Vos and De Wolf (1993), typical of clay rich intertidal environments, such as mudflats, creeks, and lagoons (Vos and de Wolf, 1988). The mesohalobous *Caloneis westii* was also present in small amounts, which is common in mudflats and the littoral zone[39] (Denys, 1991/2; Hartley, 1996; Hendey, 1964, Metcalfe *et al.,* 2000; Vos and De wolf, 1993). *Cyclotella littoralis* was found, which is a marine-brackish species tolerant of lower salinity environments (Bahi dos Santos-Fischer, *et al.,* 2016; Wachnicka, 2013; Uliana *et al.,* 2002).

[34] Brackish, salinity between 2-30‰
[35] Frequently exposed to air
[36] the marginal zone above the level of high tide
[37] salinity up to 2‰
[38] live in rocky areas
[39] part of a sea, lake, or river that is close to the shore

The combination of mesolhalobous aeriophilic and epipielic species indicates an environment near the mean high-water level, where aerophilic species such as *D. interrupta* existed in back barrier marshes, with epipelic species such as *D. didyma* in pools which periodically dry out (Vos and de Wolf, 1988).

Some marine taxa were found which had higher salinity tolerances, such as the polyhalobous[40] to mesohalobous *Diploneis smithii* (Denys, 1991/2; Vos and de Wolf, 1988). A few valves of the tychoplankton[41] *Actinoptychus senarius* were found, which is a polyhalobous to mesohalobous species common to subtidal[42] channels (Denys, 1991/2), as well as the polyhalobous *Rhaphoneis amphiceros*, which is found attached to particles in tidal inlets and large tidal channels (Denys, 1991/2; Vos and de Wolf, 1993). Single valves of the polyhalobous marine taxa *Rhabdonema arcuatum* and *Amphitetras antediluviana* were also found (Denys, 1991/2). These marine taxa may indicate the close proximity of a tidal inlet or large tidal channel (Denys, 1994).

There were also some freshwater species present in very small amounts, such as the oligohalobous *Stauroneis phoenicenteon (*Denys, 1991/2) as well as some *Pinnularia* sp. which could not be identified to species level due to poor preservation. These occurred in very low quantities and could have been in freshwater pools nearby or transported from terrestrial areas.

With the transition into the blue grey silt, diatom preservation declined abruptly. PENLD2 (492-380cm, -3.17 to -2.05m OD) had very poor diatom concentration, with many samples only counted to 100 valves. There was much lower species diversity; *D. interrupta*, *N. peregrina* and *C. pusilla* declined, possibly indicating a reduction in saltmarsh conditions. *T. navicularis, D. didyma, C. westii* and *C. littoralis* remained present, possibly indicating intertidal mudflat environments (Denys, 1991/2, 1994; Vos and de Wolf, 1988, 1993), which is reinforced by the combination of *T. navicularis* and *D. didyma* (Denys, 1994). *Petroneis humerosa* was present in small amounts, a polyhalobous

[40] Marine, salinity greater than 30‰
[41] Live in the water column
[42] below the lowest tide

mesoeuryhaline[43] species with a salinity tolerance of 5000-17000 mg Cl L^{-1}, common to sandy intertidal environments such as sandy shoals, beaches, and shallow subtidal marine basins (Denys, 1991/2; Vos and de Wolf, 1988). This assemblage indicates that the back-barrier marsh environment of PENLD1 gave way to an estuarine environment with mudflats and sand banks.

Very low concentrations of the polyhalobous *D. smithii* remained present on the transition to silt (-3.17m OD), as well as individual valves of *Campylodiscus neofastuosus* and *Amphitetras antediluviana* which are both polyhalobous species (Denys, 1991/2). *A. senarius* was also present, which is typical of subtidal channels. These marine species imply that there was a subtidal channel nearby.

Very small amounts of freshwater diatoms were present on the transition to silt (-3.17m OD), such as *Eunotia praerupta*, and *S. phoenicenteon*, which could indicate freshwater nearby (Denys 1991/2). However, the zone has very poor preservation and low diatom concentrations, so care must be taken.

In PENLD3 (380-325cm, -2.05 to -1.5m OD) preservation improved, *C. westii, C. littoris*, *T. navicularis* and *D. didyma* remained present, indicating the intertidal environment persisted. There was a high peak of *D. smithii*, which is polyhalobous, possibly indicating a higher salinity. The brackish *N. peregrina* was present with some *D. interrupta*, which are both typical of salt marsh environments (Denys, 1991/2; Vos and de Wolf, 1988). Similar to PENLD2, single valves of *A. senarius* and *C. neofastuosus* were present, as well as *S. turmida, R. arcuatum* and *Auliscus sculptus*, which are marine species (Denys, 1991/2). There were no freshwater species present in this subzone. PENLD3 represented continuing marine-brackish intertidal conditions, possibly with a higher salinity due to the absence of freshwater species, and the higher concentrations of the polyhalobous *D. smithii*.

At the onset of PENLD4 (325-276cm, -1.5 to -1.01m OD), diatom concentration declined, and preservation was poor, with some samples only counted to 100 valves. PENLD4 was dominated by

[43] organisms are able to adapt to a wide range of salinities

broken *T. navicularis*, with low concentrations of *C. westii*, *C. littoralis*, *D. didyma* and *D. smithii.* Some valves of *A. sculptus* were also present and *D. interrupta* increased slightly from PENLD3. It is possible that marine influence decreased, due to the decline of *D. smithii* and increase in the aerophilic, saltmarsh species *D. interrupta*, but that is difficult to assess due to the poor preservation and low concentrations.

In PENLD5 (276-223cm, -1.01 to -0.48m OD), diatom concentrations improved, and the zone was dominated by *D. interrupta*, with some *C. westii, T. navicularis* and *D. smithii*, which could indicate a brackish marsh, with calm conditions (Denys, 1994). However*, D. interrupta* is highly siliceous and robust to dissolution, indicating that more weakly siliceous diatoms may have been destroyed (Cooper, 1999; Flower, 1993). This zone continued across the transition from blue-grey silt and into the overlying peaty clay, indicating that brackish-intertidal conditions persisted after the transition to organic sedimentation at 232cm (-0.56m OD). Care must be taken, however, as there are very few samples in this zone, and there are no visible changes in sedimentation to salt marsh peat.

PENLD6 (223cm to the surface, -0.48m OD to +1.75m OD) represented a complete loss of diatom preservation. At 224cm (-0.48m OD), the peat was dated to 6402-6297 cal. BP (D-AMS 028035). At 214cm (-0.40m OD) and 219m (-0.45m OD) only a few broken valves were found, including *C. pusilla*. No diatoms were found between 212cm (-0.37m OD) to the surface (+1.75m OD) which implies terrestrial conditions were dominant, supported by the presence of wood macrofossil remains.

To conclude, the diatom evidence indicates a brackish-marine back barrier marsh within the basal peat between 515-492cm (-3.40 to -3.17m OD), dated to 7787-7665 cal. BP (D-AMS 028034). As sea levels rose throughout the mid Holocene an intertidal environment developed with mudflats and sand banks similar to the modern Dyfi and Mawddach estuaries. Between 380-325cm (-2.05 to -1.5m OD) there was a period of potential higher salinity, indicating a position lower in the tidal range. As the rate of sea-level rise declined, it is possible that saltmarsh developed with *D. interrupta* becoming dominant from 276cm (-1.01m OD). At 224cm (-0.48m OD), a date of 6402-6297 cal. BP

(D-AMS 028035) was recovered. In the overlying woody peat, the lack of diatoms and presence of wood macrofossil remains implies terrestrial conditions.

4.3.7 Pollen analysis

Subsamples for pollen were taken at 8cm intervals. A minimum of 250 pollen grains were counted per slide at x400 magnification. Identifications were made using Moore *et al.*'s (1991) pollen key and the University of Sheffield's reference collection. Nomenclature followed Bennett (1994). Percentages and concentrations calculated, plus diagrams created in Tilia v.2.0.2 (Grimm 2004) (Figures 4.41, 4.42,4.43 and 4.44, Table 4.40).

Table 4.32: Summary pollen zones from Penllyn

Zone	Depth (cm)	Altitude (m OD)	Pollen zone description	Environmental conditions
PENLP 1	515 to 460	-3.39 to -2.84	High *Alnus*, *Betula* and *Quercus*, *Corylus avellana*-type. Poaceae, Cyperaceae and Chenopodiaceae present. *Filipendula* spp, *Galium*-type, Asteraceae, *Urtica*, *Rumex* spp and Ranunculaceae.	A back barrier marsh with saltmarsh, fen carr, reed swamp standing freshwater, and mixed-oak woodland.
PENLP 2	460 to 280	-2.84 to -1.04	High, *Alnus* and *Betula*. *Quercus* and *C. avellana*-type declined. *Salix* increased. *Pinus* high. Poaceae high. Increase in *Filipendula* spp, *Galium*-type, Cyperaceae and Chenopodiaceae. Caryophyllaceae, Asteraceae, *Artemisia*, *Limonium vulgare*, *Plantago maritima*, *Rumex* spp, Ranunculaceae, *Taraxacum* and *Urtica*. *Plantago lanceolata* appeared at end. *Myriophyllum alterniflorum* was present. Microcharcoal high.	Saltmarsh, reed swamp and fen expanded, with mixed oak woodland and alder carr. Standing water. Some disturbed, open habitats with evidence of burning.
PENLP 3	280 to 220	-1.04 to -0.47	*Alnus* fluctuated. *Betula*, *Fraxinus* and *Pinus* declined while *Quercus* increased. *Tilia*, *Ulmus*, *Salix* and *C. avellana*-type continued. Poaceae peaked and remained high. Cyperaceae remained high. Chenopodiaceae peaked, then fell. *Filipendula* spp fell and disappeared. Asteraceae, Ranunculaceae, *Rumex* spp, *Urtica* and *Valeriana* were present. Low amounts of *P. lanceolata*, *Artemisia* and *Taraxacum* present. Microcharcoal began high, fluctuated, then fell	Mixed-oak woodland with a decline in saltmarsh and fen. Reed swamp persisted. There were some open, disturbed habitats locally with some evidence of burning.
PENLP 4	220 to 140	-0.47 to 0.36	*Alnus* up to 80%. *Quercus* continued. Other trees declined. Poaceae declined. Cyperaceae continued. Low amounts of *P. lanceolata*, very little microcharcoal.	Alder carr expanded onto the infilled marsh.
PENLP 5	140 to 49	0.36 to 1.27	*Alnus* declined, replaced by *Betula* up to 60%. *Quercus* slightly increased, *C. avellana*- type and *Salix* increased. peak in *Pteropsida*. *Tilia, Ulmus* and *Fraxinus* present. Cyperaceae high, Poaceae increased. Asteraceae, *Artemisia*, *Filipendula* spp, Ranunculaceae, *Rumex* spp, *Taraxacum, Urtica* and *P. lanceolata* present.	Alder carr replaced by birch with mixed-oak woodland , reed swamp and disturbed, open ground.
PENLP 6	49 to surface	1.27 to 1.76	*Calluna vulgaris* and *C. avellana*-type increased. All trees fell. Poaceae and Cyperaceae high. *Frankenia laevis* appeared. *P. lanceolata*, *Artemisia*, *Anthemis*-type, Asteraceae, *Filipendula* spp present. Low *Galium*-type, Ranunculaceae, *Scabiosa, Urtica Taraxacum* and *M. alterniflorum*. Microcharcoal increased	Widespread anthropomorphic impact in the landscape

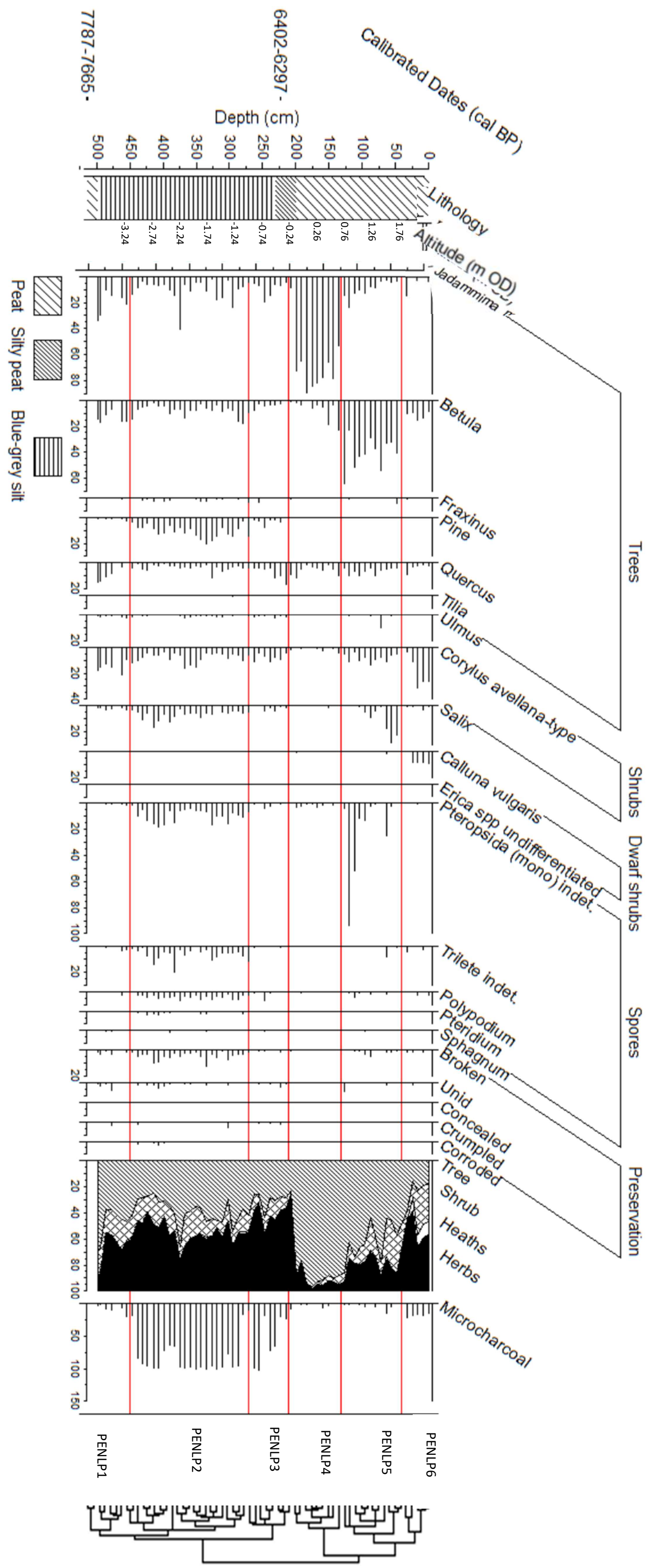


Figure 4.41: Pollen percentage diagram for Penllyn showing trees, shrubs, dwarf shrubs and spores with preservation, microcharcoal, stratigraphy and radiocarbon dates.

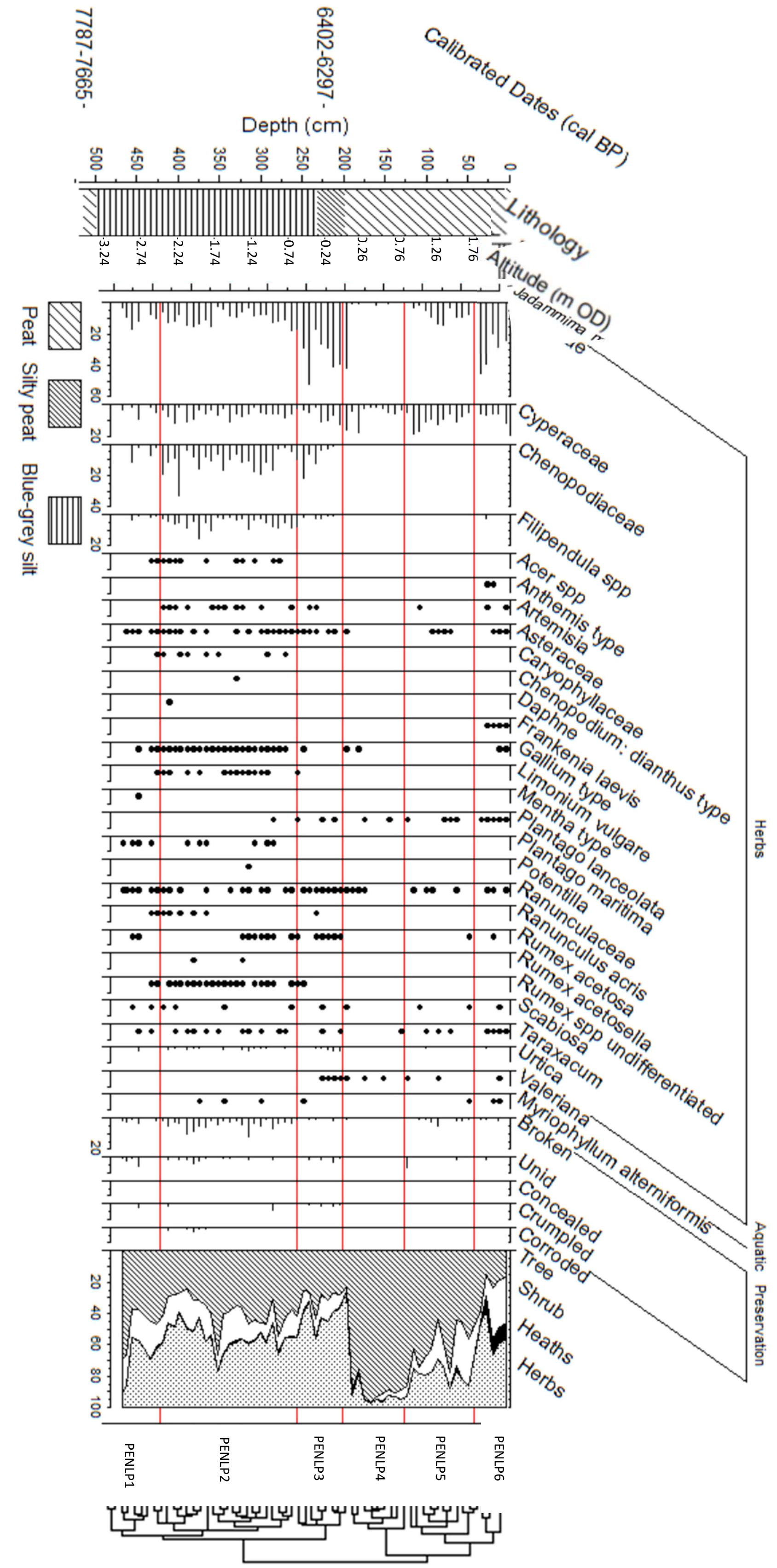
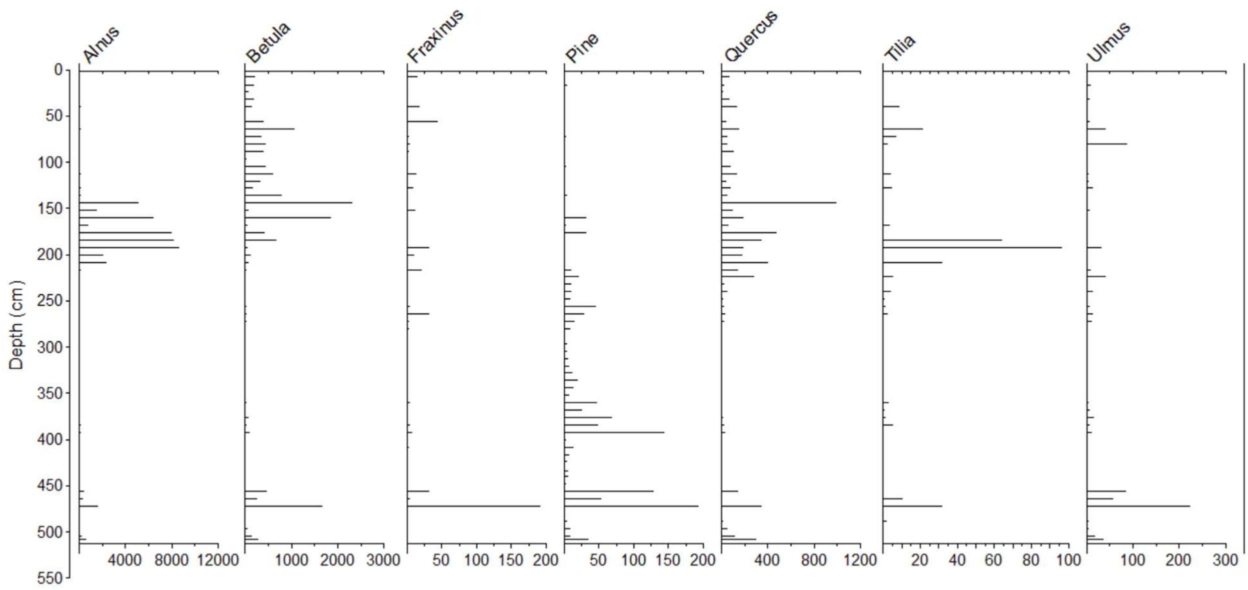


Figure 4.42: Pollen percentage diagram for Penllyn showing herbs and aquatics with preservation, stratigraphy and radiocarbon dates. Dots indicate presence below 5%



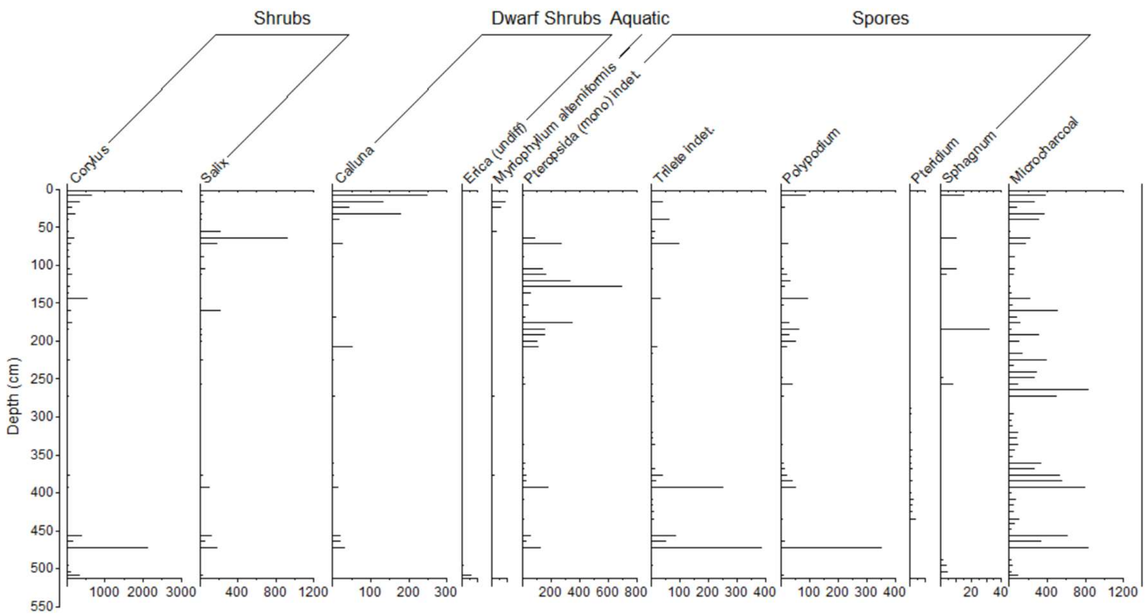


Figure 4.43: Pollen concentration diagrams for Penllyn showing trees, shrubs, dwarf shrubs, spores and aquatics with microcharcoal

Figure 4.44: Pollen concentration diagram for Penllyn showing herbs

Zone PENLP1 represents the basal peat, which was dated at 515cm (-3.39m OD) to 7787-7665 cal. BP (D-AMS 028034). PENLP1 (515 to 460cm. -3.39 to -2.84m OD) was dominated by tree pollen, in particular *Betula, Alnus* and *Quercus* as well as the dwarf-shrub *Corylus avellana*-type. There were low amounts of *Pinus, Salix, Tilia, Fraxinus* and *Ulmus*. Some *Calluna vulgaris* was present as well as limited *Erica* undifferentiated. There were high percentages of Poaceae, Cyperaceae and Chenopodiaceae, with diverse herb species such as *Filipendula* spp, *Galium*-type, Asteraceae, *Urtica, Rumex* spp and Ranunculaceae. Small quantities of *Plantago maritima*, *Scabiosa, Taraxacum* and *Mentha*-type were also present. Very few spores were present, with *Pteropsida* and Trilete spores, as well as negligible microcharcoal. Chenopodiaceae, *Plantago maritima* and *Artemisia* are associated with saltmarsh communities, while *Alnus* and *Filipendula* are associated with fen carr (Waller *et al.,* 1999; Bell 2007). Poaceae and Cyperaceae are associated with fen, saltmarsh, and bog communities (Waller *et al.,* 1999; Bell, 2007). *Mentha*-type has been associated with unshaded freshwater conditions alongside *Rumex obtusifolius*-type (Waller *et al.,* 1999). This indicates a diverse back barrier marsh was present, with saltmarsh, fen carr and reed swamp as well as some standing freshwater, and mixed-oak woodland in the valley uplands.

Through the blue-grey silt, the start of PENLP2 (460 to 280cm -2.84 to -1.04m OD) saw a reduction in pollen preservation and concentration. The proportion of *Alnus* and *Betula* fluctuated but remained high. *Quercus* and *C. avellana*-type declined while *Salix* increased. The proportion of *Pinus* was high but poorly preserved; overrepresentation of *Pinus* is common in marine sediments as the pollen grains float (Hopkins, 1950; Caseldine, 1992, 2000; Scaife, 1993, 1994). *Tilia, Ulmus* and *Fraxinus* were present in low amounts. *C. vulgaris* was present throughout the zone, with large quantities of spores, such as *Polypodium, Pteridium* and *Sphagnum*. Poaceae remained high, with a dramatic increase in *Filipendula* spp, *Galium*-type, Cyperaceae and Chenopodiaceae. Herbs remained diverse, with Caryophyllaceae, Asteraceae, *Artemisia, L. vulgare*, *P. maritima, Rumex* spp, Ranunculaceae, *Taraxacum* and *Urtica*. *Plantago lanceolata* appeared at the top of zone. The aquatic *Myriophyllum alterniflorum* was present indicating the presence of standing water, with microcharcoal high throughout. Care must be taken, however, as the spike in microcharcoal could be caused by the change in the catchment area due to the transition from peat to silt, rather than a change in the use of fire. An expansion of saltmarsh is reflected in the increase in Chenopodiaceae, as well as Asteraceae, Caryophyllaceae, *Ligustrum vulgare*, and *P maritima*. A further expansion of reed swamp and fen is reflected in the presence of *Filipendula*, *Gallium*-type, Poaceae and Cyperaceae as well as *Betula* and *Alnus* (Bell, 2007; Waller *et al.,* 1993, 1999). *Betula* can be found in poor fen and acidic conditions, with *Alnus* indicating the presence of fen carr in the area (Waller *et al.,* 1993, 1999). Herbs common to disturbed, open habitats were present such *as P. lanceolata, Rumex* sp., Ranunculaceae, *Urtica, Artemisia* and *Taraxacum*, which may reflect a diverse woodland edge community (Bell, 2007).

As the blue-grey silt transitioned to silty peat*,* the onset of PENLP3 (280 to 220cm, -1.04 to -0.47m OD) marked a decline in *Betula, Fraxinus* and *Pinus* while *Quercus* increased. *Alnus* fluctuated, while *Tilia, Ulmus, Salix* and *C. avellana*-type continued. Poaceae peaked and remained high through the zone. Cyperaceae remained high while Chenopodiaceae peaked, then fell to the end of the zone. *Filipendula* spp fell, disappearing by the end of the zone. These changes reflect the decline of saltmarsh and fen in the surrounding area, with some reed swamp possibly persisting (Bell, 2007). Some Asteraceae, Ranunculaceae, *Rumex* spp, *Urtica* and *Valeriana* were present. Very low amounts of *P. lanceolata*, *Artemisia* and *Taraxacum* were present, indicating some open, disturbed habitats locally. Herbs were less diverse with the absence of *Gallium*-type, Caryophyllaceae and *P. maritima*. Spores declined, while microcharcoal began high, fluctuated, then fell at the top of the zone.

As the silty peat transitioned to peat, preservation improved at the onset of PENLP4 (220 to 140cm. -0.47 to 0.36m OD). At 224cm (-0.48m OD), the peat was dated to 6402-6297 cal. BP (D-AMS 028035). *Alnus* expanded to dominate up to 80% of the profile reflecting *Alder* carr colonising the infilling marsh. *Quercus* continued, while other trees declined. Herbs were much less diverse, and Poaceae dramatically declined. Cyperaceae continued, while Chenopodiaceae, *Filipendula* spp, *Gallium*-type, Ranunculaceae. and *Rumex* spp largely disappeared, implying a reduction in saltmarsh, reed swamp and fen, however, care must be taken as *Alnus* produces a high quantity of pollen and may be skewing the pollen profile (Waller *et al.,* 1999). This can be seen from the pollen concentration diagrams as the concentration of Alnus peaks to very high levels, while all other species fall. Some very low amounts of *P. lanceolata* were present, with few spores and very little microcharcoal.

On the onset of PENLP5 (140 to 49cm 0.36 to 1.27m OD) *Alnus* rapidly declined, replaced by *Betula*, which dominated up to 60% of the profile, indicating drier conditions persisted (Waller *et al.,* 1999).

Quercus increased slightly, while *C. avellana*- type and *Salix* rapidly increased. *Tilia, Ulmus* and *Fraxinus* were present in low amounts. Cyperaceae remained high, while Poaceae rapidly increased. Asteraceae, *Artemisia, Filipendula* spp, Ranunculaceae, *Rumex* spp, *Taraxacum* and *Urtica* were present as well as *P. lanceolata*. It is possible reed swamp and fen carr persisted in the local environment, although no evidence of saltmarsh was found. There was a sharp peak in *Pteropsida* spores, while microcharcoal remained low.

In PENLP6 (49cm to surface 1.27 to 1.76m OD) *C. vulgaris* increased sharply, with very high Poaceae and Cyperaceae. *C. avellana*-type increased up to 40 % while all other trees declined. *P. lanceolata, Artemisia, Anthemis*-type, Asteraceae, *Filipendula* spp were present with low amounts of *Galium*-type, Ranunculaceae, *Scabiosa* and *Urtica*. *Frankenia laevis* appeared, and there was a small peak in *Taraxacum* and *M. alterniflorum*. Microcharcoal increased slightly through the zone. The sharp drop in tree pollen, combined with increase in disturbance indicators such as *P. lanceolata* and the presence of microcharcoal indicates anthropomorphic impact in the landscape (Turner, 1964; Dumayne-Peaty, 1998).

To conclude, the basal peat (515 to 460cm, -3.39 to -2.84m OD) was characterised by a high level of *Alnus* with evidence for a mixed-*Quercus* woodland. The peat at 515cm (-3.39m OD) was dated to 7787-7665 cal. BP (D-AMS 028034) at 515cm (-3.39m OD). Saltmarsh, reed swamp and tall herb fen was present indicated by Cyperaceae and Chenopodiaceae. *Filipendula* spp, *Galium*-type with diverse woodland edge herbs such as Asteraceae, *Urtica*, *Rumex* spp and Ranunculaceae. On the transition to silt between 460 to 280cm (-2.84 to -1.04m OD) preservation decline with *Pinus* becoming dominant indicating transportation of buoyant grains in estuarine silt. Evidence of saltmarsh, fen and reed swamp increases with high levels of Poaceae, *Filipendula* spp, *Galium*-type, Cyperaceae and Chenopodiaceae as well as Caryophyllaceae, Asteraceae, *Artemisia*, *Limonium vulgare*, *Plantago maritima* and diverse herbs. From 280cm (-1.04m OD) tree pollen fluctuated with an increase in saltmarsh, fen and reed swamp indicated by Poaceae, Cyperaceae, Chenopodiaceae and *Filipendula* before all but Poaceae declined towards 220cm (-0.47m OD). At 224cm (-0.48m OD), the peat was dated to 6402-6297 cal. BP (D-AMS 028035). On the transition to peaty silt at 224cm (-0.48m OD) *Alnus* became dominant, replaced by *Betula* at 140cm (0.36m OD) with evidence for mixed *Quercus* woodland. From 49cm to the surface (1.27 to 1.76m OD) trees declined with evidence for human impact indicated by the presence of *P. lanceolata*, *Artemisia*, *Anthemis*-type, Asteraceae, Ranunculaceae, *Scabiosa, Urtica, Taraxacum* and microcharcoal.

4.4 Statistical Analysis

4.4.1 Transfer functions

4.4.1.1 Transfer functions developed.

Three proxies were used in this study, with diatoms and foraminifera suitable for transfer function application. Due to constraints of time, a local training set could not be developed. The diatom training sets of Zong and Horton (1999) and Hill *et al.,* (2007) were used, as well as the foraminifera training sets of Horton and Edwards (2006) and Rushby *et al.,* (2019). These training sets have been combined into a multiproxy training set. Ideally, a multiproxy training set would have both diatoms and foraminifera from the same sites and samples, but this was not available.

Samples and species with low counts were removed from the training set and an initial model was run using WA-PLS. The training sets were then pruned, with samples removed which had residuals greater than the standard deviation of the whole range of SWLI represented in the training set (Edwards *et al.,* 2004; Gehrels *et al.,* 2005; Horton and Edwards, 2006). By pruning the dataset, the model precision and performance is improved, as samples with poor relationship to elevation can skew the results (Horton and Edwards, 2006). Alternatively, some researchers argue that it is beneficial to leave all samples in the training set, as it better reflects the full range of modern environments (Hill *et al.,* 2007). For this reason, it was decided to compare the results of each model by firstly leaving all samples in the training set, and secondly pruning samples based on the residuals. Each training set is outlined in Table 4.33 below – to summarise they are as follows:

- D-A – Diatom training set with all samples / unpruned dataset
- D-P – Diatom training set with samples pruned based on residuals
- F-A – Foraminifera training set with all samples / unpruned dataset
- F-P – Foraminifera training set with samples pruned based on residuals
- M-A – Multiproxy training set with all samples / unpruned dataset
- M-P – Multiproxy training set with samples pruned based on residuals

Table 4.33: Summary of transfer functions developed including proxies used, number of sites and samples and whether the samples were pruned.

Name	Proxy	Pruned or unpruned	Sites	Samples	Sources
D-A	Diatoms	All samples	6	88	Zong and Horton 1999 Hill *et al.,* (2007)
D-P	Diatoms	Pruned	6	82	Zong and Horton 1999 Hill *et al.,* (2007)
F-A	Foraminifera	All samples	14	280	Horton and Edwards 2006, Rushby *et al.,* 2019
F-P	Foraminifera	Pruned	14	247	Horton and Edwards 2006, Rushby *et al.,* 2019
M-A	Multiproxy	All samples	21	367	Zong and Horton 1999 Hill *et al.,* (2007) Horton and Edwards 2006, Rushby *et al.,* 2019
M-P	Multiproxy	Pruned	21	351	Zong and Horton 1999 Hill *et al.,* (2007) Horton and Edwards 2006, Rushby *et al.,* 2019

A number of training sets were used for this research. Zong and Horton (1999) developed a diatom training set from six sites across the UK. The closest site to the Dysynni Valley was Roundsea Marsh on the west coast of England. DCCA analysis produced a gradient length of 3.84 standard deviation units, which indicated that unimodal methods were suitable. Only diatom species that accounted for at least 5% total diatom valves were included, leaving 73 taxa out of 153.

Canonical correspondence analysis (CCA) and partial CCAs were used to determine the relationship between the diatom abundance and the environmental variables in the training set, and a Monte

Carlo permutation test determined whether the results were statistically significant. The CCA results showed that axes one and two explained 12.9% of the total variance in the data, while the two axes represented 59.7% of the species-environment relationship (Zong and Horton, 1999). SWLI accounted for 23.3% of the explained variance, as well as organic content (19.7%) sand content (5.2%), silt (5.2%) and clay (5.1%). Intercorrelation between variables explained 41.5% of variation and 78% of the total variance remained unexplained. Zong and Horton (1999) determined that although a robust transfer function could be developed, the models could not be considered completely independent of other variables.

Hill *et al.,* (2007) produced a diatom training set for the macrotidal Severn Estuary. Two sites were sampled in order to cover the entire tidal frame. DCCA analysis confirmed a unimodal model would be suitable, and CCA analysis revealed that axes one and two explained 16.4% of the total variance. A total of 68.6% of the species-environment relationship was represented and the environmental parameters accounted for 25% of the explained variation. Partial CCAs revealed that 27.2% of explained variation was characterised by altitude and 27% by flooding duration, as well as organic matter (10.7%), clay fraction (5.2%), silt fraction (5.1%) and sand fraction (5%) (Hill *et al.,* 2007). The remaining 19.8% of explained variation represented intercorrelation between variables and 75% of the variation could not be explained.

Horton and Edwards (2006) produced a foraminifera training set from 15 sites across the UK. There were, however, no sites in Wales and the closest site to the Dysynni Valley was Roundsea Marsh. Although the training set included some sites used by Zong and Horton (1999), such as Roundsea Marsh, the transects and samples were not the same. DCCA analysis revealed an axis length of 3.16 standard deviation units, therefore unimodal methods were suitable. CCA indicated that axes one and two represented 71% of the total variance, with 93% of species-environment relationship explained. Six environmental variables were assessed, which accounted for 76% of the explained variance including elevation (42%), salinity (17%), organic content (14%), pH (14%) and clay (5%).

The associated Monte Carlo permutation tests showed that all variables except clay were highly significant. Species were only included if they accounted for at least 3% in one sample, resulting in 110 samples and 19 species (Horton and Edwards 2006).

A foraminifera training set for the Maltreath marshes, Anglesey, has also been developed, including 48 samples from a 218m transect (Rushby *et al.*, 2019). DCCA analysis revealed a unimodal response was suitable, but details of DCCA and CCA analyses were not published (Rushby *et al.,* 2019).

Once developed, the transfer function must be evaluated. Transfer functions are initially evaluated based on the RMSEP and R^2 values, and the best component out of 5 is chosen. A good model will have a low RMSEP value out of 100 and a high R^2_{boot} out of 1.

Table 4.34: Summary of transfer function model evaluation in terms of the number of components, the RMSEP and the R^2_{boot}

Name	Proxy	Pruned or unpruned	Component	RMSEP	R^2_{boot}	Sources
D-A	Diatoms	All samples	2	22.884	0.643388	Zong and Horton 1999
D-P	Diatoms	Pruned	1	19.78166	0.649398	Zong and Horton 1999
F-A	Foraminifera	All samples	1	59.6199	0.20007	Horton and Edwards 2006, Rushby *et al.,* 2019
F-P	Foraminifera	Pruned	2	31.11894	0.483432	Horton and Edwards 2006, Rushby *et al.,* 2019
M-A	Multiproxy	All samples	1	62.0681	0.526723	Zong and Horton 1999 Horton and Edwards 2006, Rushby *et al.,* 2019
M-P	Multiproxy	Pruned	2	34.77961	0.769853	Zong and Horton 1999 Horton and Edwards 2006, Rushby *et al.,* 2019

The diatom-only transfer function gives a very precise model with good values of root mean square error of prediction (RMSEP) and coefficient of determination (R^2_{boot}). The relationship between the fossil and modern samples must be evaluated (Chapter 4.4.1). Initially, component 2 was chosen based on an improvement in RMSEP of 5.2%.

Table 4.35: Model precision statistics for the first 5 components of the unpruned diatom transfer function. RMSE and R^2 are produced by the transfer function, Boot_ R^2 and RMSEP is produced by bootstrapping cross validation.

	Component 1	Component 2	Component 3	Component 4	Component 5
RMSE	20.79806	16.04301	11.56891	10.05512	9.384468
R^2	0.660147	0.797761	0.894832	0.920544	0.93079
Boot_ R^2	0.566078	0.643388	0.689229	0.715205	0.727608
RMSEP	24.16278	22.884	22.74427	22.67723	23.75854
%Change	0	5.292361	0.610575	0.294771	-4.76828

Pruning the training set based on residuals made little difference to the model precision, with a very slight improvement in RMSEP. After pruning, component 1 was chosen as there was only an improvement in RMSEP of 1.5% for the second component.

Table 4.36: Model precision statistics for the first 5 components of the pruned diatom transfer function. RMSE and R^2 are produced by the transfer function, Boot_ R^2 and RMSEP is produced by bootstrapping cross validation.

	Component 1	Component 2	Component 3	Component 4	Component 5
RMSE	16.73636	13.33924	11.00337	9.464012	8.68125
R^2	0.729595	0.828215	0.88312	0.913528	0.92724
Boot_ R^2	0.649398	0.689475	0.711273	0.715514	0.709445
RMSEP	19.78166	19.47299	20.20867	21.30663	23.11026
%Change	0	1.560387	-3.77797	-5.4331	-8.46513

The foraminifera-only model with all species produces a poor model in terms of precision, with an RMSEP of 59.6199 and an R^2_{boot} of 0.20007. The first component was chosen due to a decrease of 2.1% for the second component.

Table 4.37: Model precision statistics for the first 5 components of the unpruned foraminifera transfer function. RMSE and R^2 are produced by the transfer function, Boot_ R^2 and RMSEP is produced by bootstrapping cross validation.

	Component 1	Component 2	Component 3	Component 4	Component 5
RMSE	57.66061	54.30492	52.36916	51.68947	50.99054
R^2	0.236965	0.323201	0.370577	0.386821	0.403292
Boot_ R^2	0.20007	0.230238	0.179592	0.14773	0.124411
RMSEP	59.6199	60.88888	73.78151	86.85842	101.1136
%Change	0	-2.12846	-21.174	-17.7238	-16.4119

There was significant improvement in precision after pruning the training set, with an RMSEP of 31.11894 and R^2boot of 0.353328. The second component was chosen based on an improvement of 8.1% RMSEP.

Table 4.38: Model precision statistics for the first 5 components of the pruned diatom transfer function. RMSE and R^2 are produced by the transfer function, Boot_ R^2 and RMSEP is produced by bootstrapping cross validation.

	Component 1	Component 2	Component 3	Component 4	Component 5
RMSE	30.72034	27.03918	25.77437	25.31107	24.78846
R^2	0.452931	0.576173	0.614873	0.62858	0.643745
Boot_ R^2	0.353328	0.483432	0.520075	0.491878	0.41131
RMSEP	33.89007	31.11894	30.62932	32.96223	39.90195
%Change	0	8.176832	1.573386	-7.61658	-21.0536

The multiproxy model before pruning had reasonable performance with 0.526723 R^2_{boot} and 62.0681 RMSEP. Component 1 was chosen as there is only an improvement of 2.3% RMSEP.

Table 4.39: Model precision statistics for the first 5 components of the unpruned multiproxy transfer function. RMSE and R^2 are produced by the transfer function, Boot_ R^2 and RMSEP is produced by bootstrapping cross validation.

	Component 1	Component 2	Component 3	Component 4	Component 5
RMSE	60.63546	55.9359	54.25828	52.51239	51.86973
R^2	0.544005	0.611937	0.634875	0.657986	0.666306
Boot_ R^2	0.526723	0.56721	0.529236	0.43687	0.378607
RMSEP	62.0681	60.59274	66.02126	81.01218	93.32686
%Change	0	2.376992	-8.95903	-22.7062	-15.201

Pruning based on residuals improved model performance considerably, with an R^2boot value of 0.769853 and RMSEP of 34.77961. The second component was chosen based on an improvement of 6.08% RMSEP.

Table 4.40: Model precision statistics for the first 5 components of the pruned multiproxy transfer function. RMSE and R^2 are produced by the transfer function, Boot_ R^2 and RMSEP is produced by bootstrapping cross validation.

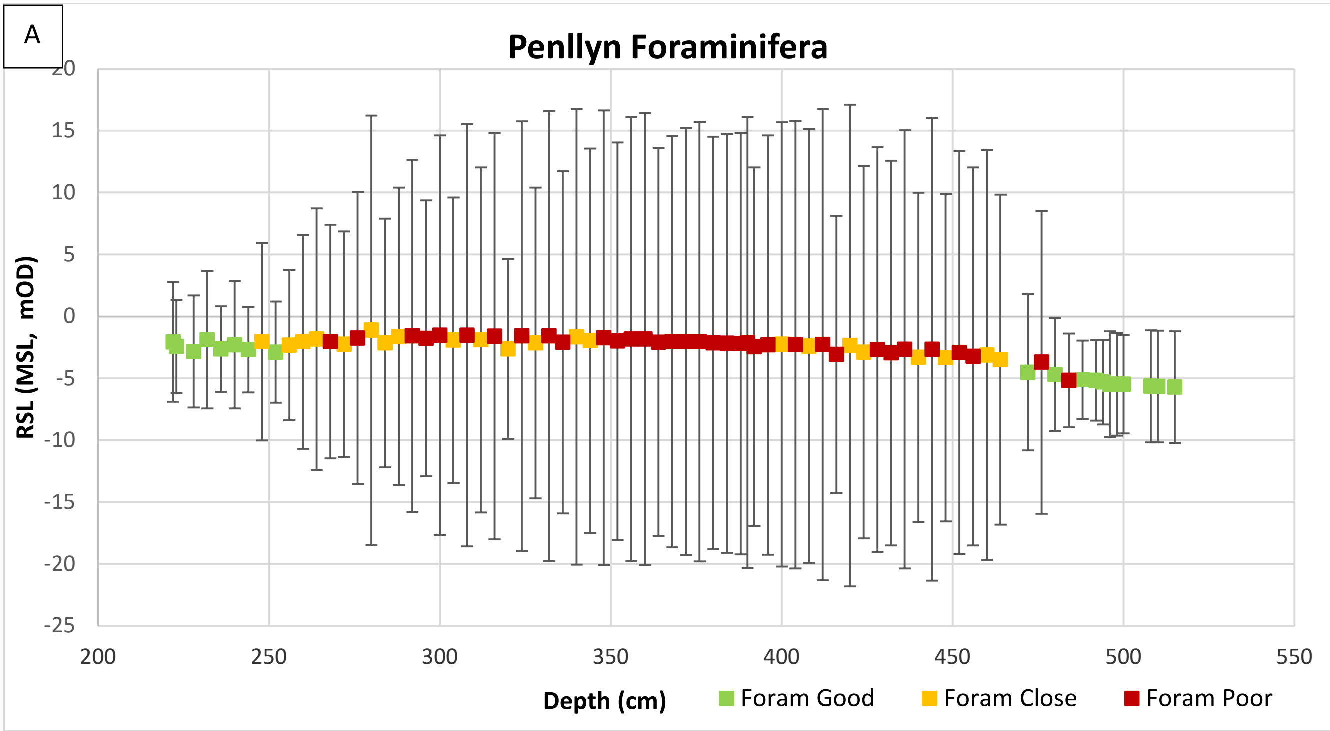
	Component 1	Component 2	Component 3	Component 4	Component 5
RMSE	35.61914	31.38561	29.37889	28.21292	27.69854
R^2	0.747087	0.803618	0.827934	0.841312	0.847045
Boot_ R^2	0.72947	0.769853	0.789597	0.788721	0.763909
RMSEP	37.03475	34.77961	34.04601	35.19153	38.71087
%Change	0	6.089269	2.109282	-3.36464	-10.0005

The most precise model in terms of RMSEP is D-P (RMSEP 19.78166, R^2 boot 0.649398). The best model in terms of R^2_{boot} is M-P (RMSEP 34.77961, r2 boot 0.769853). Model F-P is the worst in terms of R^2_{boot} (0.483432) but reasonable RMSEP (31.11894). Every pruned model gives an improvement in RMSEP, and most models give an improvement of R^2_{boot}. The pruned diatom model only improves marginally, with a similar R^2_{boot} (0.643388 and 0.649398) and RMSEP reduced from 22.884 to 19.78166. The first component must be used, however, which will equate to a WA model, not WA-PLS. Pruning is most effective with the foraminifera and multiproxy models, which are very poor in terms of precision without pruning. In each case the second component can be used after pruning which increases model performance using residuals. The foraminifera model increases R^2_{boot} from 0.20007 to 0.483432 and RMSEP decreases from 59.6199 to 31.11894. The multiproxy model R^2_{boot} increases from 0.526723 to 0.769853 and RMSEP decreases from 62.0681 to 34.77961. All pruned models will be used, and results compared in terms of reconstruction and modern analogues (see Chapters 4 and 5)

Both fossil diatoms and foraminifera were available in the Penllyn core, therefore all three transfer functions were employed including foraminifera, diatom and multiproxy. The modern analogue technique (MAT) was used to establish how well the fossil core reflected assemblages found in the modern training set (Chapter 3.4). For each fossil sample MAT determines if there are good, close, or poor analogues in the modern training set. If there are no good or close analogues, then the results from the transfer function cannot be used.

4.4.1.2 Foraminifera Transfer Function

The foraminifera transfer function appears to show an overall rise in sea-level from -6 to -1.5m OD. From the base of the core, sea level rose from -6m to -3m OD, followed by a steadier rise in sea-level to -1.5m OD. This was followed by an apparent fall in sea level to between -2 and -3m OD before freshwater conditions dominated the core with the transition to peat.



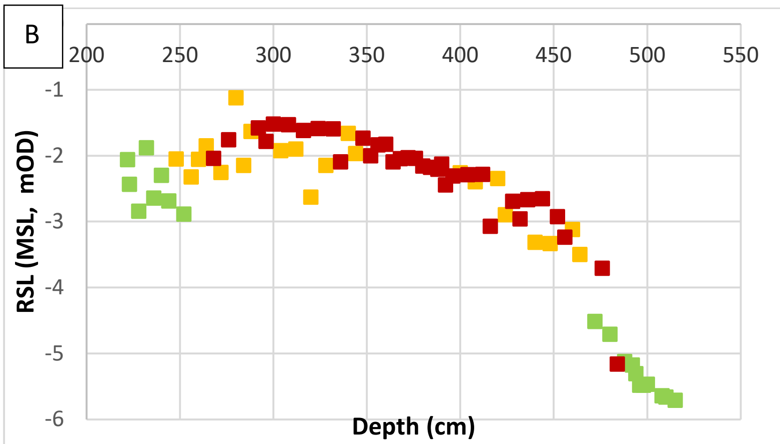


Figure 4.45: A: Graph showing the reconstruction from the Penllyn foraminifera transfer function against depth with standard error bars.

B: The same results, with no error bars on a shorter y-axis scale

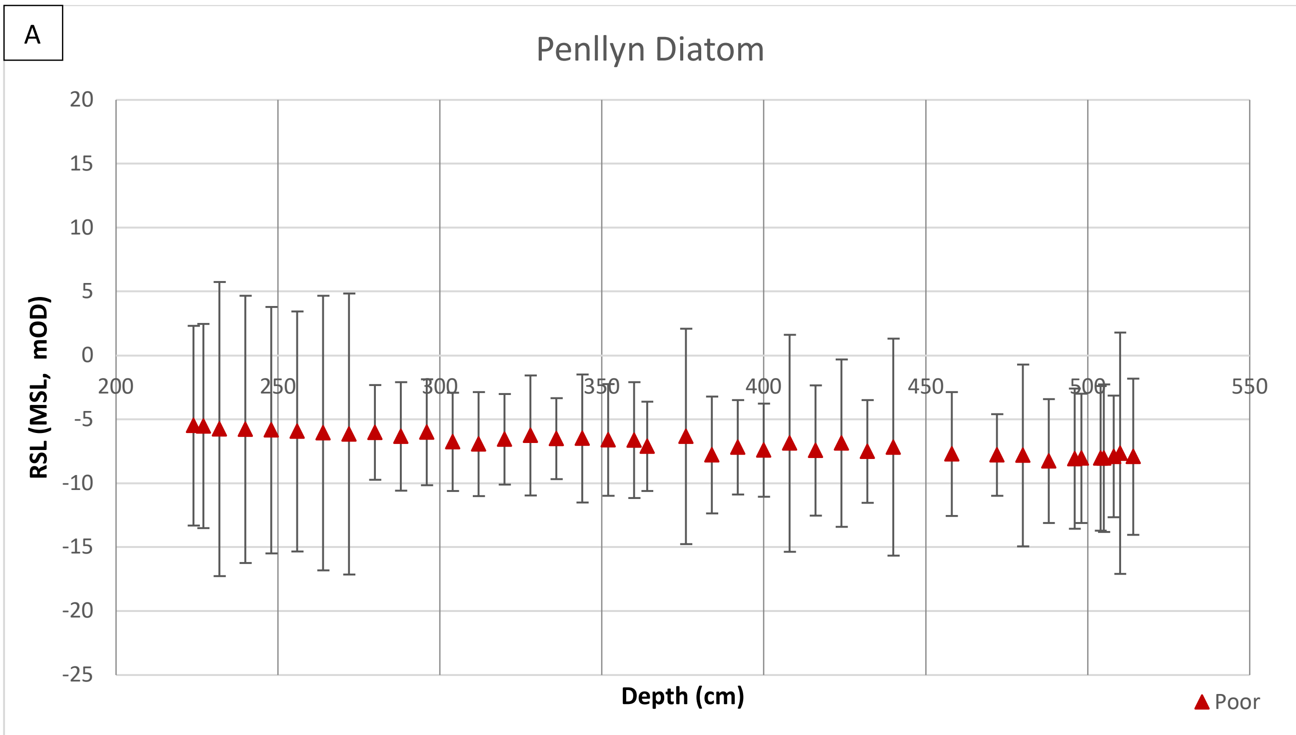
The samples correlating to the high marsh, basal peat (from 515 to 472cm) have very good analogues in the modern training set with very narrow error ranges. This relates to zones PENLF1 and PENLF2 of the foraminifera diagram (515 to 470cm, -3.39 to -2.94m OD) where a monospecific *J. macrescens* assemblage transitioned to a high marsh assemblage at the onset of blue-grey silt with dominant *J. macrescens* as well as *T. inflata* and *M. fusca*. High marsh species have narrow tolerance ranges and are often considered more accurate sea-level indicators than low marsh species (Scott and Medioli, 1978; Gehrels 1999, Wright *et al.,* 2011, Horton and Edwards, 2006). High marsh assemblages of *J. macrescens, T. inflata* and *M. fusca* are well represented in the training set which has resulted in good analogues with the fossil samples.

Samples between 464 and 256cm have both poor and close analogues with very wide errors, correlating with PENLF3 (470-270cm, -2.94 to -0.94m OD). In PENLF3 the calcareous *Ammonia* spp. was dominant with very few other calcareous species. In the modern training set, low marsh, calcareous assemblages are very diverse and samples with dominant *Ammonia* spp. are not well represented. There is very little correlation between the fossil samples of zone 3 and the modern training set, which may be due to poor preservation, as calcareous foraminifera are known to dissolve upon burial (Gehrels *et al.,* 2001, Horton and Edwards, 2006). Furthermore, low marsh, calcareous species have wide tolerance ranges which may also be reflected in the much wider standard errors.

Zone 4a (270-221cm, -0.94 to -0.44m OD) marked a return to high marsh conditions with dominant *J. macrescens* and low levels of high marsh species *T. inflata, M. fusca* and *Haplophragmoides* spp. This is reflected in the transfer function as there are good analogues and narrow error ranges between 252 and 223cm. There is, however, a wider scatter in the results compared to the high marsh samples in the basal peat which may reflect that sea-level retreat was not linear.

4.4.1.3 Diatom Transfer Function

The diatom transfer function appears to show a rising sea level from -8.5 to -5.5m OD, which is significantly lower than the foraminifera reconstruction. There is a wide scatter in the reconstruction between 440 and 304cm, although there are narrower error bars. The standard error is greater at the extremes, particularly between 480 - 414 cm and 272-224cm.



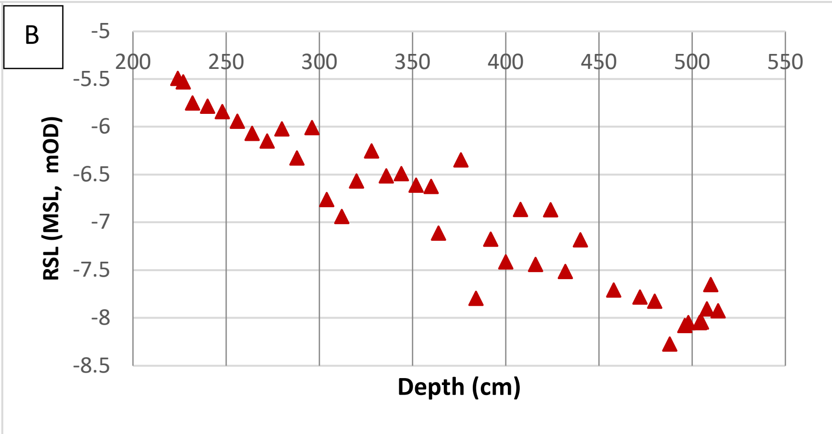


Figure 4.46: A: Graph showing the reconstruction from the Penllyn diatom transfer function against depth with standard error.
B: No error bars and enlarged to show trends.

Despite a very accurate model in terms of RMSEP and R^2_{boot}, there are no close or good analogues at all for the diatom transfer function, which makes the data unusable. The poor analogues may explain the significant offset between the reconstructions provided by the diatom and foraminifera transfer functions. The transfer function may have been impacted by preservation within the fossil core. For example, between 272 and 224cm (diatom zone PENLD5) the samples were dominated by the robust species *D. interrupta*, implying poor preservation, which correlates with a wider standard error in the transfer function. Furthermore, diatom communities are very diverse, and a local training set may be needed for future sea-level reconstructions. However, even with a poor modern analogues, the diatom assemblage has great value which highlights the need for a local transfer function to produce accurate reconstructions.

4.4.1.4 Multiproxy Transfer Function

The reconstruction from the multiproxy transfer function has been adversely affected by the sampling intervals of the two proxies: diatoms were sampled at 8cm intervals while foraminifera were sampled at 4cm throughout the fossil core. The samples which contain both diatoms and foraminifera have poor analogues and plot much lower than the samples with only foraminifera. The training set did not contain sites and samples with both proxies, which would be ideal. This transfer function cannot be used for reliable reconstructions, and in future, a local multiproxy training set will be needed which has the same sites and samples containing both proxies.

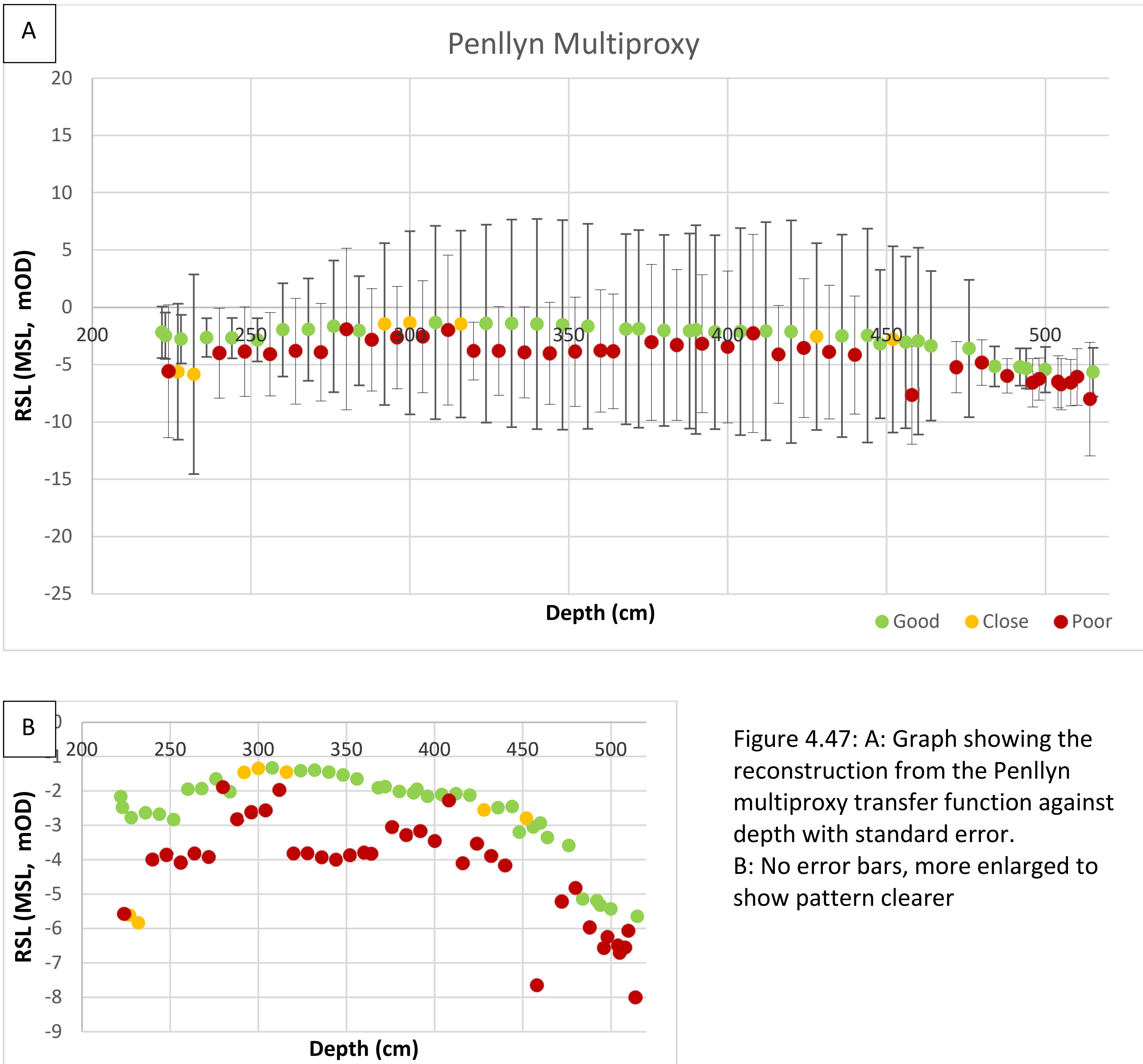


Figure 4.47: A: Graph showing the reconstruction from the Penllyn multiproxy transfer function against depth with standard error.
B: No error bars, more enlarged to show pattern clearer

4.4.2 Sea Level Index Points

4.4.2.1 Indicative meaning

To calculate sea-level index points the altitude, indicative range, tendency, and age are needed (Chapter 3.4). Two sea-level index points were calculated for the dated horizons at Penllyn. The calibrated median age, as well as the positive and negative range were established using Oxcal Online v4.4, calibration curve IntCal20 (Reimer *et al.,* 2016*,* 2020).

Table 4.41: Radiocarbon dates from Penllyn, with calibrated ages and range calculated using Oxcal Online v4.4

Depth (cm)	Elevation (m OD)	Radiocarbon age		Calibrated Age (cal. BP, median)	Positive error range	Negative error range
		BP	1 σ error			
224	-0.49	5558	30	6346.00	53.00	50.00
514	-3.39	6881	28	7708.00	82.00	84.00

As both foraminifera and diatoms were present in the core, all three transfer functions were applied. The indicative meaning was established for the SLIPs from the microfossil analyses. The diatom and foraminifera results indicated a high marsh environment for both dated horizons, which Hijma *et al.,* (2015) places between highest astronomical tide (HAT) and mean tide level (MTL). Following Hijma *et al.,* (2015), therefore the indicative meaning of both SLIPs was calculated as (HAT+MTL)/2.

4.4.2.2 Errors

Errors were established following Smith *et al.,* (2020) and Hijma *et al.,* (2015) (Table 4.42). Compaction was established by multiplying the depth to compressible substrate by a factor of 2.5 following Hijma *et al.,* (2015). Compaction corrections were not applied to the basal SLIP as it was deposited on gravel, which was recovered in the gouge core. A tide model was not used, instead the SLIPs were plotted against mean high water spring tide (MHWST) following Smith *et al.,* (2020) as this gives a good reflection of tidal change within an estuary (Chapter 3.4).

Table 4.42: Error breakdown with elevation, positive and negative errors for both SLIPs (-0.49 and -3.39m OD) for each proxy (diatom, foraminifera, multiproxy and microfossil).

	Elevation (m OD)	Positive error	Negative error	Sample thickness (m)	Sampling (m)	Core shortening /stretching (m)	dGPS	Non-vertical drilling (m)	Compaction
Diatom	-0.49	1.89	1.31	0.01	0.01	0.01	0.10	0.45	1.43
	-3.39	1.61	1.25	0.01	0.01	0.01	0.10	1.03	0.00
Foram	-0.47	1.75	1.09	0.01	0.01	0.01	0.10	0.45	1.43
	-3.39	1.50	1.10	0.01	0.01	0.01	0.10	1.03	0.00
Multi	-0.47	1.76	1.10	0.01	0.01	0.01	0.10	0.45	1.43
	-3.39	1.51	1.12	0.01	0.01	0.01	0.10	1.03	0.00
Microfossil	-0.49	2.03	1.50	0.01	0.01	0.01	0.10	0.45	1.43
	-3.39	1.17	0.59	0.01	0.01	0.01	0.10	1.03	0.00

4.4.2.3 Sea-level index points

The results of each SLIP can be seen in Table 4.43 and Figure 4.48 a, b and c. The multiproxy SLIPs will not be considered further, transfer function was adversely affected by the sampling intervals of the two proxies.

Table 4.43: Summary SLIPs with elevation, RSL, error and calibrated ages for two SLIPs (at-0.49 and -3.39m OD) for each proxy (diatom, foraminifera, multiproxy and microfossil).

	Elevation (m OD)	RSL (m OD MHWST)	+ error	- error	Calibrated Age (cal. BP, median)	age + error	age - error
Diatom	-0.49	-6.00	1.89	1.31	6346.00	53.00	50.00
	-3.39	-8.42	1.61	1.25	7708.00	82.00	84.00
Foram	-0.47	-2.88	1.75	1.09	6346.00	53.00	50.00
	-3.39	-6.16	1.50	1.10	7708.00	82.00	84.00
Multiproxy	-0.47	-2.78	1.76	1.10	6346.00	53.00	50.00
	-3.39	-6.02	1.51	1.12	7708.00	82.00	84.00
Microfossil	-0.49	-4.91	2.03	1.50	6346.00	53.00	50.00
	-3.39	-7.82	1.17	0.59	7708.00	82.00	84.00

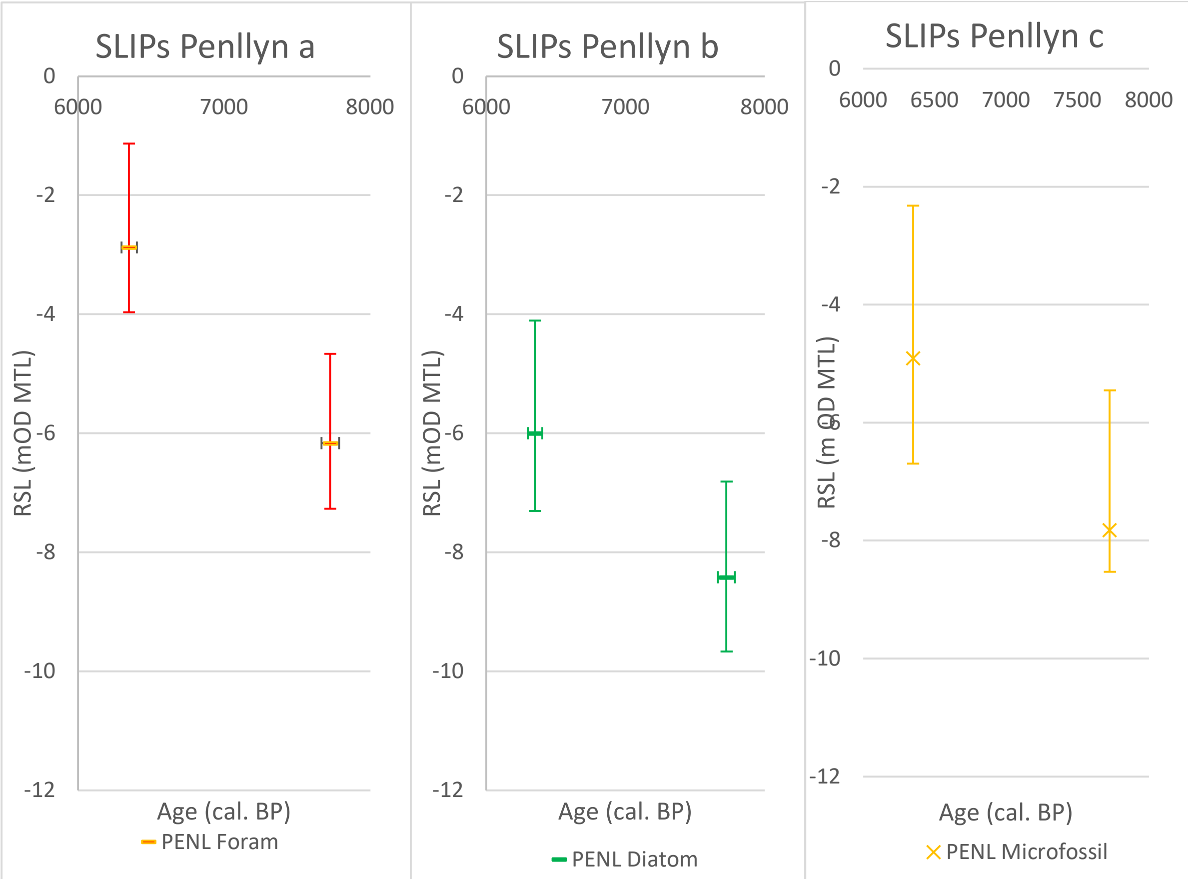


Figure 4.48 a, b, and c: SLIP graph with foraminifera, diatom and microfossil SLIPS plotted on separate axes.

The SLIP results all show a rise in sea level from 7787-7665 cal. BP to 6402-6297. For the foraminifera, sea-level rose from -6.16 to -2.88m OD MHWST, however the diatom SLIP shows a rise from -8.42 to -6.00 m OD. The indicative meaning calculated from microfossils shows a rise from-7.82 to -4.91m OD, similar to the diatoms. It is unclear why foraminifera and diatoms have plotted so differently, and it is challenging to choose which SLIP to use. The foraminifera transfer function has the best modern analogues; however, it plots consistently higher than both the diatom and microfossil SLIPs. The indicative meaning determined from microfossil evidence plots very close to the diatom transfer function, implying the diatom results could be the most accurate. Care must be taken when using the results of the diatom transfer function, however, as there are no modern analogues in the training set which makes the transfer function unreliable. There may also be preservation problems within the silt. For example, between 276-223cm (-1.01 to -0.48m OD) the diatom assemblage is dominated by *D. interrupta* which is a very robust species and may indicate more delicate diatoms have been lost. Future work is needed to produce a multiproxy, local training set in order to produce SLIPs and transfer functions which have good modern analogues for diatoms.

4.5 Environmental change at Penllyn

The basal peat (515-500cm, -3.39 to -3.24 m OD) was dated at 515cm (-3.39m OD) to 7787-7665 cal. BP (D-AMS 028034). The diatom evidence from the basal peat suggested a back-barrier saltmarsh or lagoon had developed at Penllyn, indicated by the presence of *D. interrupta* and *C. pusilla*. The presence of *D. didyma*, *N. peregrina*, *T. navicularis* and *S. turmida* suggested intertidal mudflats may have existed at a lower elevation in the tidal frame (Figures 4.49, 4.50 and 4.51) (Vos and de Wolf 1993). The polyhalobous marine taxa *R. arcuatum* and *T. antediluvian* were found, as well as *A. senarius* and *R. amphiceros* which are both species commonly found in tidal inlets and subtidal channels (Denys (1991/2). Some poorly preserved, potentially allochthonous freshwater species were also found such as *S. phoenicentron* and *Pinnularia* sp. Overall, the diatom evidence implied an environment near mean high-water level with aerophilic species such as *D. interrupta* common, as well as pools containing intertidal species such as *D. didyma* (Vos and de Wolf 1993).

The foraminifera analysis complemented the diatom data, with evidence for a high marsh environment on the very edge of marine influence indicated by a monospecific *J. macrescens* assemblage (Scott and Medioli, 1978; Gehrels, 1999; Wright *et al.,* 2011; Horton and Edwards, 2006). Similarly, the pollen evidence from the basal peat indicated local reed swamp and fen communities represented by grass, sedges, meadowsweet and bedstraw, as well as saltmarsh indicated by goosefoot, daisy, and sea plantain. There was also alder carr represented by alder and willow, as well as mixed oak-birch woodland. Woodland edge herb species were present such as nettles, buttercup, and docks. The pollen, diatom and foraminifera evidence combined indicates that at

Penllyn there was a back barrier saltmarsh on the edge of marine influence, with a diverse vegetation of saltmarsh, reed swamp and fen communities. Nearby, there was alder carr with diverse herb species, and mixed-oak woodland in the surrounding valley.

At 500cm (-3.24m OD) there was a transition to blue-grey silt with orange mottling, and poorer diatom preservation. Orange mottling is indicative of iron pan which implies that the sediment was exposed to air during deposition, and can cause problems for diatom preservation (Tom Hill, pers comm). The diatom evidence suggests that saltmarsh conditions were replaced by intertidal mudflats, with *D. interrupta* and *C. pusilla* replaced by *C. westii*, *T. navicularis* and *C. littoralis*. The evidence is unclear, however, due to low counts and poor preservation. The foraminifera evidence supports an increased marine influence, with *T. inflata* and *M. fusca* appearing alongside *J. macrescens*. Foraminifera indicate high marsh conditions, in contrast to the intertidal diatom assemblage. The combination of *J. macrescens, T. inflata* and *M. fusca* have been found in high marsh environments globally (Gehrels 1999, Horton *et al.,* 1999, Horton and Edwards 2006, Wright *et al.,* 2011, Kemp *et al.,* 2015). The foraminifera analysis confirms marine conditions were increasing but indicates that the transition from high saltmarsh to intertidal mudflats was more gradual than the diatom evidence implied. Similarly, the pollen evidence suggests comparable vegetation communities persisted after the transition from peat to silt, with little change in the pollen profile.

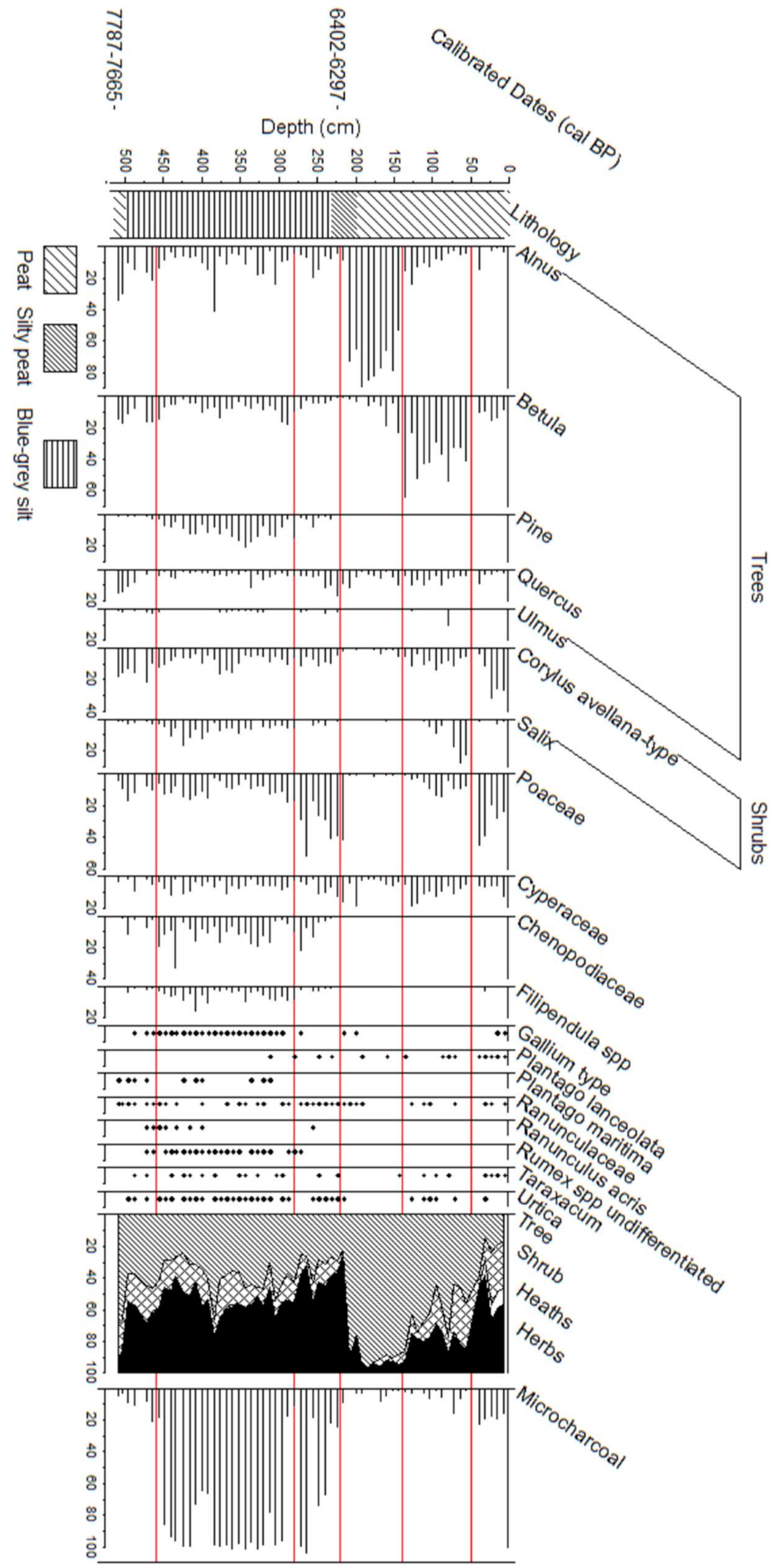


Figure 4.49: Summary pollen percentage diagram for Penllyn showing trees, shrubs, herbs, microcharcoal, stratigraphy and radiocarbon dates.

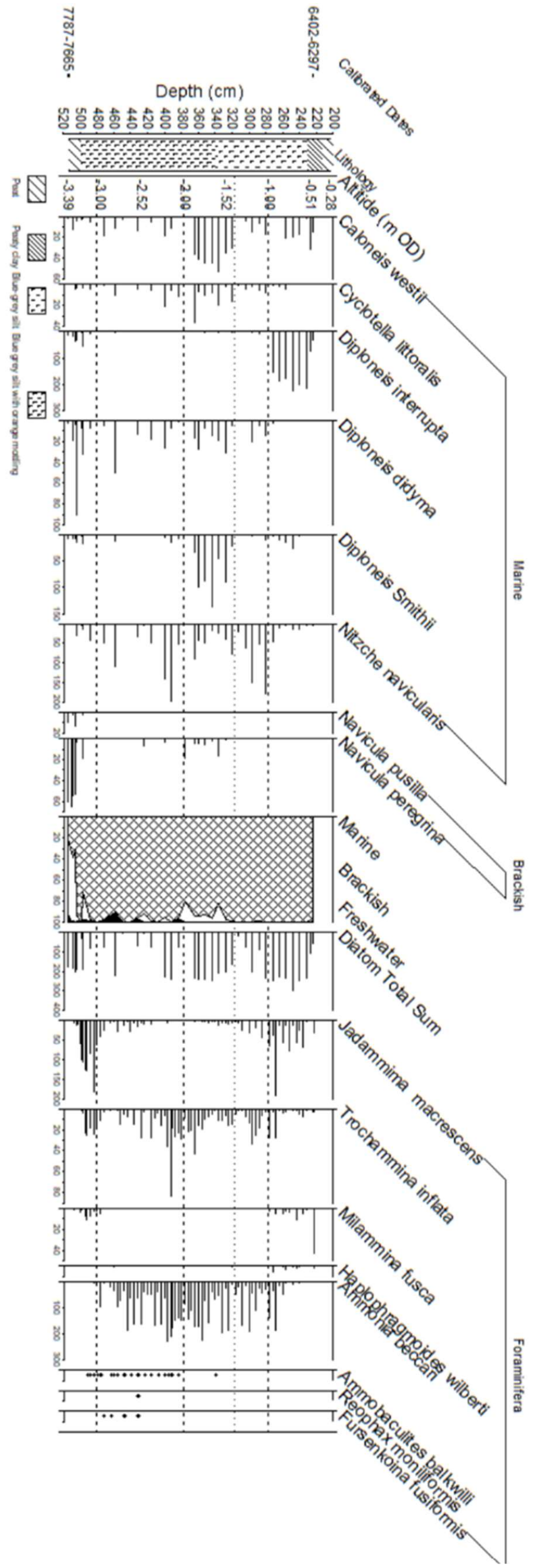


Figure 4.50: Summary foraminifera and diatom diagram for Penllyn with diatom percentages (species less than 20% removed) and foraminifera concentrations (species less than 5% removed).

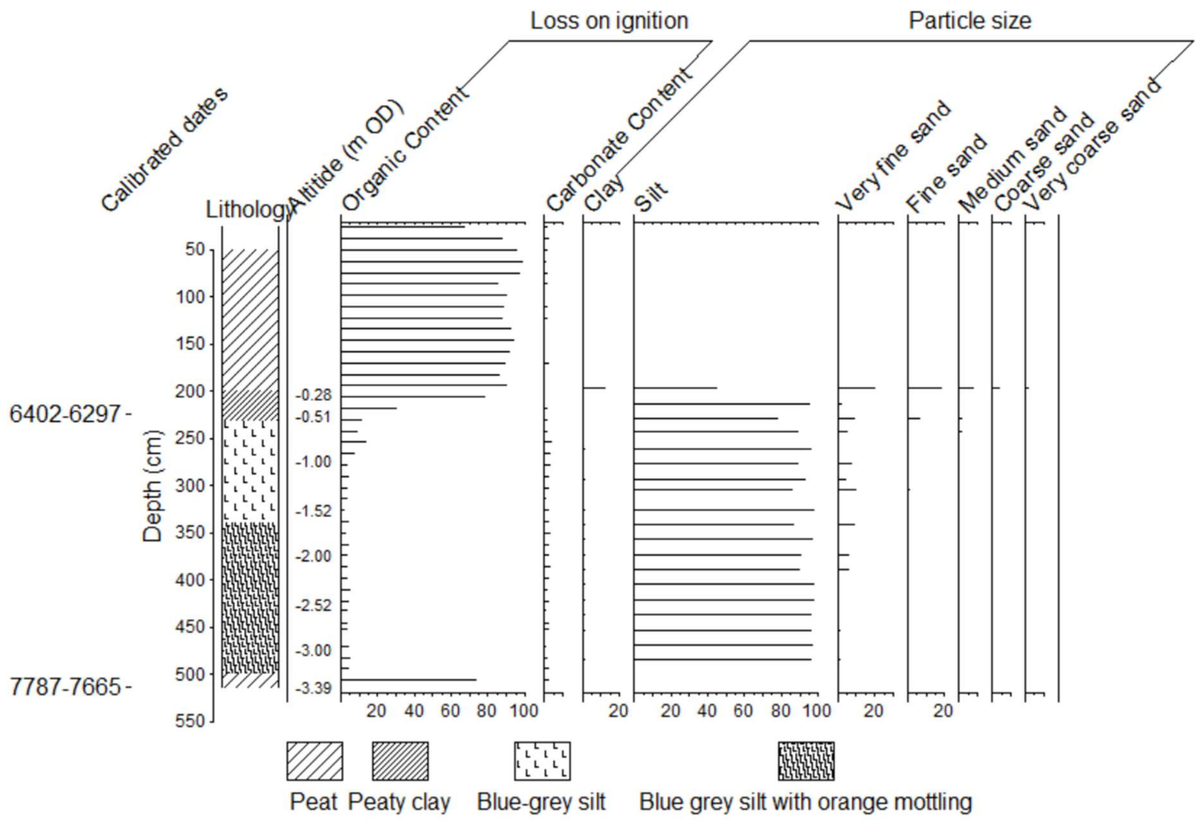


Figure 4.51 Summary diagram for Penllyn with loss on ignition and particle size.

From 480cm (-3.04m OD) diatom preservation continued to be poor, with intertidal mudflat species persisting, such as *C. westii*, *D. didyma* and *C. littoralis*. Marine diatoms such as *A. senarius, C. neofastuosus, S. turmida*, *R. arcuatum* and *A. sculptus* were also present, but no freshwater species were found. The foraminifera evidence also indicated a shift towards a low marsh, intertidal environment as *Ammonia* sp. became dominant, with some inner shelf species washed onto the marsh such as *F. fusiformis* and *R. moniliformis* (Murray, 1971, 2000).

From 460cm (-2.84m OD) pollen preservation and concentration declined. On the transition from peat to silt, there was a change in the pollen catchment area, from a more local marsh to an estuarine environment, with pollen transported from the whole river catchment. This can be seen from the high proportion of poorly preserved pine pollen, which is common in estuarine sediments as pine pollen floats (Hopkins, 1950; Caseldine, 1992, 2000; Scaife, 1993, 1994). Pollen evidence suggested an expansion of alder carr surrounding the estuary, while some mixed-oak woodland persisted in the area. An expansion of reed swamp and tall herb fen can be seen from high levels of grass, with a dramatic increase in meadowsweet, bedstraw and sedges. There was a corresponding expansion of saltmarsh indicated by an increase in goosefoot, as well as herbs such as daisy, carnation, common privet, and sea plantain. Herbs common to open, disturbed habitats such as ribwort plantain, mugworts, docks, buttercups, nettles and dandelions were found, which may be representative of the woodland edge nearby (Bell *et al.,* 2007). The aquatic species watermilfoil was present indicating the presence of standing water nearby. There was also a dramatic increase in microcharcoal, however this could be due the change in the catchment area caused by transition from peat to silt, rather than a change in the use of fire. From 460cm (-2.84m OD), foraminifera and diatom evidence were indicative of a low marsh or intertidal environment, with pollen indicating an expansion of diverse saltmarsh and reed swamp communities in the local area, as well as an expansion of alder carr. There was an increase in transported, allochthonous species due to the change in catchment area, with diatom and foraminifera evidence of allochthonous marine species, as well as an overrepresentation of microcharcoal and pine pollen.

At 380cm (-2.04m OD) foraminifera evidence indicated a continued low marsh, tidal flat environment with dominant *Ammonia* sp. and *T. inflata,* but lower species diversity. The pollen profile remained unchanged with poor preservation. In contrast, diatom preservation increased. Intertidal mudflat species such as *D. didyma*, *C. westii* and *C. littoralis* continued while a higher salinity was represented by a peak in *D. smithii*. Above 340cm (-1.64m OD) there was a reduction in orange mottling and iron pan within the silt, implying a reduction in aerial exposure. Between 325-276cm (-1.49 to -1.00m OD) diatom evidence indicated brackish conditions, with the profile dominated by broken *T. navicularis* and increasing levels of *D. interrupta*. Foraminifera evidence suggested low marsh conditions persisted as *Ammonia* sp. remained dominant, with declining marine conditions, implied from increasing *J. macrescens.*

From 280cm (-1.04m OD), pollen indicated that the valley remained largely occupied by woodland, with an increase in oak while birch, ash and pine declined. Willow and alder continued suggesting that alder carr persisted. Woodland edge and disturbed habitat species such as buttercups, docks and nettles were present as well as low amounts of ribwort plantain, mugworts and dandelion. Grass and sedges peaked and remained high through the zone, indicating that reed swamp persisted. Valerian was present, which is found in many environments including marshes and fen (Grimes, 1988). There was a peak in goosefoot followed by a steady decline throughout the zone, but other saltmarsh species were not present, indicating a decline in saltmarsh communities locally. Meadowsweet fell, disappearing by the end of the zone, implying the loss of tall herb fen. The pollen evidence therefore supports the foraminifera and diatom evidence of reducing marine conditions, with saltmarsh and fen communities disappearing, replaced by reed swamp, while the proportion of woodland remained unchanged.

Between 276-223cm (-1.00 to -0.47m OD) diatom evidence indicated a return to a brackish saltmarsh environment frequently exposed to the air, with calm conditions (Denys 1991/2). *D. interrupta* became dominant, with *T. navicularis*, *C. westii* and *D. smithii* present. The foraminifera evidence supported the diatom analysis, with indicators of high marsh conditions. *J. macrescens* was dominant with *M. fusca*, *T. inflata* and *Haplophragmoides* sp. present. From 260cm (-0.84m OD) there was a shift towards an increase in sand fraction with a corresponding rise in organic content from <5 to 10%. At 232cm (-0.54m OD) there was a transition to peaty silt, with diatom, pollen and foraminifera evidence indicating that brackish, high saltmarsh conditions persisted across the boundary.

At 224cm (-0.48m OD), the peat was dated to 6402-6297 cal. BP (D-AMS 028035) with no diatoms found above this level. From 223 to 221cm (-0.47 to -0.45m OD) there were very low counts of *J. macrescens* and *M. fusca*, indicative of high marsh conditions, with no foraminifera preserved above 221cm (-0.45m OD) to the surface. Organic content rose to 80% with low carbonate content from

214cm (-0.38) to the surface. Particle size analysis indicated an increase in all sand fractions at 220cm (-0.44m OD).

Between 220 to 140cm (-0.47 to 0.36m OD) pollen preservation improved as the silty peat gave way to woody peat, and the pollen catchment became more local in extent. Alder expanded to dominate up to 80% of the profile, reflecting alder carr encroaching onto the marsh. Other tree species declined, while oak continued indicating mixed oak woodland persisted in the region. Evidence of saltmarsh, fen or reed swamp was sparse. Herbs were much less diverse, as goosefoot, meadowsweet, bedstraw, buttercups and docks largely disappeared. Grass dramatically declined while sedges continued. The occurrence of some very low amounts of ribwort plantain implied the presence of some disturbed or open habitats, but care must be taken, as the dominance of alder may obscure the profile.

At 140cm (0.36m OD) alder rapidly declined, replaced by birch, which dominated up to 60% of the pollen profile, implying conditions were drier. Regionally mixed-oak woodland expanded as oak increased, as well as heather and willow. Lime, elm, and ash were present in low amounts. Many herb species, which prefer open, disturbed habitats were present such as mugwort, buttercup, dock, dandelion and nettles as well as ribwort plantain. There was a rapid increase in grass, and sedges remained high, with some meadowsweet indicating an expansion of reed swamp and fen communities. No saltmarsh species were present.

At the top of the pollen profile (49cm to surface, 1.27 to 1.76m OD), there was evidence for the opening up of the landscape, as hazel increased up to 40 % while all trees declined. Heather increased sharply indicating an expansion, likely in the surrounding uplands and surrounding coastal plain. Very high grass and sedges indicated grasslands and possible reed swamp, with some tall herb fen indicated by meadowsweet and bedstraw. Sea heath appeared which is common to marshes or dunes. Open habitat and woodland edge taxa were present such as ribwort plantain, daisy, mugwort and chamomile, as well as low amounts of buttercup, nettle, and honeysuckle. Microcharcoal increased slightly through the zone, with no corresponding change in catchment or preservation, possibly indicating increased burning in the local area. An increase in anthropogenic impact may be interpreted from the combination of disturbance indicators such as ribwort plantain a fall in tree pollen and an increase in microcharcoal.

4.6 Conclusions

The gouge survey at Penllyn revealed a basal peat overlain by a blue-grey silt and finally a woody peat, which may reflect the submerged forest seen at the foreshore (Leng and Pratt, 1987; Smith *et al.,* 2002). A sample core was analysed which produced two radiocarbon dates as well as loss on ignition, particle size, pollen, diatom, and foraminifera analyses. Transfer functions were developed, and two SLIPs were produced, indicating a rise in sea level from -7.82 to -4.91m OD between 7787-7665 cal. BP to 6402-6297 cal. BP.

The environmental reconstruction indicated a back barrier saltmarsh existed from 7787-7665 cal. BP with mixed oak woodland, alder carr, saltmarsh, reed swamp and tall herb fen. The transition to blue-grey silt at -3.24m OD led to a shift to intertidal mudflats and high marsh, with increasing marine conditions. Low marsh developed with an expansion of alder carr, saltmarsh, fen and reed swamp. From -1.49m OD marine conditions began to decline, with low marsh, intertidal conditions transitioning to high marsh with a reduction in saltmarsh, reed swamp and fen pollen. From 6402-6297 cal. BP (-0.48m OD) diatoms and foraminifera were absent, as alder carr expanded onto the infilled marsh. From 0.36m OD birch replaced alder, with mixed-oak woodland and diverse woodland edge taxa. From 1.27m OD to the surface the landscape opened up with some tall herb fen and evidence of widespread human impact.

The sequence at Penllyn represents the period from 7787-7665 cal. BP to the present, with multiproxy analyses revealing the coastal and vegetation changes. Two sea-level index points reveal the vertical sea-level change. However, due to the estuarine nature of the sediment, it is possible that the pollen catchment may represent a much wider area with transported pollen grains.

Furthermore, although the vertical sea-level change is established, the horizontal sea-level change is unclear.

Chapter 5: Perfeddnant

5.1 Introduction

Northeast of Bryncrug, the broad floodplain of the Dysynni Valley narrows to an over-deepened, glacial valley that extends to Craig yr Aderyn (Birds' Rock) (Figure 5.52) (Snowdonia National Park Authority, 2014b). The post-medieval gentry estate of Peniarth overlooks the valley from a raised area of till (20m AOD) (Snowdonia National Park Authority, 2014b). Blundell *et al.* (1969) undertook a seismic survey across the Dysynni Valley near Peniarth and found the rock floor of the valley was between 91 and 167m deep beneath the glacial till (Figure 5.53). The steep valley side to the south is punctuated by narrow valleys which produce small alluvial fans at intervals. The northern valley side slopes more gradually to the valley bottom (Snowdonia National Park Authority, 2014b). The river Dysynni was canalised upriver from Peniarth, and the LiDAR data reveals a number of paleochannels in the valley bottom. There is a single, clear river channel at Perfeddnant which gives way to multiple smaller paleochannels further upriver at Gesail (Snowdonia National Park Authority, 2014b) (Figure 5.54).

Figure 5.52: Photograph of Perfeddnant looking northeast towards Craig yr Aderyn

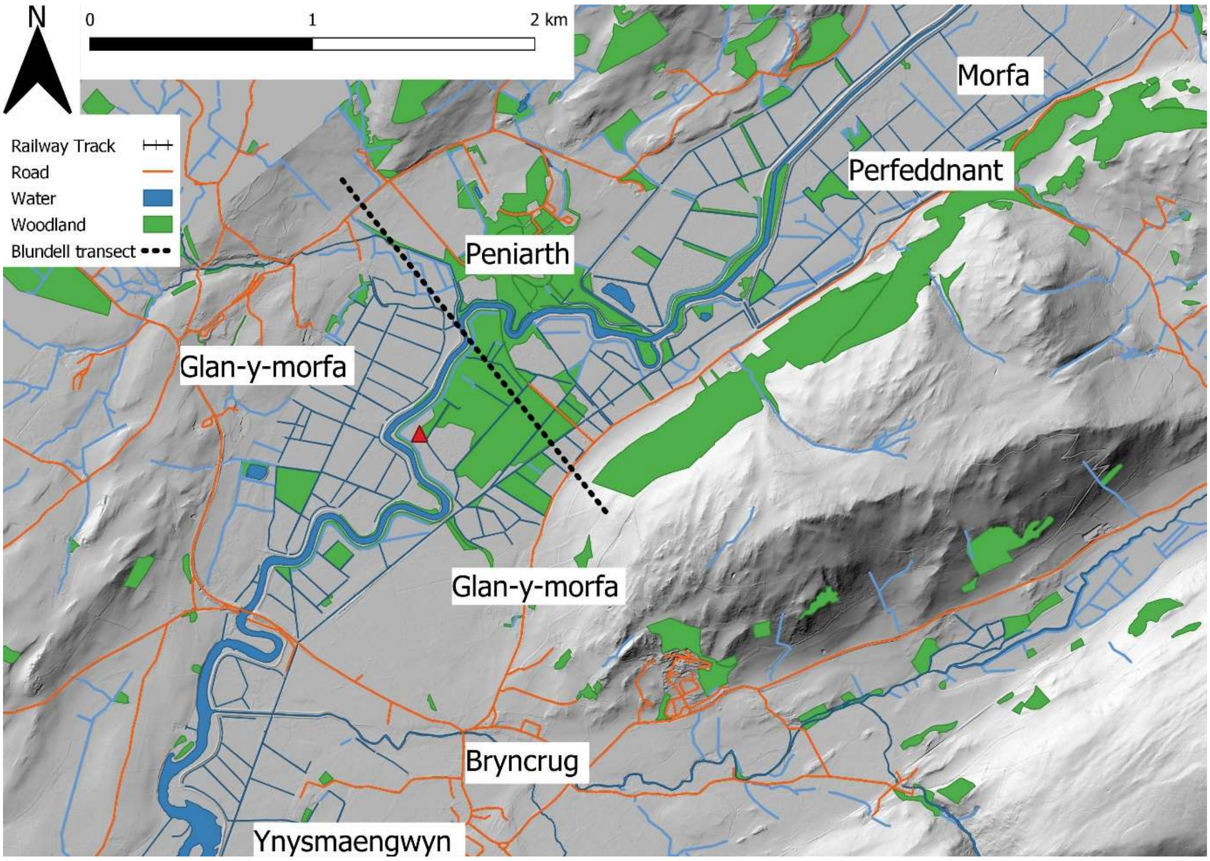


Figure 5.53: Location map showing Perfeddnant, and locations mentioned in the text and transect by Blundell *et al.,* (1969). Triangle indicates location of 'allotment in Gwyddelfynydd marsh'. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved

The valley bottom at Perfeddnant is primarily improved grassland with thistles and nettles as well as spiky grass in the damper areas, such as palaeochannels (Figure 5.54). Some fields have *sphagnum* moss, particularly to the north. The ditches contain nettles, grass, ferns, small bushes and trees such as oak and willow, brambles, reeds and rushes. At the edge of the ditches, where the ditch fill has been left, there are nettles and docks. There is a wooded lane from the road towards the river with a hedgerow including yew, damson and plum, holly, brambles, dock and ivy. The hedgerow gives way to a wall with ivy, small bushes, oak and willow saplings which eventually merge into a patch of woodland. The two small areas of woodland are mixed oak-birch.

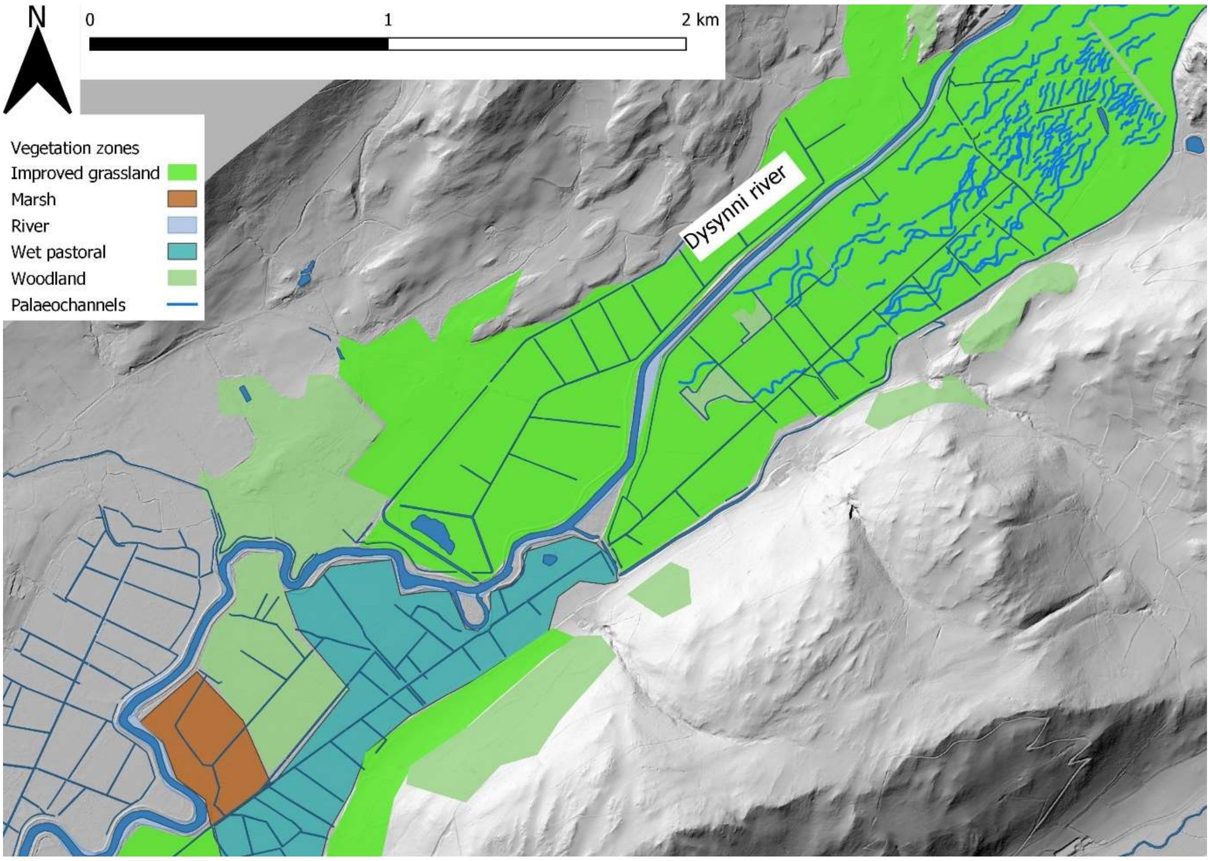


Figure 5.54: Map showing results of the vegetation survey and palaeochannels. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved

A number of British Geological Survey (BGS) boreholes have been undertaken in the area (Figure 5.55). As the primary aim was to establish groundwater depth the overlying sediments were not recorded in detail. At Tal-y-bont farm, 55m of light grey mudstone with fractures was recovered, overlain by 0.3m of soil. At Peniarth, boreholes BH1 and BH2 revealed soft light grey mudstone and hard, fractured, purple-grey-orange rock up to 121m deep overlain by 3m of soil and stone. At Glan Machlas, Peniarth Ganol and Peniarth Uchaf Llanegrin, gravel and rock were recorded overlain by 1.5 m of soil. All of these records were low resolution with superficial descriptions. At Dysynni Bridge, eight boreholes revealed basal gravel up to 10m deep, overlain by firm grey silt, gravel and peat. Pratt *et al.,* (1995) refer to an estuarine alluvium which was deposited at Craig yr Aderyn, overlain by river alluvium. The alluvium was described as a pale grey clay with rootlets, representing vegetated mud banks and blown sand deposited within tidal channels. No stratigraphic records were published, however, and no BGS borehole records have been published near to Craig yr Aderyn.

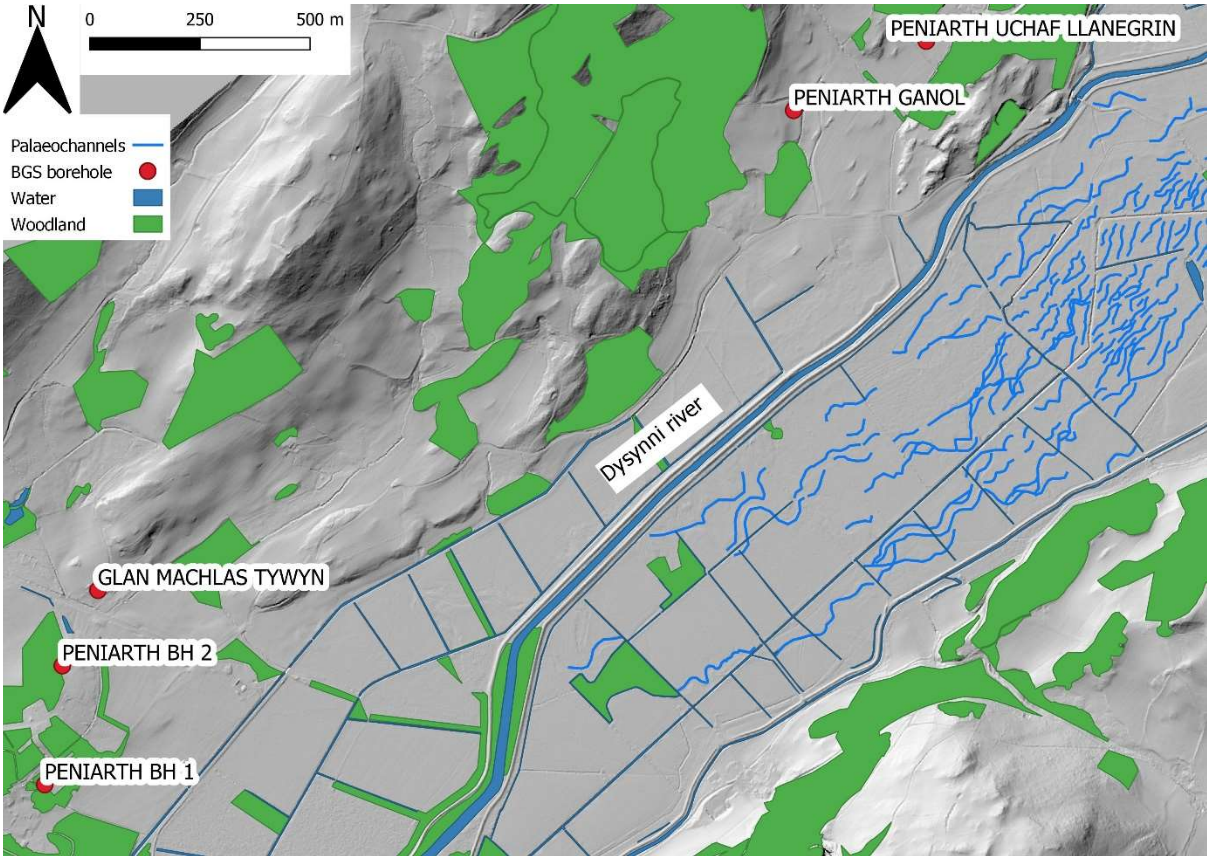


Figure 5.55: Map showing British Geological Survey borehole locations and palaeochannels. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved

5.1.2 Archaeological Evidence

There are extensive crop marks near Bryncrug including circular enclosures indicating roundhouses and field boundaries that extend over 2.3 hectares, although the site remains undated and may represent later prehistory or medieval settlement (Frost, 2012). There are also early medieval square barrow cemeteries near Bryncrug (NPRN 423308) and Croes Faen (NPRN 310263). Cook (2017) undertook geophysical survey on the cropmarks near Bryncrug and Croes Faen and transcribed the results (Figures 2.56, 2.57, 2.58). The survey at Croes Fan produced a number of square structures, previously thought to be a square barrow cemetery (Knight, 2011). At Bryncrug circular, rectangular, and curvilinear features were identified (Cook, 2017). It is important to note that settlement at Bryncrug is located on an alluvial fan, which would have been similar dry hills in a potentially marshy estuary.

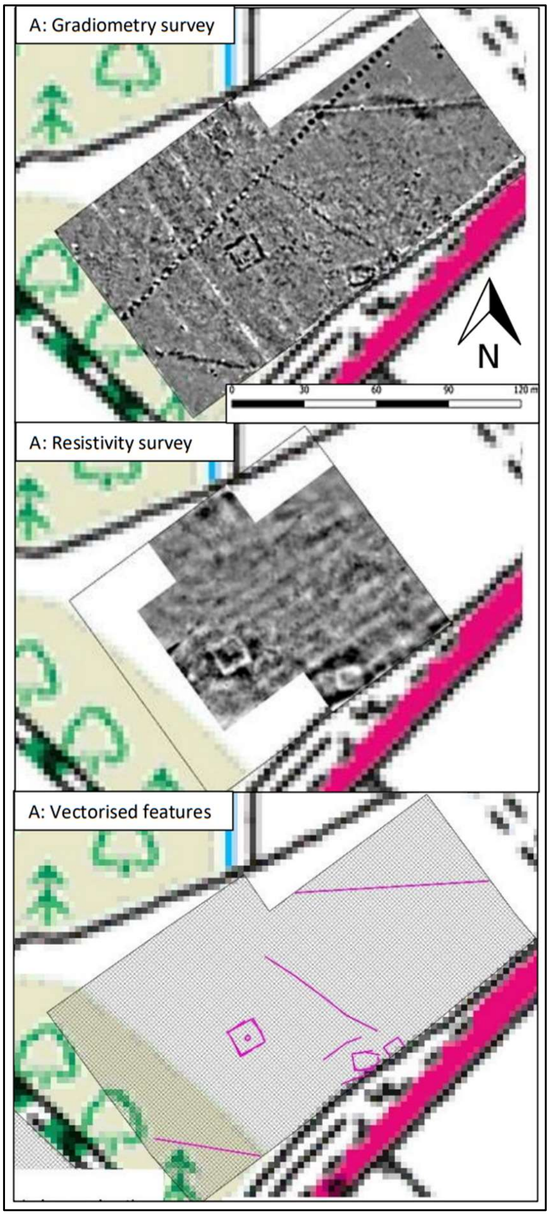


Figure 2.56: From Cook (2017) Geophysical survey plots near Croes Faen (area A, field 2) showing the features identified. Both resistivity and gradiometry were undertaken and were used to render the subterranean features as a vector file in QGIS. Crown copyright and database right 2019 Ordnance Survey 100025252

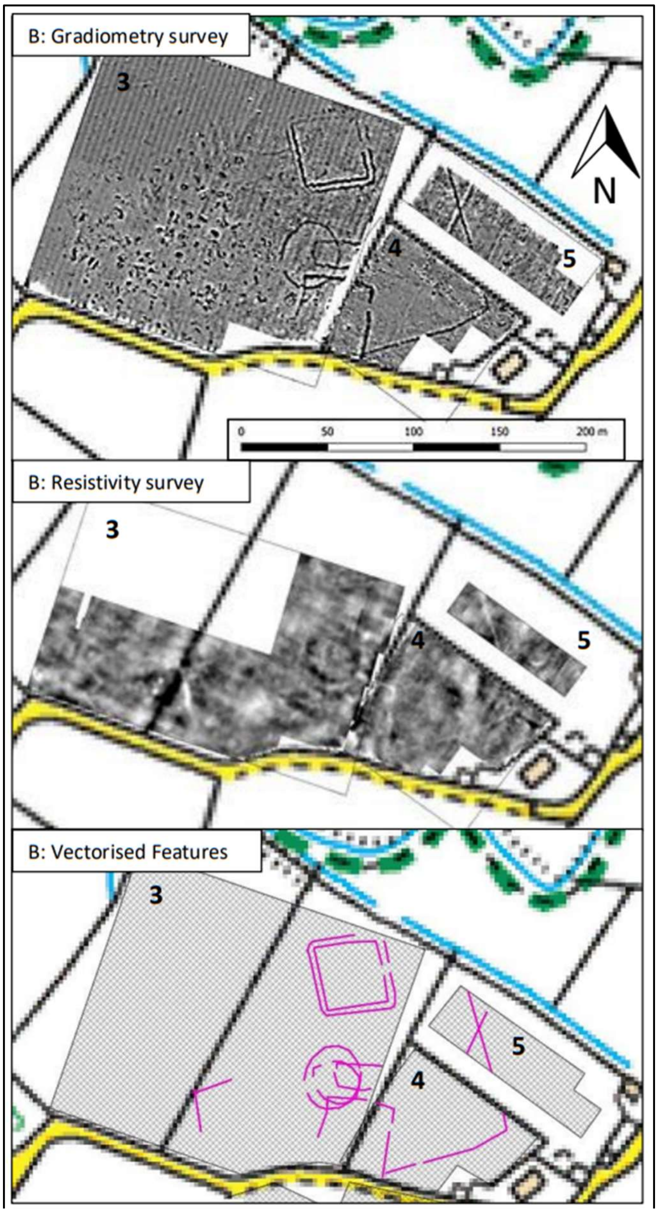


Figure 2.57: From Cook (2017): Geophysical survey plots near Bryncrug (area B, fields 3, 4 and 5) showing the features identified. Both resistivity and gradiometry were undertaken and were used to render the subterranean features as a vector file in QGIS. Crown copyright and database right 2019 Ordnance Survey 100025252

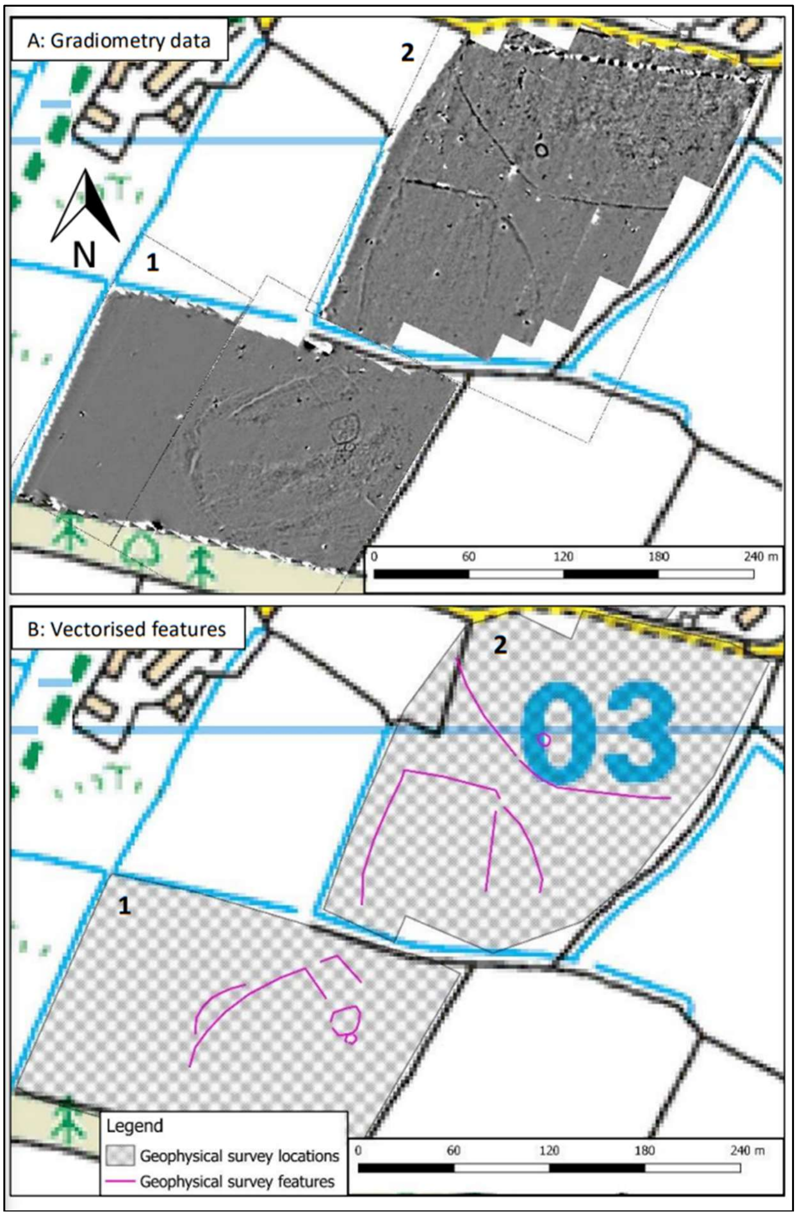


Figure 2.58: From Cook (2017) Processed gradiometry data (A) and vectorised geophysical survey results (B) from near Bryncrug (area B fields 1 and 2) showing the features identified. Crown copyright and database right 2019 Ordnance Survey 100025252

A number of castle mounds in the Dysynni valley have been attributed to the medieval period, including Talybont Castle, also known as Domen Ddreiniog (NPRN 302714) located at a former river crossing, which may have been associated with a local *llys* or princely court, with an associated *maerdref*, or township, still known locally as Tal-y-bont (Frost, 2012). Castell Cynfall (NPRN 302770) was an isolated motte with a short-lived castle that was destroyed in 1147, located on a ridge to the south of the Dysynni (Smith *et al.*, 2001).

The post-medieval gentry estates were both located on moraine hills, for example the estate at Ynysmaengwyn (NPRN 54223) (see Figure 5.59) which was rebuilt in 1757 for the Corbett family and was later demolished in 1964 (Frost, 2012). It has been argued that Ynysymaengwyn may have been within the range of the tidal estuary prior to land reclamation, which would have greatly enhanced potential for trade (Gwynedd Archaeological Trust, 2017c). Peniarth (NPRN 28633) was built as early as the fifteenth century and was expanded around 1700 by Richard Owen (Frost, 2012). Edward Corbett, owner of the Ynysmaengwyn estate in the late eighteenth century, undertook extensive agricultural improvements throughout the valley (Frost, 2012). Prior to improvement the tithe maps list much of the land use as 'turbay' or peat cutting, while after land reclamation and drainage much of the Dysynni valley was under hay. During the seventeenth and eighteenth centuries there was an increase in population of the valley, taking advantage to improved communication in the form of roads (Frost, 2012). There is also a rich industrial history of the Dysynni valley such as quarries and mines. The Talyllyn Railway is a narrow-gauge railway built in 1865 for the slate quarries at Bryneglwys from Abergynolwyn to the standard-gauge railway at Tywyn (Gwynedd Archaeological Trust 2017d).

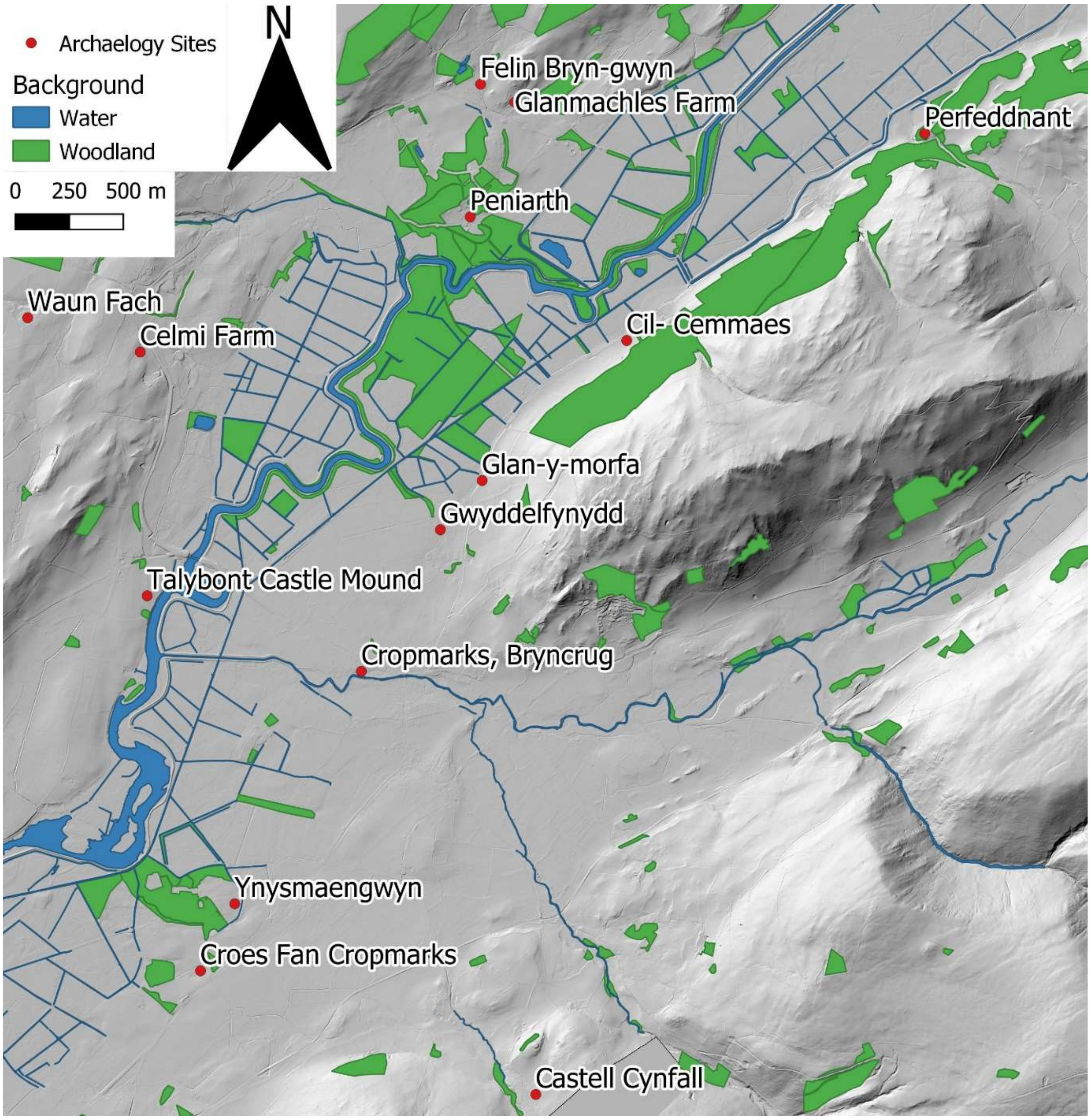


Figure 5.59: Archaeological evidence near Perfeddnant, indicating sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales and Archwilio HER information © Natural Resources Wales and Database

The 1841 tithe map near Perfeddnant reflects a similar field system to modern day, with a very large, regular enclosure, which is more modern than the small, piecemeal enclosure of the uplands. Many field names indicate the presence of marsh, such as 'morfa' which translates to marsh, as well as 'allotment in Gwyddelfynydd marsh' (Figure 5.53). There are three places named 'Glan-y-morfa' including a post medieval farmstead (NPRN 28434) which can be translated to 'next to the coastal marsh' indicating the presence of saltmarsh or tidal estuarine environments nearby, despite the local river having no tidal influence. The medieval and post medieval buildings are located on dry land, either in the valley uplands or along the valley sides. Postmedieval farmsteads include Pont-y-garth (NPRN 24202), Perfeddnant (NPRN 28663), Cil- Cemmaes (NPRN 302238), Glanmachles Farm (NPRN 412907), Felin Bryn-gwyn (NPRN 412906), and Gwyddelfynydd (NPRN 41683).

5.2 Fieldwork

5.2.1 Stratigraphy

A gouge core survey was undertaken from the alluvial fan at Bryncrug towards the northeast, in order to trace the extent and depth of the estuarine alluvium found at Penllyn (Figure 5.60 and 5.61). The west to east transect along the valley (Figures 5.62 and 5.63) revealed a basal gravel, overlain by an intermittent, thin layer of basal peat at locations 6, 7, 9, 10, 12, 13 and 33. The base of the sequence was found between -5.5 and -1.9m OD, but despite multiple attempts, the base of the sequence was not reached at location 1, with a maximum depth of -5.8m OD reached. The basal peat was overlain by blue-grey silt. The upper contact of the blue-grey silt was found between -0.1 and -1.1 m OD. At locations 1 to 10, 12 and 14 the blue grey silt was overlain by woody peat. From location 15 towards the northeast, the overlying peat was replaced by a grey, compact clay with iron staining and some organic banding. The northwest to southeast transect between locations 28 to 40 represents the sequence of blue-grey silt overlain by woody peat (Figure 5.63), while the transect locations 17 to 27 represents the blue-grey silt overlain by compact, grey clay and organic layers (Figure 5.64).

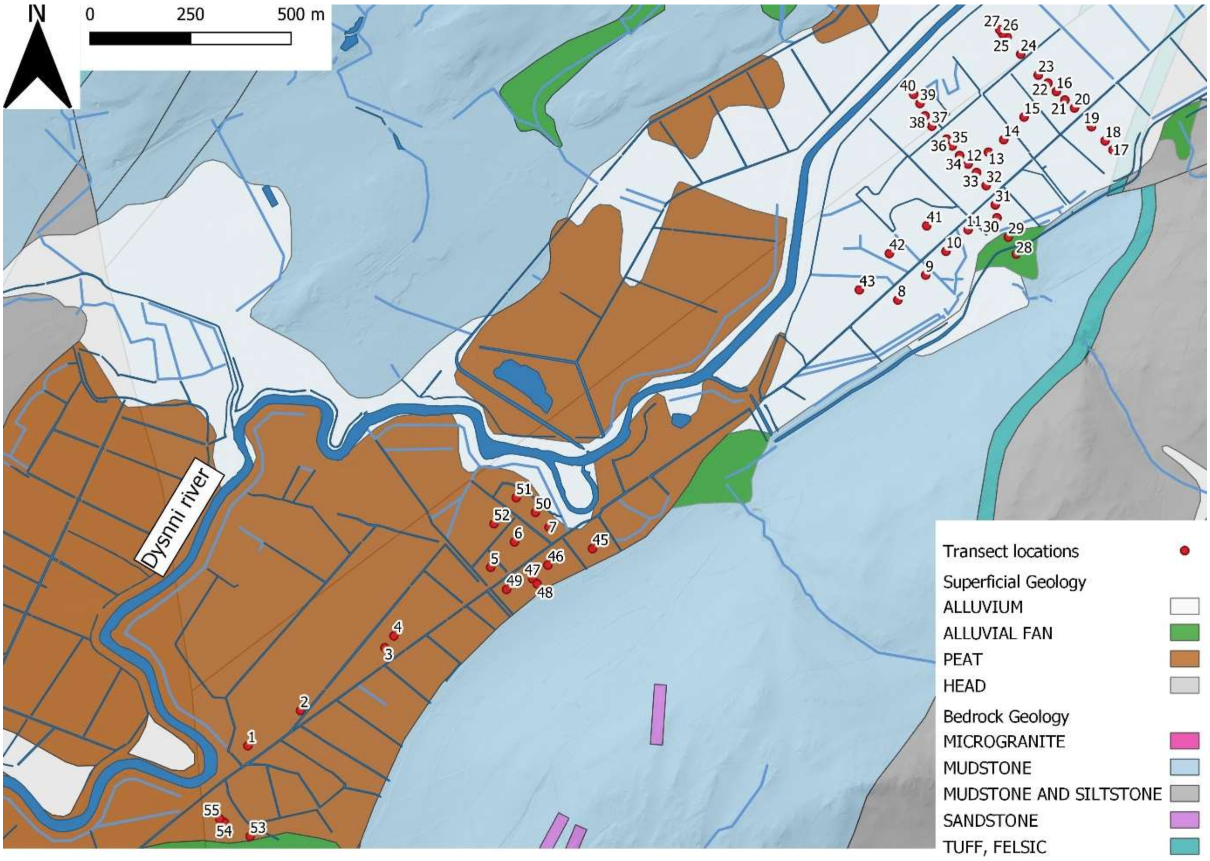


Figure 5.60: Map showing bedrock and superficial geology with location of gouge survey. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved. Geological Map Data BGS © UKRI 2022

Location 33 was chosen for further analysis as it represented the full stratigraphic sequence, with upper and lower peat suitable for dating (Figure 5.63). A core was obtained with a 10cm wide percussion corer, which recovered 5m of sediment, including basal gravel in a clay matrix, basal peat, blue grey silt, and overlying woody peat. The core was placed in guttering and wrapped in cling film and foil and taken to the cold store (<5°C).

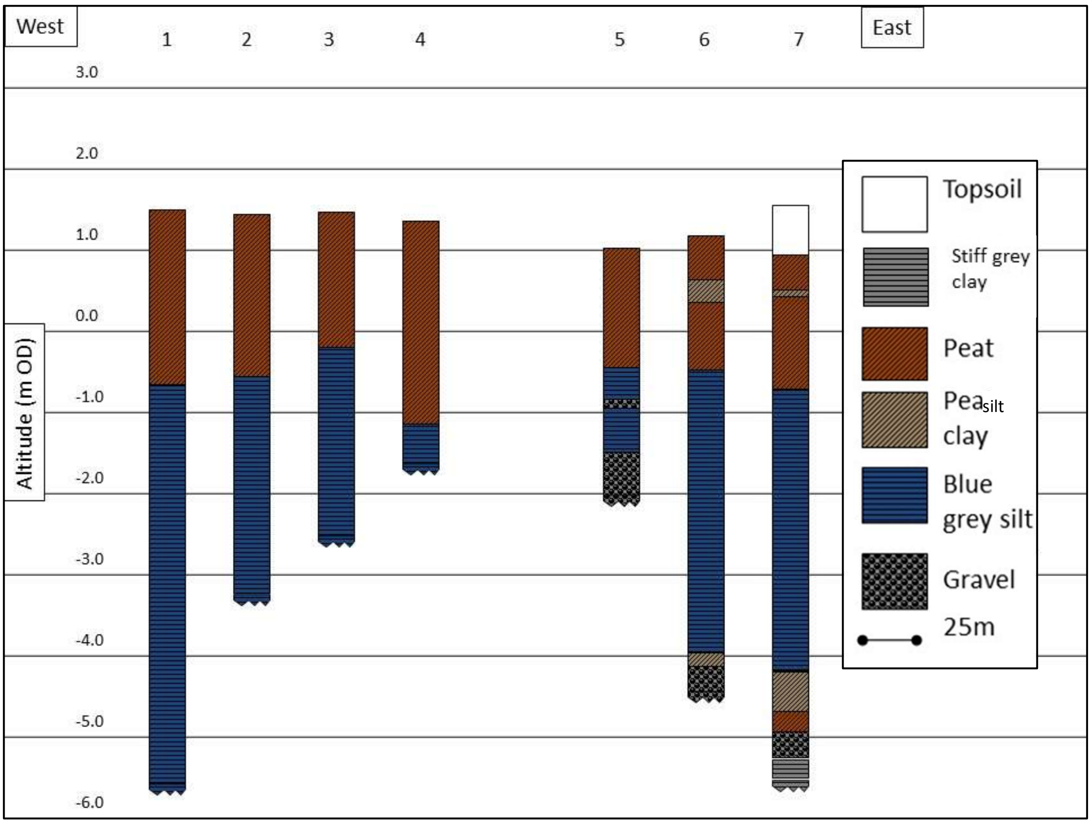


Figure 5.61: Stratigraphic sequence of the transect located west to east, boreholes 1 to 7

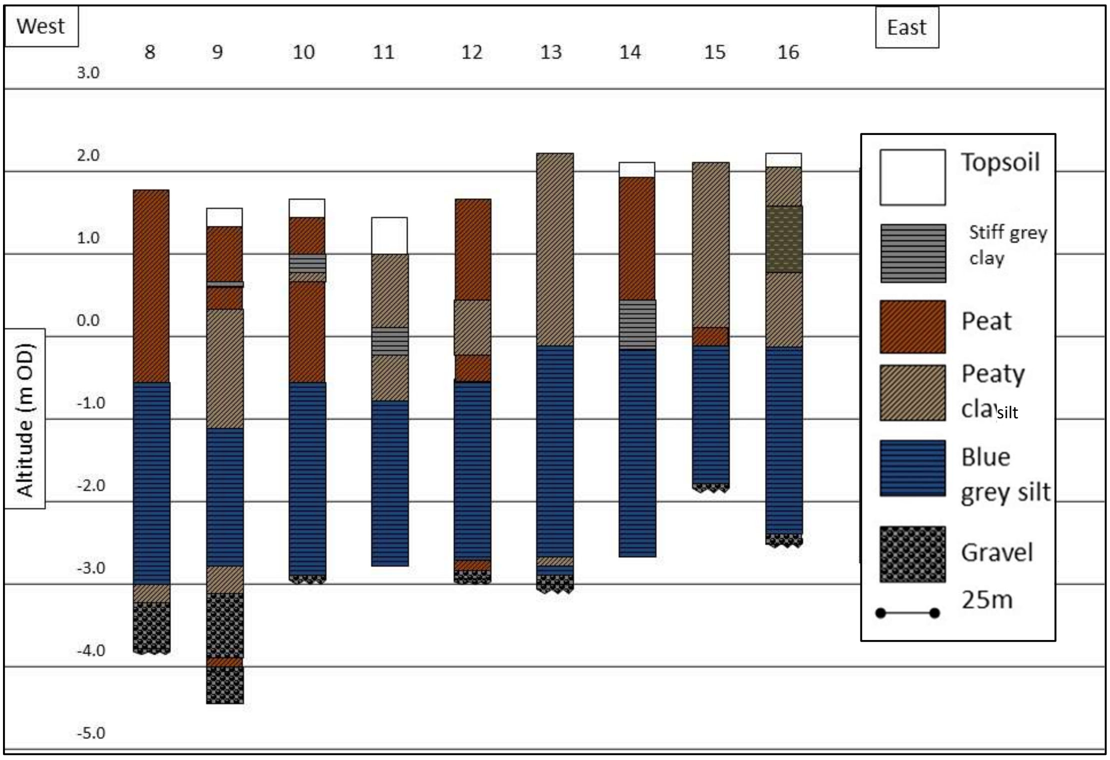


Figure 5.62: Stratigraphic sequence of the transect located west to east, boreholes 8 to 16

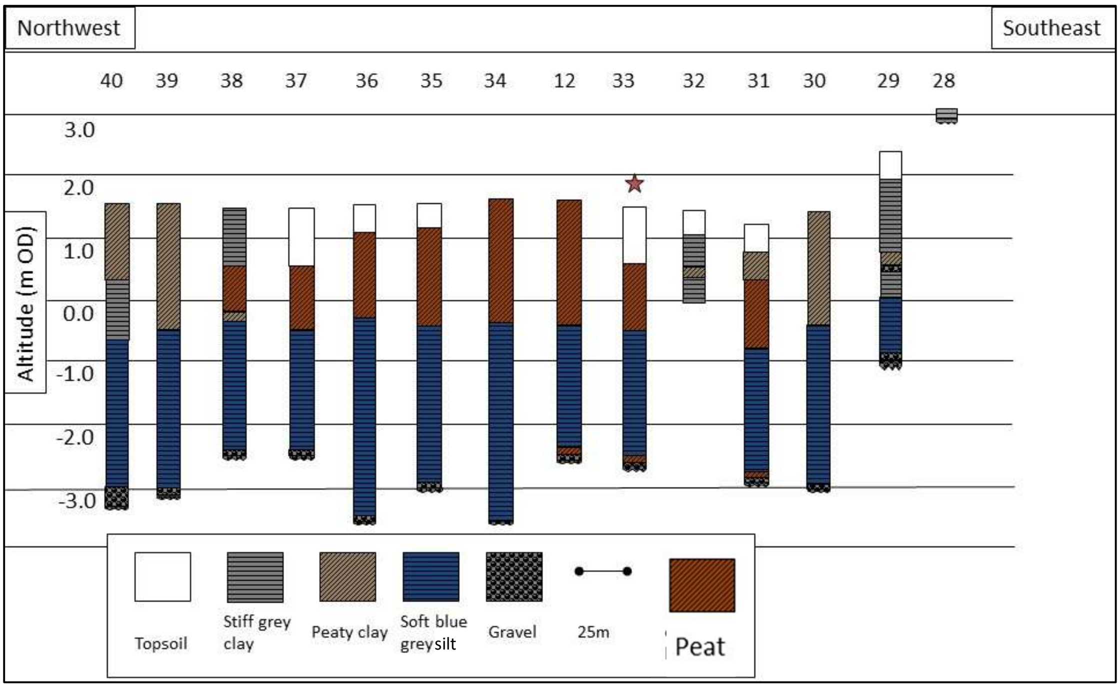


Figure 5.63: Stratigraphic sequence of the transect located northwest to southeast, boreholes 40-34, 12, 34-28. Star indicates sample location

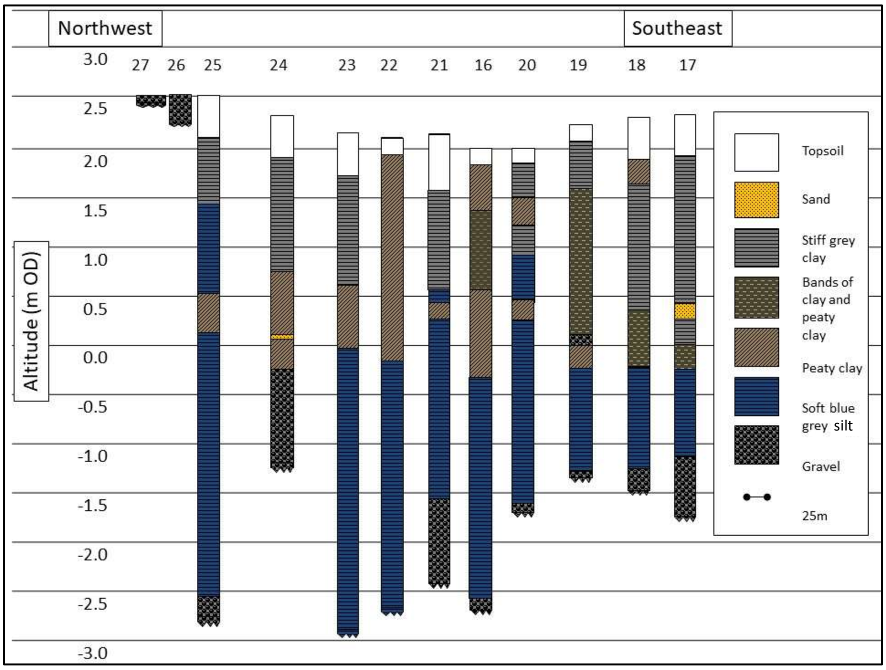


Figure 5.64: Stratigraphic sequence of the transect located northwest to southeast, boreholes, locations 27-21, 16, 20-19

5.3 Laboratory work

In the laboratory, the stratigraphy core 33 was described using Tröels-Smith system (1955). The core was subsampled for loss on ignition, particle size, diatom, foraminifera analyses and radiocarbon dating.

5.3.1 Stratigraphy

A basal gravel was recovered between 500 and 436cm (-3.25 to -2.61m OD), which consisted of poorly sorted angular shale up to 20mm long, in a blue-black clay matrix, with some infrequent quartz inclusions. The basal gravel was overlain by a thin layer of well humified black peat between 436 and 435cm (-2.61 to -2.60m OD), overlain by soft blue-grey clay with iron speckling and reeds to 214cm (-0.39m OD). This in turn was overlain by layers of peaty clay, woody peat and silt up to the surface with some variations in colour, texture and inclusions.

Table 5.44: Description of Perfeddnant stratigraphy following Tröels-Smith (1955). Nig=Nigro, Strf-Stratificatio, Elas = Elasticitas, Sicc = Siccitas, Strf = Structura, lim = Limes, Hum = Humositas. See methods section.

Depth (cm)	Altitude (m OD)	Description	Nig	Strf	Elas	Sicc	Colour	Stru	Lim	Hum	Composition
0-6	1.75 to 1.69	Clay topsoil	1	0	2	1	Grey	Homogeneous	0	0	2Th (*Phragmites)* 1Tb +1Ag
6-25	1.69 to 1.50	Red brown peat with roots	2	0	2.5	2	Red Brown	Fibrous	0	1	3Tb plus 1Th (*Phragmites*)
25-50	1.50 to 1.25	Brown silty peat with roots	2	0	2	1	Brown	Fibrous	0	1	2Th plus 2Ag
50-93	1.25 to 0.82	Grey silty peat with wood	1	0	1	1	Grey	Homogeneous	0	1	1As 1Ag 2Tb plus Wd
93-139	0.82 to 0.36	Red-brown woody peat	1	0	2.5	2	Red-brown	Fibrous	0	1	Ag1 Tb3 + *Phragmites*, wood, roots
139-185	0.36 to -0.10	Black silty peat	2.5	0	1	1	Black	Homogeneous	0	2	Ag2 Tb 2 + *Phragmites*, wood
185-214	-0.10 to -0.39	Brown peaty clay with wood	1	0	2.5	2	Brown	Homogeneous	0	2	Ag2 Tb 2 + As, wood
214-435	-0.39 to -2.6	Blue-grey clay with reeds	1	0	3	3	Blue-grey	Homogeneous	0	0	As4 + *Phragmites*
435-436	-2.60 to -2.61	Compact, black peat	3	0	1	2.5	Black	Homogeneous	1	3	Ag2 Tb2 +As
436-500	-2.61 to -3.25	Angular, poorly sorted shale in a clay matrix. Inclusions up to 40mm long, some quartz	3	0	0	0	Blue-grey	Granular	-	0	Gmaj1 Gmin2 As1

5.3.2 Loss on Ignition

Subsamples were taken for loss on ignition at 4cm intervals to establish changes in organic and carbonate content (Figure 5.65). Some horizons could not be sampled for loss on ignition due to insufficient sediment. The method used followed Heiri *et al.,* (2001, Chapter 3.3.2). The basal peat was too thin for a sufficient sample to be obtained for loss on ignition, but the clay directly above it had a low organic and carbonate content below 10%. At 214cm (0.39m OD) the organic content increased to 30-40% in the peaty clay, indicating the change from estuarine sediemnts to terrestrial peat. The organic content dropped to 20% before rising again to 50-60% at 169cm (0.06m OD) in the woody peat layer. The organic content fell to 10-20% at 80cm (0.95m OD) in the silt layer before rising to 40-80% in the surface topsoil. The varied organic content in the overlying oody peat indicates variations in the peat not visible from the naked eye. The carbonate content remained low throughout the core, with a small peak near the surface.

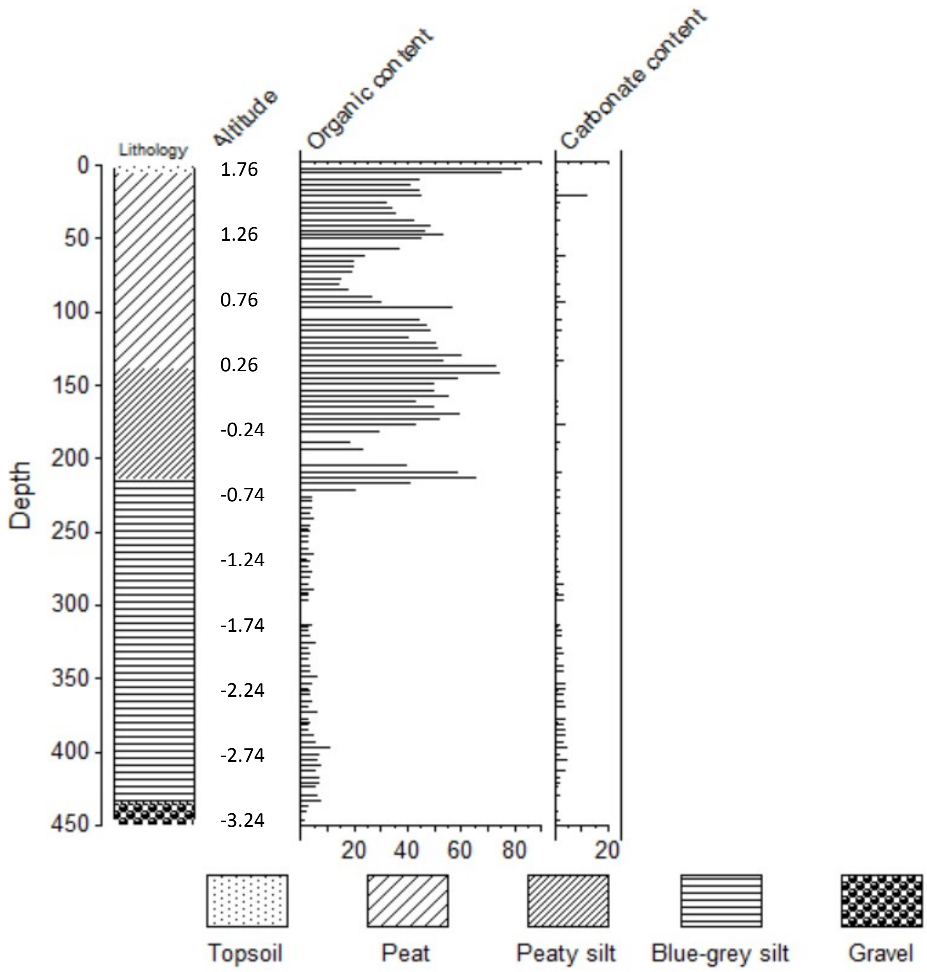


Figure 5.65: Graph showing organic carbon and carbonate content for the sample core at Perfeddnant.

5.3.3 Particle Size

Subsamples for particle size analysis were taken at 12cm intervals (Figure 5.66). Samples could not be obtained between 297-321cm (-1.22 to -1.46m OD) due to lack of sediment available. Samples were pre-treated with hydrogen peroxide to remove organic content and a Malvern laser granulometer was used to determine particle size (Chapter 3.3.3). The basal gravel consisted primarily of the very coarse and coarse sand fractions. The basal gravel contained 0.2g. of material larger than 2mm and consisted of a single, angular shale piece 3mm long. This basal gravel reflects the glacial gravel recovered from Gouge cores across the Dysynni Valley. The blue grey silt was dominated by the silt fraction, with some clay and very low percentages of sand in some samples, indicating a largely low energy environment with some fluvial input. The peat contained very little sediment due to high organic content, but the small sample of the inorganic fraction obtained was dominated by silt with some sand particles, including very coarse sand. There were sharp increases in the sand fraction at 130 cm (0.45m OD) and 214cm (-0.39mOD) which correlate with a small increase in organic or carbonate content, likely indicating an inwash of sediment. There was also a peak in sand fraction at 230cm (-0.55m OD) combined with a higher organic content, which could also indicate an in-wash of sediment.

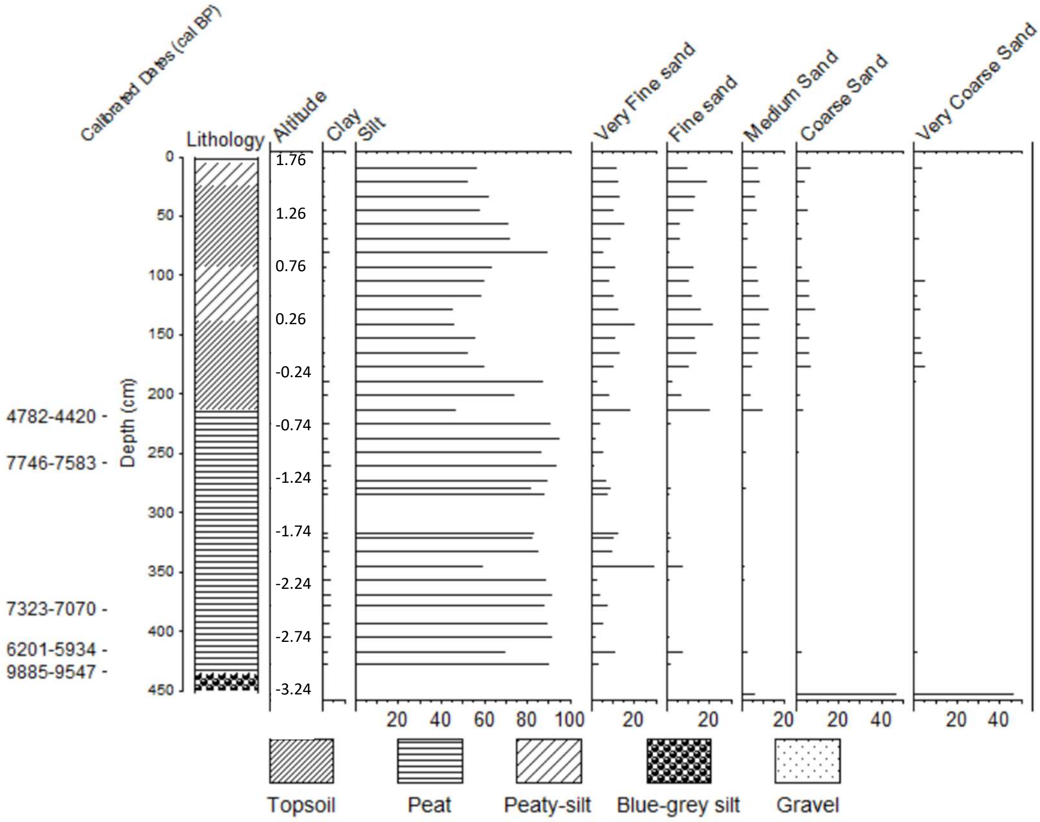


Figure 5.66: Graph showing particle size data for the sample core at Perfeddnant.

5.3.4 Radiocarbon dates

Two bulk samples of peat for radiocarbon determination were submitted from the upper and lower boundaries of the blue-grey silt. No *in situ* macrofossils were available, with only large wood fragments and degraded reeds visible. Three samples of organic silt were dispatched for bulk dating within the silt, however, there appeared to be a reversal as the dates were out of sequence. Two further samples were submitted to date the humin and humate fractions in order to identify the correct sequence of dates. An offset in the humin and humate dates would indicate disturbance has occurred. The humin dates were considerably older than the humate fraction implying an inwash of old carbon. Only the upper and lower dates from the peat will be used for analysis (D-AMS 031519 and D-AMS 031520), as the dates from the organic silt appear to be disturbed.

Table 5.45: Radiocarbon dates for Perfeddnant. Results presented in units of percentage modern carbon (pMC), the uncalibrated radiocarbon age before present (BP) and the calibrated age before present (cal. BP). All results corrected for isotopic fractionation with an δ^{13}C value measured on the prepared carbon by the accelerator. The pMC requires no further correction for fractionation.

DirectAMS code	Sample depth (cm)	Sample depth (m OD)	Sample type	δ (13C)	Fraction of modern carbon		Radiocarbon age		Calibrated age (cal. BP) 95.4% probability
					pMC	1σ error	BP	1 σ error	
D-AMS 031519	219-220	-0.44 to -0.45	Bulk Peat	-37.4	60.52	0.26	4034	35	4567-4436
D-AMS 037471	257-8	-0.82 to -0.83	sediment (humin)	-27.9	18.76	0.12	13443	51	16366-15971
D-AMS 037472	257-8	-0.82 to -0.83	sediment (humic acid)	-21.3	39.77	0.19	7407	38	8338-8170
D-AMS 036266	258-9	-0.83 to -0.84	Bulk Organic silt	-26.7	34.11	0.19	8640	45	9695-9532
D-AMS 036267	382-3	-2.07 to -2.08	Bulk Organic silt	-23.9	36.10	0.20	8185	45	9272-9019
D-AMS 037473	415-6	-2.40 to -2.41	sediment (humin)	-46.9	30.36	0.18	9576	48	11127-10734
D-AMS 037474	415-6	-2.40 to -2.41	sediment (humic acid)	-27.5	39.28	0.21	7506	43	8397-8201
D-AMS 036268	416-7	-2.41 to -2.42	Bulk Organic silt	-30.1	40.94	0.20	7174	39	8150-7883
D-AMS 031520	434-435	-2.60 to -2.61	Bulk Peat	-26.8	33.81	0.17	8711	40	9763-9551

5.3.5 Foraminifera analysis

Subsamples were taken at 4cm intervals throughout the core, where foraminifera were preserved, with contiguous samples taken across biostratigraphic boundaries (Figure 5.67, Table 5.46). No foraminifera were found between -2.75 to -2.43 and -0.83 to 1.75m OD. Concentrations were calculated based on the total volume counted, and seven subzones were established using CONISS.

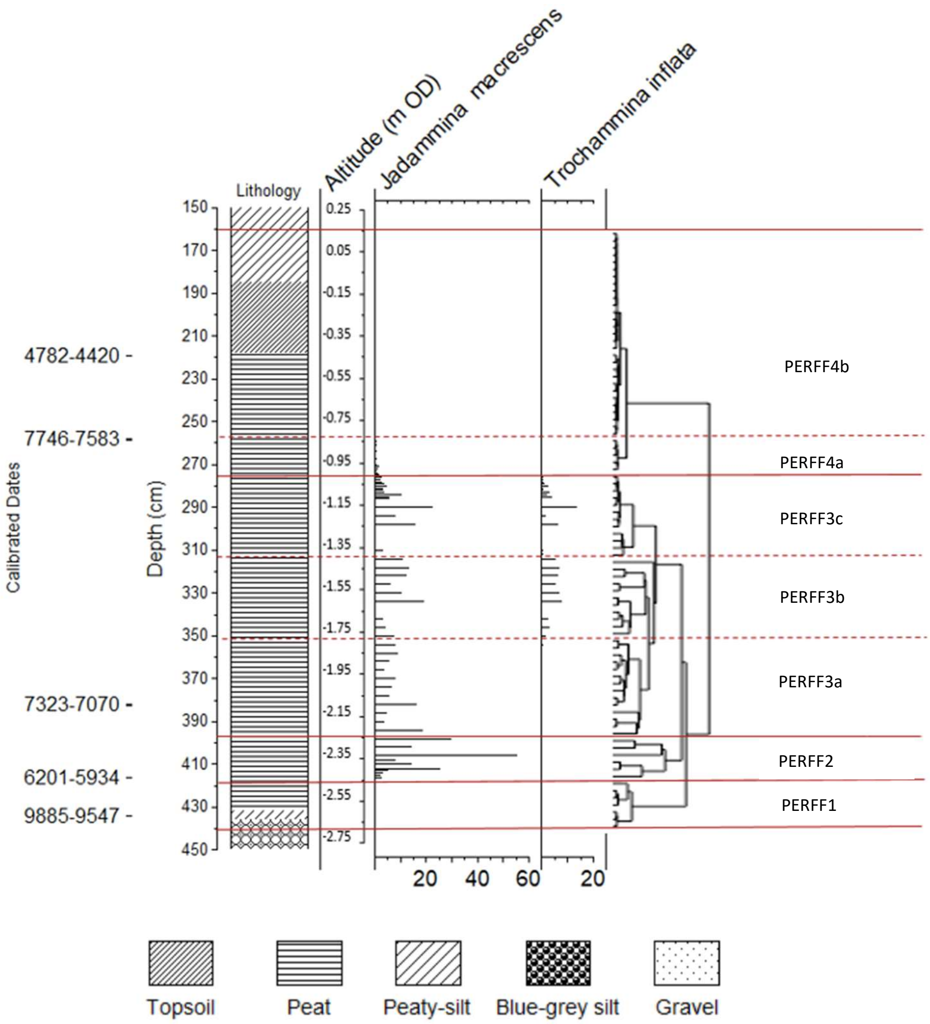


Figure 5.67: Concentration foraminifera frequency diagram for Perfeddnant with radiocarbon dates, stratigraphy and CONISS.

Table 5.46: Summary table of foraminifera zones

Zone	Depth (cm)	Altitude (m OD)	Description	Environmental conditions
PERFF1	450-418	-2.75 to -2.43	Barren of foraminifera	No marine influence
PERFF2	418-398	-2.43 to -2.14	*J. macrescens* appears at 418cm. Monospecific *J. macrescens* assemblage, rising to a peak of 56 per cm^3.	High marsh at the edge of marine influence
PERFF3a	398-351	-2.14 to -1.76	Monospecific *J. macrescens* assemblage continues, concentrations decline to between 19 to 8 per cm^3	High marsh at the edge of marine influence
PERFF3b	354-312	-1.76 to -1.37	*T. inflata* appears at 354cm, low *J. macrescens* and *T. inflata* concentrations throughout	High marsh with increasing salinity
PERFF3c	312-276	-1.37 to -1.01	*J. macrescens* and *T. inflata* rise to a peak of 23 and 14 per cm^3 respectively. *T. inflata* disappears at 276cm.	High marsh with decreasing salinity
PERFF4a	276-258	-1.01 to -0.83	Monospecific *J. macrescens* assemblage. *J. macrescens* disappears at 258cm	High marsh at the edge of marine influence
PERFF4b	258-0	-0.83 to 1.75	Barren of foraminifera to the surface	No marine influence

Zone PERFF1 (450-418cm, -2.75 to -2.43m OD) was barren of foraminifera, indicating the environment was outside marine influence. *Jadammina macrescens* appeared at 418cm (-2.43m OD) and throughout zone PERFF2 (418-398cm, -2.43 to -2.14m OD) a monospecific assemblage of *J. macrescens* rose to a peak. *J. macrescens* is a brackish species common to high marsh, marginal marine saltmarshes worldwide (Murray 1979, Scott and Medioli 1978, Gehrels 1999, Wright *et al.,* 2011). Scott and Medioli (1978) found a monospecific assemblage of *J. macrescens* in a narrow zone at the edge of marine influence at Chezzetcook marsh in Nova Scotia. At Perfeddnant, a monospecific assemblage of *J. macrescens* persisted throughout zone PERFF3a (398-351cm, -2.14 to -1.76m OD) at much lower concentrations compared to zone PERFF2. The presence of a monospecific assemblage of *J. macrescens* at Perfeddnant implies a high marsh environment persisted on the very edge of marine influence from 418 to 354cm (-2.43 to -1.79m OD).

At the onset of zone PERFF3b (354-312cm, -1.76 to -1.37m OD) *Trochammina inflata* appeared alongside *J. macrescens*, and both species continued through subzone PERFF3c (312-276cm, -1.37 to -1.01m OD), rising to a peak at 290cm (-1.15m OD). Murray (1979) described *T. inflata* as a high marsh, sediment dwelling species with wide environmental tolerances. The co-appearance of *T. inflata* and *J. macrescens* is common in saltmarshes globally and is indicative of a high marsh environment (Murray 1979, Gehrels 1999, Horton *et al.,* 1999, Wright *et al.,* 2011, Kemp *et al.,* 2015). The high marsh species *Millamina fusca* is absent from Perfeddnant, however, despite often being found in conjunction with *T. inflata* and *J. macrescens* (Horton *et al.,* 1999). It has been suggested that *M. fusca* is more susceptible to degradation than *T. inflata* and *J. macrescens*, so care must be taken (Horton *et al.,* 1999; De Rijk and Troelstra, 1999; Goldstein and Watkins, 1999). *T. inflata* disappeared at the end of subzone PERFF3c at 276cm (-1.01m OD). Subzone PERFF4a (276-258cm, -1.01 to -0.83m OD) reflected a low concentration, monospecific assemblage of *J. macrescens* before all foraminifera disappeared in subzone PERFF4b (258-0cm, -0.83 to 1.75m OD). The presence of high marsh foraminifera throughout the profile implied that sedimentation rates kept pace with sea-level rise (Barlow *et al.,* 2013; Gehrels *et al.,* 2006).

5.3.6 Diatom analysis

Subsamples for diatom analysis were taken at 8cm intervals throughout the clay, with samples taken at 4 and 2cm over stratigraphic boundaries (Figures 5.68, 5.69, Table 5.47). Where possible, 250 valves were counted, however very poor preservation made this impossible for much of the silt. Very few diatoms were found between -2.33 to -0.51 with no diatoms above -0.43m OD. Two preparation methods were used which produced the same results (see Chapter 3.3.4) which confirmed that the poor preservation was not caused by the preparation procedure. Between 408 and 220 cm (-2.33 to -0.45m OD), only broken valves *of Diploneis interrupta* and *Tryblionella navicularis*, as well as a broken valve of *Auliscus sculptus,* were found. Fifteen diatom species were identified in total, with 11 identified only to genus level due to poor preservation. Due to very low counts, CONISS was not used to establish zones.

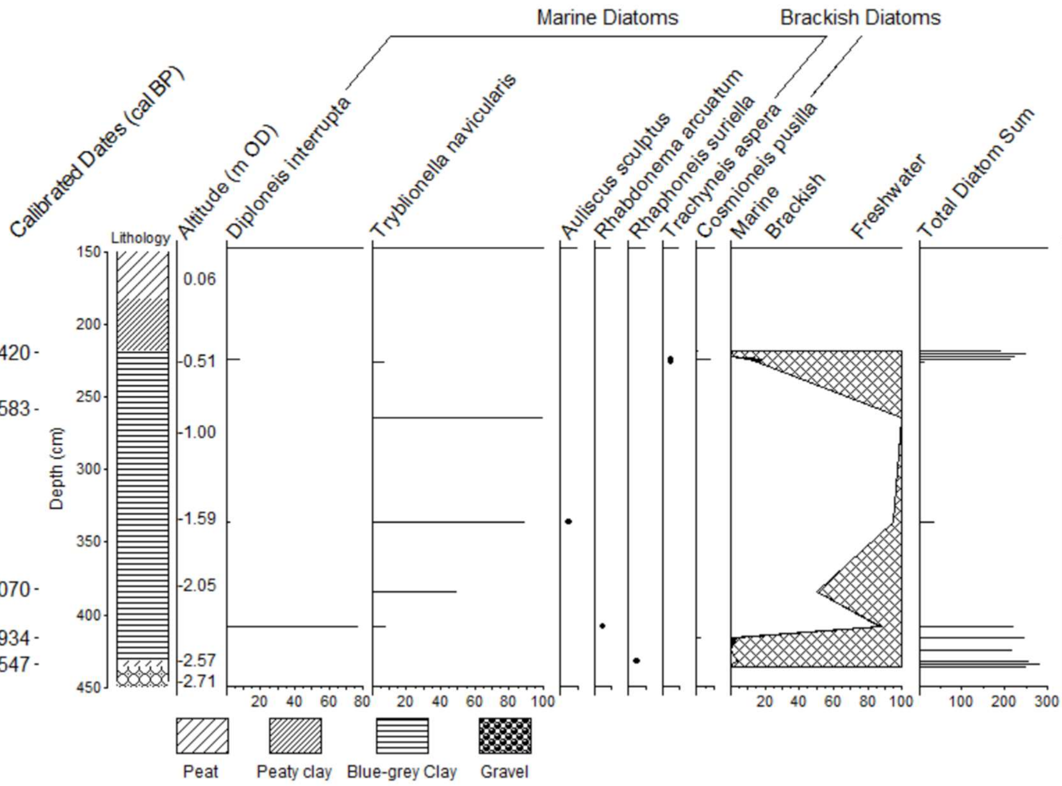


Figure 5.68: Diatom Summary Percentage Diagram from Perfeddnant showing Marine and Brackish diatoms with stratigraphy and radiocarbon dates

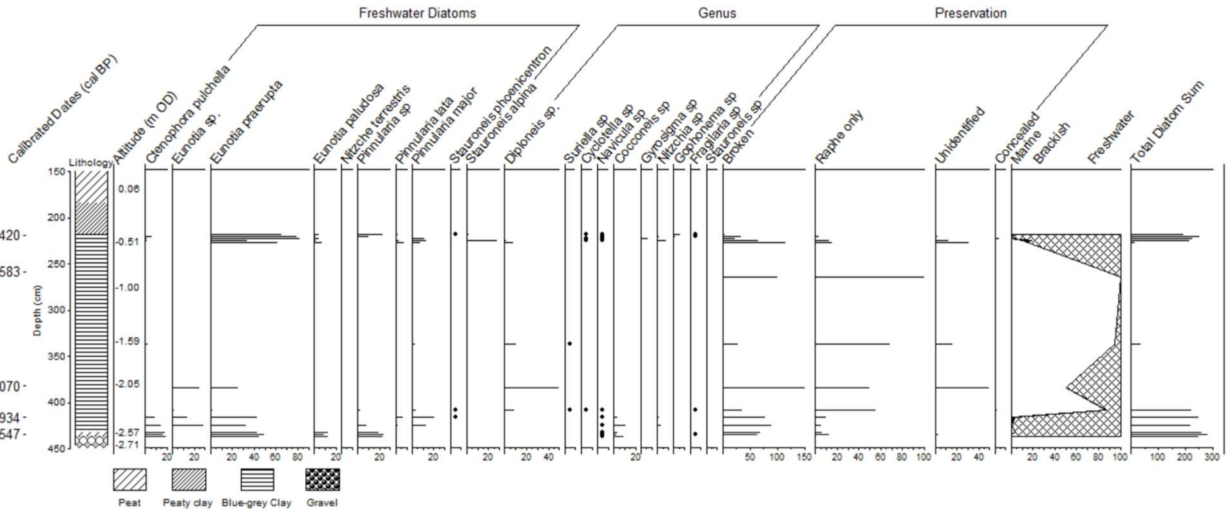


Figure 5.69: Diatom summary percentage diagram from Perfeddnant with stratigraphy, preservation plot and radiocarbon dates.

Table 5.47: Diatom zone summaries for Perfeddnant

Depth (cm)	Altitude (m OD)	Description	Environmental Conditions
436-409	-2.61 to -2.34	Dominated by *Ctenophora pulchella,* as well as *Eunotia praerupta, Eunotia paludosa, Pinnularia* sp., *Pinnularia major* and *Pinnularia lata.* Some *Stauroneis alpina* and *Cocconies* sp were found as well as one broken valve of the marine *Rhaphoneis surirella.*	Freshwater and weakly brackish with some tidal influence washing marine species onto the marsh
408	-2.33	A count of 200-250 was possible. Dominated by *Diploneis interrupta* with some *Tryblionella navicularis. Navicula pussila* was present as well one valve of *Rhabdonema arcuatum.* Freshwater species *C. pulchella* and *S. phoenicentron* present as well as *Nitzchia terrestris, E. praerupta, E. paludosa, P. major, P. lata* and *S. alpina.*	Brackish, saltmarsh environment with freshwater conditions nearby. Care must be taken as it is only one sample.
408-218	-2.33 to -0.51	Preservation very poor, very few diatoms found apart from broken *D. interrupta* and *T. navicularis*, and one broken valve of *Auliscus sculptus*	Possibly brackish conditions with tidal action washing marine species onto the marsh.
226to 218	-0.51m OD to -0.43	Preservation improved, dominated by *C. pulchella* and *E. praerupta*, as well as *E. paludosa*, *P. lata*, *P. major, S. phoenicentron* and *S. alpina. Cosmioneis pusilla, D. interrupta* and *T. navicularis* were also present, as well as *Trachyneis aspera*	Freshwater and weakly brackish marsh
218cm to surface	-0.43 to +1.76	No diatoms found	

Between 436-409cm (-2.61 to -2.34m OD) the diatom assemblage was dominated by *Ctenophora pulchella*, which is a freshwater-brackish epiphytic species (Roe *et al.,* 2009, Williams and Round 1986). *Stauroneis phonicentron* was also found, which is an oligohalobous benthic species (Denys 1994).

The profile was predominantly terrestrial and freshwater between 436-409cm (-2.61 to -2.34m OD), for example the stenohaline, epontic species *Eunotia praerupta* was commonly found (Denys 1991/2, Van dam *et al.,* 1994). *Eunotia paludosa* was also identified, which is found in mossy, moist, temporarily dry environments (Liu *et al.,* 2011, Van der vijver and Beyens 1998). A number of

Pinnularia sp. valves were also present, although due to poor preservation many could not be identified to species level. *Pinnularia major* was identified, which is an oligotrophic species common in lentic water and peatlands (Noga *et al.,* 2014), as well as *Pinnularia lata*, which is a terrestrial species common to cool, oxygen-rich, oligotrophic waters (Krammer 2000). The terrestrial, freshwater *Stauroneis alpina* was recovered (Kociolek *et al.,* 2018) as well as *Cocconeis* sp., but due to preservation it could not be identified to species level. The assemblage at Perfeddnant between 436-409cm (-2.61 to -2.34m OD) indicates a terrestrial, weakly brackish environment with some mosses and dry areas indicated by *E. paludosa*. In contrast, one broken valve of *Rhaphoneis surirella* was found, which is a polyhalobous sessile species (Denys 1991/2), which could be allochthonous.

At 408cm (-2.33m OD) a diatom count between 200-250 was possible, but preservation was very poor with 93.24% of valves broken. The assemblage was dominated by *Diploneis interrupta* with some *Tryblionella navicularis*, which are marine-brackish diatoms typical of saltmarshes (Denys 1991/2, 1994, Vos and de Wolf 1988, 1993). The brackish *Navicula pussila* was also found, which is an oligohalobous benthic species common to saltmarshes (Denys 1991/2). These are robust species that preserve well, however, and it is likely that more fragile diatoms may have been lost. An individual valve of *Rhabdonema arcuatum* was identified, which is a polyhalobous, epontic species (Denys 1991/2). The weakly brackish *C. pulchella* and *S. phoenicentron* remained present, as well as *Nitzchia terrestris* which is an oligohalobous, stenohaline species (Denys 1991/2). Other terrestrial freshwater species were found including *E. praerupta, E. paludosa, P. major, P. lata* and *S. alpina.* The assemblage at 408cm (-2.33m OD) at Perfeddnant implies that marine-brackish saltmarsh conditions developed, with some terrestrial, freshwater environments continuing. The preservation is very poor, however, and as this is a single sample, interpretations must be considered with care. Between 407 and 226 cm (-2.32 to -0.51m OD) diatom counts did not exceed 38 valves. Some broken *D. interrupta* and *T. navicularis* were found, as well as one broken *Auliscus sculptus*, which is a polyhalobous tychoplankton (Denys 1991/2).

Between 226cm to 218cm (-0.51m OD to -0.43m OD) preservation returned, with a profile dominated by *C. pulchella* and *E. praerupta*, as well as *E. paludosa*, *P. lata*, *P. major, S. phoenicentron* and *S. alpina*, indicating a return to freshwater conditions. There was evidence for continuing brackish conditions as *C. pusilla, D. interrupta* and *T. navicularis* were also present, as well as the polyhalobous, epipelic *Trachyneis aspera* (Denys 1994). No diatoms were found above 218cm (-0.43m OD).

The diatom assemblage from Perfeddnant implies that a terrestrial, freshwater environment persisted until 409cm (-2.34m OD) which is well above the dated peat-clay boundary. Furthermore, terrestrial conditions returned at 226cm (-0.51m OD) which is also below the dated upper clay-peat boundary.

5.4 Statistical analysis

5.4.1 Transfer functions

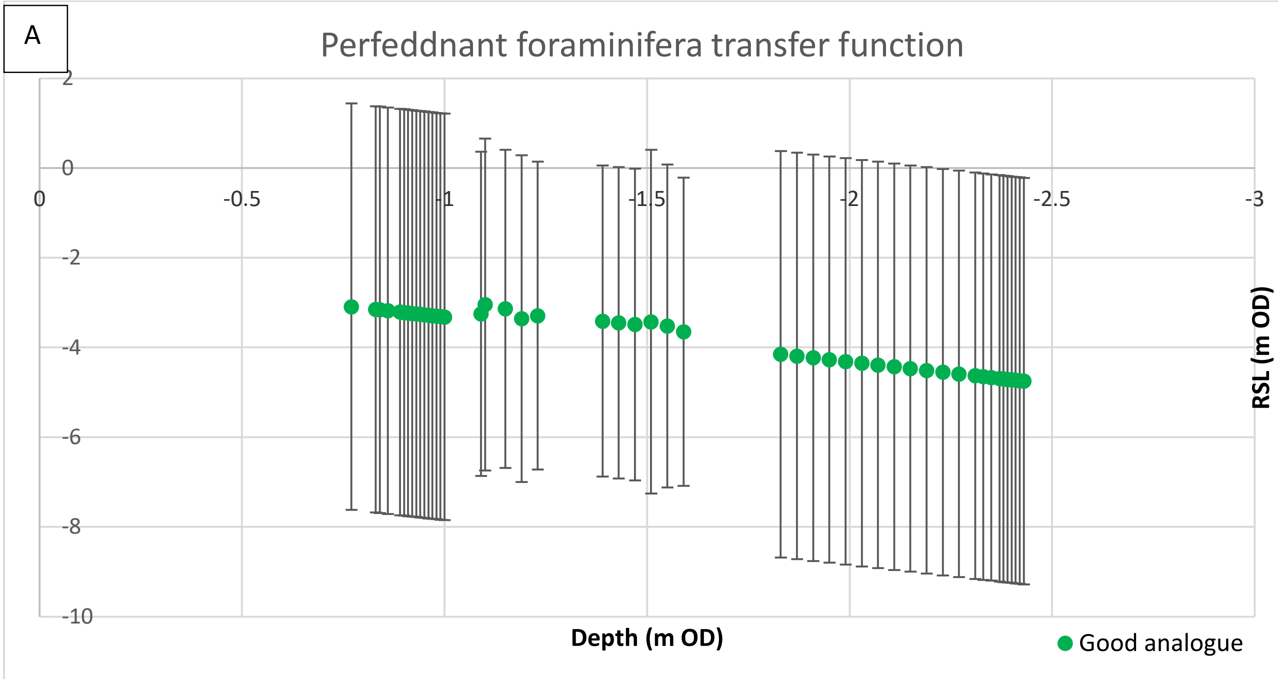
All three transfer functions were applied to the data from Perfeddnant, and the modern analogue technique (MAT) was used to establish the relationship between the fossil samples and the environment. Although there were both diatoms and foraminifera in the Perfeddnant core, there were many problems with preservation which affected the transfer function results.

5.4.1.1 Foraminifera transfer function.

The results of the foraminifera transfer function appear to show a rising sea level from -5 to -3m OD (Figure 5.70). Many samples had very low counts and samples with less than 30 tests were excluded, apart from the monospecific *J. macrescens* assemblages following Horton and Edwards (2006), due to the low species diversity and strong relationship to sea-level.

It is possible that there are preservation problems in the Perfeddnant core, as no fragile species such as *M. fusca* were found, in contrast to the Penllyn core (see Chapter 4). Furthermore, dissolution of calcareous species in saltmarsh sediments has been well recorded (Gehrels *et al.,* 2001; Horton and

Edwards, 2001, 2006) and if the species such as *Ammonia* spp. has dissolved, a low marsh environment would be missed. The results of this transfer function should be treated with caution.



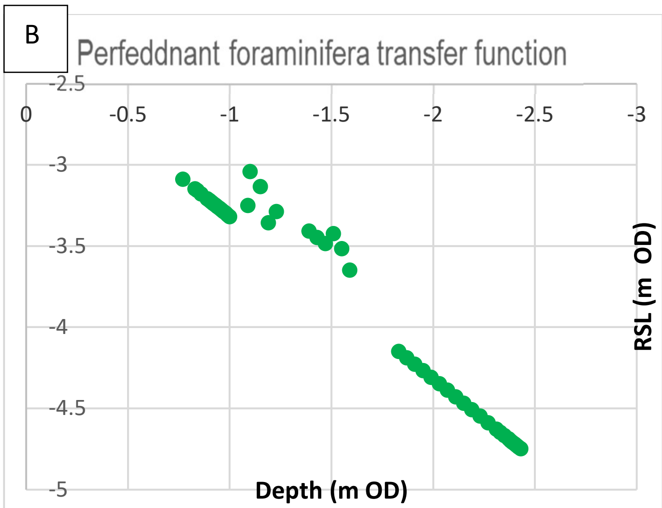
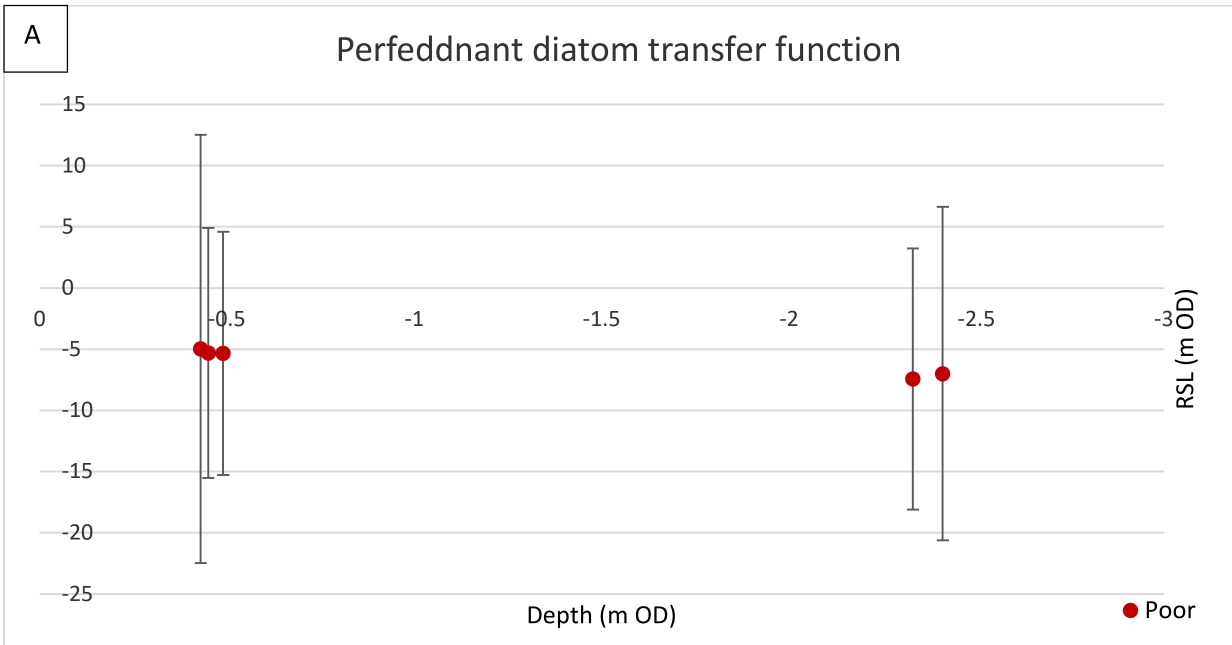


Figure 5.70: A: Sea-level reconstruction from the foraminifera transfer function at Perfeddnant with error bars.

B: Sea-level reconstruction from the foraminifera transfer function at Perfeddnant without error bars.

5.4.1.2 Diatom Transfer function

Very few diatom samples had sufficient counts to produce a transfer function reconstruction, particularly between 220 and 408cm (-0.44 and -2.32m OD), where predominantly broken valves were found (Figure 5.71). Furthermore, samples dominated by freshwater species did not produce any transfer function reconstructions. There are no good analogues for the few samples that did produce a reconstruction, and the diatom results plot much lower than the foraminifera, similar to Penllyn.



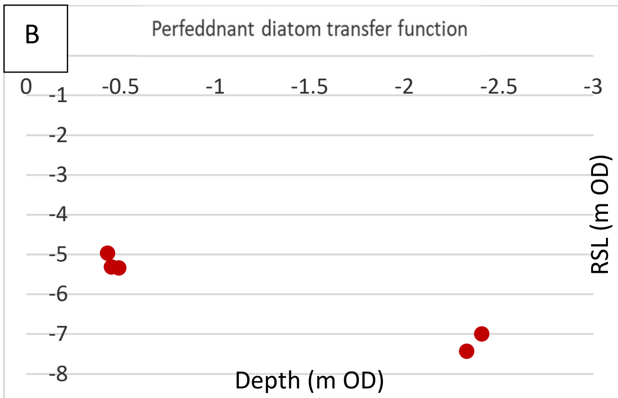
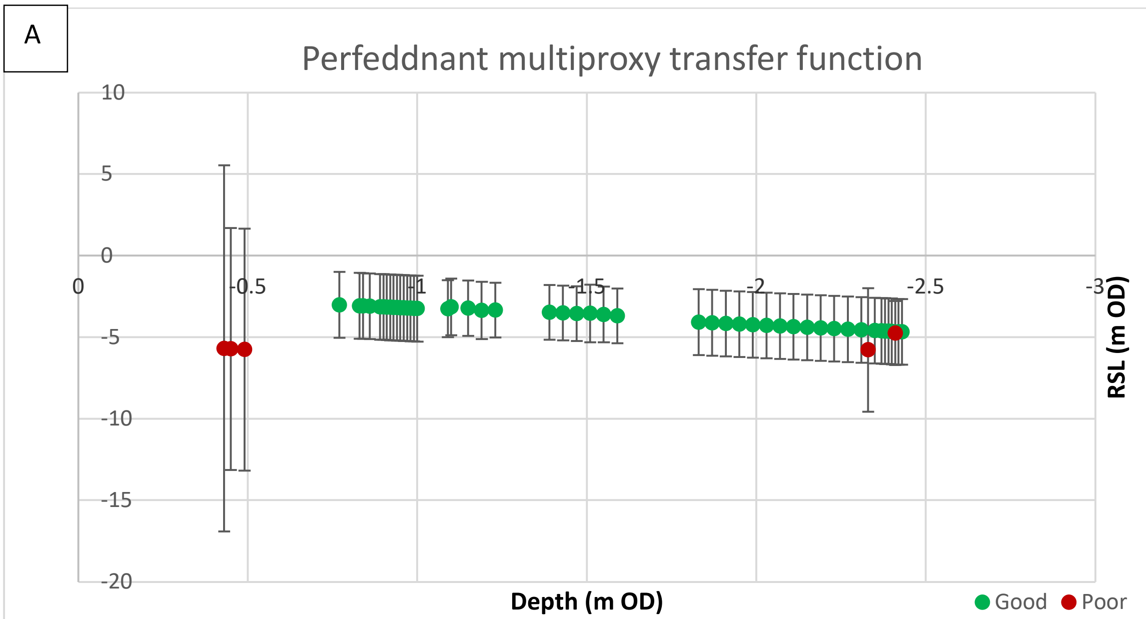


Figure 5.71: A: Sea-level reconstruction from the diatom transfer function at Perfeddnant with error bars.

B: Sea-level reconstruction from the diatom transfer function at Perfeddnant without error bars.

5.4.1.3 Multiproxy transfer function

Within the multiproxy transfer function there is a significant offset between the samples with diatoms and the samples with foraminifera between -0.49 and -0.45m OD. In contrast, the two diatom samples at -2.41 and -2.22m OD plot within the error range of the foraminifera reconstruction (Figure 5.72). Furthermore, the foraminifera samples have much smaller error bars compared to the diatoms. Given the problems of preservation within the fossil diatom dataset, the multiproxy transfer function cannot be used.



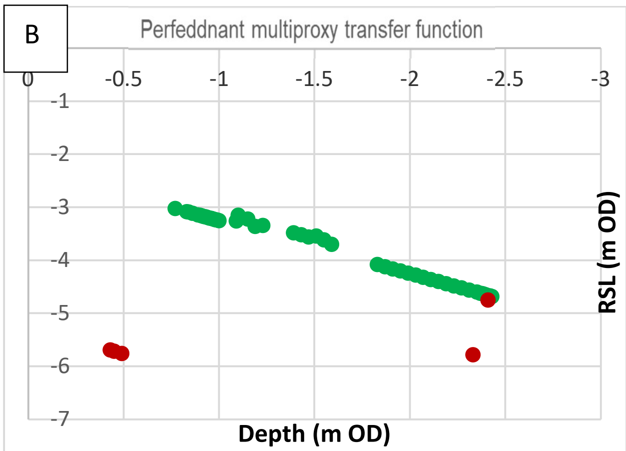


Figure 5.72: A: Sea-level reconstruction from the multiproxy transfer function at Perfeddnant with error bars

B: Sea-level reconstruction from the multiproxy transfer function at Perfeddnant without error bars.

5.4.2 Sea-level index points and limiting points

To calculate sea-level index points the altitude, indicative range, tendency, and age are needed (see method section). At Perfeddnant, only two dates were accepted due to reversals. The calibrated median age, as well as the positive and negative range were established using Oxcal Online v4.4, calibration curve IntCal20 (Reimer *et al.,* 2020; Bronk Ramsey, 2009a).

Table 5.48: Radiocarbon dates from Perfeddnant, with calibrated ages and positive/negative error range calculated using Oxcal Online v4.4

Depth (cm)	Elevation (m OD)	Radiocarbon age		Calibrated Age (cal. BP, median)	Positive error range	Negative error range
		BP	1 σ error			
219	-0.44	4034	35	4491	292	75
434	-2.59	8711	40	9652	237	107

At-2.59mOD, no foraminifera were found with only freshwater diatoms, such as *Ctenophora pulchella, Eunotia praerupta* and *Eunotia paludosa*. The transfer function could not be applied, and as the sample was freshwater a limiting point can be established.

At -2.41m OD a bulk date of 8150-7883 cal. BP D-AMS 036268 was established, which was out of sequence. The humin and humic aciddates were very different (Humin: 11127-10734 D-AMS 037473; Humic acid: 8397-8201 cal. BP D-AMS 037474), and therefore could not be used. The foraminifera transfer function produced an indicative meaning of -4.728m OD MHWST. A monospecific *J. macrescens* assemblage indicated a high marsh environment, between HAT and MTL, which was calculated as -4.2225m OD MHWST, similar to the foraminifera. In contrast, the diatom transfer function produced an indicative meaning of -6.98m OD MHWST. However, there were no modern analogues in the diatom training set, and the sample was dominated by the freshwater *E. praerupta*, with *Ctenophora pulchella*, so the transfer function could not be used.

At -0.83m OD a bulk date of 9695-9532 cal. BP (D-AMS 036266) was produced, which was also out of sequence. Humin and humic acid dates were produced, which were very different (Humin: 16366-

15971, D-AMS 037471; Humic acid: 8338-8170, D-AMS 037472), therefore the date could not be used. No diatoms were preserved, but the foraminifera transfer function was applied, producing an indicative meaning of -3.148m OD MHWST. A monospecific *J. macrescens* assemblage indicated a high marsh environment. Following Hijma *et al.,* (2015), the indicative meaning was calculated as -2.6425m OD MHWST. However, with no dating evidence a SLIP could not be produced.

At -0.44m OD, no foraminifera were preserved. The diatom transfer function appeared to show an indicative meaning of -4.95m OD MHWST. However, no modern analogues were found in the training set, and the diatom profile was dominated by the freshwater *Eunotia praerupta*, therefore the profile was freshwater, and can be established as a limiting point.

According to the foraminifera transfer function sea level rose from -4.74 to -3.03m OD MHWST between -2.43m OD and -0.77m OD in the core. In contrast, the diatom transfer function showed a rise from -6.98m OD MHWST to-4.95m OD MHWST between -2.41m and -0.43m OD in the core. The foraminifera and diatom microfossil evidence indicated a rise in sea-level from -4.2225 to -2.6425m OD MHWST between -2.41m OD and -0.82m OD in the core. No SLIPs could be calculated, however, due to problems of dating. Two freshwater limiting points were calculated between 9652 and 4491 cal. BP from -2.59 to -0.44m OD.

Table 5.49: SLIP and limiting points from Perfeddnant, with errors and ages.

elevation	RWL	SLIP	Upward Error	Downward error	Calibrated Age (cal. BP, median)	Positive error range	Negative error range
-0.44	-	-0.44	2.9066	0.2191	4491	292	75
-0.83	1.8225	-2.6425	2.4191	0.2191	-	-	
-2.41	1.8225	-4.2225	0.4441	0.2191	-	-	
-2.59	-	-2.6	0.21905	0.21905	9652	237	107

5.5 Environmental Change at Perfeddnant

A glacial gravel was found across the Dysynni Valley with poorly sorted, angular shale pieces and occasional quartz inclusions. The upper level of the gravel was encountered at Penllyn between -4.0 and 0.5 m OD, at Peniarth and Perfeddnant between -6.0 and -1.0 m OD and at Gesail between -1.5 and 3.0 m OD. The sample core at Perfeddnant revealed the glacial gravel between -3.25 to -2.61m OD (500 to 436 cm) and corresponds with descriptions of glacial till recorded in BGS boreholes at the drumlin in Tywyn (Leng and Pratt, 1987).

An intermittent layer of basal peat was found overlying the glacial gravel at some locations at Perfeddnant. In the sample core, the basal peat was found between 436-435cm (-2.61 to -2.60m OD) and dated to 9,885-9,547 cal. BP (D-AMS 031520). Freshwater conditions were prevalent, which can be seen from the lack of foraminifera and the dominance of freshwater diatoms such as *C. pulchella*. The presence of *E. paludosa* indicates the occurrence of mosses and periodically dry areas (Liu *et al.,* 2011; Van der vijver and Beyens, 1998).

Overlying the basal peat at Perfeddnant was blue-grey estuarine silt, however at many locations the peat was not present, and the clay overlies the glacial gravel directly. It is probable that isolated pockets of peat existed within the undulating gravel environment, however it may also be possible that fluvial erosion has taken place, removing much of the peat in the valley bottom. In the sample core, freshwater conditions persisted into the overlying estuarine silt up to 418cm (-2.43m OD). At 418cm (-2.43m OD) conditions remained predominantly freshwater, indicated by the persisting freshwater diatom assemblage. A brackish saltmarsh began to develop with the onset of marginal marine conditions indicated by very low counts of the high marsh foraminifera *J. macrescens* (Scott and Medioli, 1978; Gehrels, 1999).

At 409cm (-2.34m OD) brackish saltmarsh conditions dominated, indicated by high concentrations of a monospecific, *J. macrescens* foraminifera assemblage. The freshwater diatoms declined, with some weakly brackish species recovered such as *S. phoenicentron.* However, brackish species such as *D.*

interrupta, T. navicularis and *C. pusilla* were prevalent, as well as a single valve of the marine species *R. arcuatum*. Care must be taken however, as the diatom preservation was very poor, with very low counts and many broken valves.

From 409cm (-2.34m OD) there was evidence for a brackish, marginal marine saltmarsh, with both aerial exposure and flooding episodes from tidal action, river flooding or slope wash. High concentrations of a monospecific, *J. macrescens* foraminifera assemblage combined with brackish diatom species such as *D. interrupta, T. navicularis* and *C. pusilla* indicated the presence of a marginal marine, brackish saltmarsh. Diatom preservation was very poor, however, with many broken valves and poor concentrations. Poor preservation could be caused by aerial exposure, indicated by iron banding as well as the presence of *J. macrescens* which can survive exposure (Murray, 1989) and the aerophilic diatom species *D. interrupta* (Denys 1991/2). It is also possible for transportation to cause broken diatoms, which is supported by the presence of low levels of sand within the clay, indicating a high-energy input (de Groot *et al.,* 2011). At 358cm (-1.83m OD) there was a small peak in sand which correlates with an increase in organic content, which could potentially indicate inwash of sediment. Low levels of freshwater diatoms were present, such as *C. pulchella, S. phoenicentron, E. praerupta, E. paludosa, P. major, P. lata* and *S. alpina*; as well as one valve of the marine species *R. arcuatum*. It is possible that the site was influenced by tidal action, washing marine diatoms onto the saltmarsh and either slope wash or fluvial input bringing freshwater diatoms and sand downstream.

At 354cm (-1.79m OD) high marsh conditions persisted, with evidence for an increase in marine conditions with the presence of both *J. macrescens* and *T. inflata* as well as some poorly preserved, broken brackish diatoms such as *T. navicularis* and one broken valve of *A. sculptus*. At 276cm (-1.01m OD) brackish saltmarsh conditions continued, however marine influence began to decline with first *T. inflata* then *J. macrescens* disappearing at 276cm (-1.01m OD) and 258cm (-0.83 m OD) respectively. No foraminifera were found above 258cm (-0.83 m OD), and poor diatom preservation continued until 226cm (-0.51m OD). It is possible that 258cm (-0.83 m OD) marks the end of marine conditions, but it is also conceivable that erosion or disturbance occurred.

At 226cm (-0.51m OD) evidence for freshwater conditions returned with diatom species such as *C. pulchella* returning. There were low percentages of brackish diatoms present such as *C. pusilla, D. interrupta, T. navicularis* and *T. aspera,* which may have been allochthonous as all the valves were broken. At 219cm (-0.44m OD) a date of 4782-4420 cal. BP (D-AMS 031519) was obtained and above 218 cm (-0.43m OD) no diatoms of foraminifera were found. The blue-grey silt continued with a small percentage of sand, indicating low-energy, still water was dominant with some high energy input either from river flooding or tidal action.

At 214cm (-0.39m OD) conditions shifted towards a more terrestrial environment, with a transition to peaty clay and an increase in organic content to 40%. At Penllyn, the transition from blue-grey silt to silty peat is at a similar altitude, between 0 and -1m OD (Chapter 4.3.1). The same transition from minerogenic, marine clay to freshwater, terrestrial peat occurred at a similar altitude across Wales. At Borth, Wilks (1977, 1979) dated the transition from clay to peat at 6179-5663 cal. BP (HAR1020) and 5586 to 5312 cal. BP (HAR 1019). This has been correlated with the lower rate of sea-level rise across Wales, when the sedimentation rate overtook low levels of sea-level rise leading to coastal expansion (Wilks, 1977, 1979).

There is evidence for a potential inwash of sediment at 214cm (-0.39m OD), with a peak in the sand fraction which corresponds with the sudden increase in organic content. At 133cm (0.42m OD) there was a peak in the sand fraction and a small increase in carbonate content, which could indicate a second inwash of sediment. At 139cm (0.36m OD) there was a transition to peat with an organic content between 60-80%, with evidence for fluvial input from the small silt and sand fraction present.

5.6 Conclusions

The gouge survey at Perfeddnant revealed a glacial gravel overlain by an intermittent thin layer of peat. A blue-grey silt was found across the site, overlain by woody peat to the southwest and stiff, grey clay to the northeast. A sample core was taken which represented the sequence of basal peat, blue-grey silt, and woody peat. Loss on ignition, particle size, diatom and foraminifera analyses were undertaken and transfer functions were applied. Two freshwater limiting points were established but no sea-level index points. Nine dates were obtained but an age reversal resulted in only two dates being accepted.

The basal peat was freshwater, dated to 9885-9547 cal. BP (D-AMS 031520). The overlying blue-grey silt contained some brackish diatoms, although preservation was poor, with high marsh foraminifera. Despite multiple attempts, no accurate dates were obtained from the silt. The overlying woody peat was freshwater, dated to 4782-4420 cal. BP (D-AMS 031519).

The site at Perfeddnant gives an indication of the horizontal penetration of marine influence in the Dysynni Valley and established that a high marsh saltmarsh existed, keeping pace with sea-level rise through much of the Holocene. Microfossil preservation was poor, however, and there were problems with dating the estuarine silt, limiting the sea-level and environmental reconstructions possible.

Chapter 6: Gesail

6.1 Introduction

Northeast of Perfeddnant, the valley bottom rises to 4m OD at the base of Craig yr Aderyn (Figure 6.73). The modern river has been canalised with river defences and embankments, but the LiDAR reveals ribboned paleochannels running southwest to northeast following the valley (Figure 6.74, 6.75). Pratt *et al.,* (1995) described estuarine silts underlying river alluvium at the base of Craig yr Aderyn, but no stratigraphic records were cited.

Figure 6.73: Photograph taken from Perfeddnant looking northeast towards Craig yr Aderyn, showing the site of Gesail in the distance, near the line of trees.

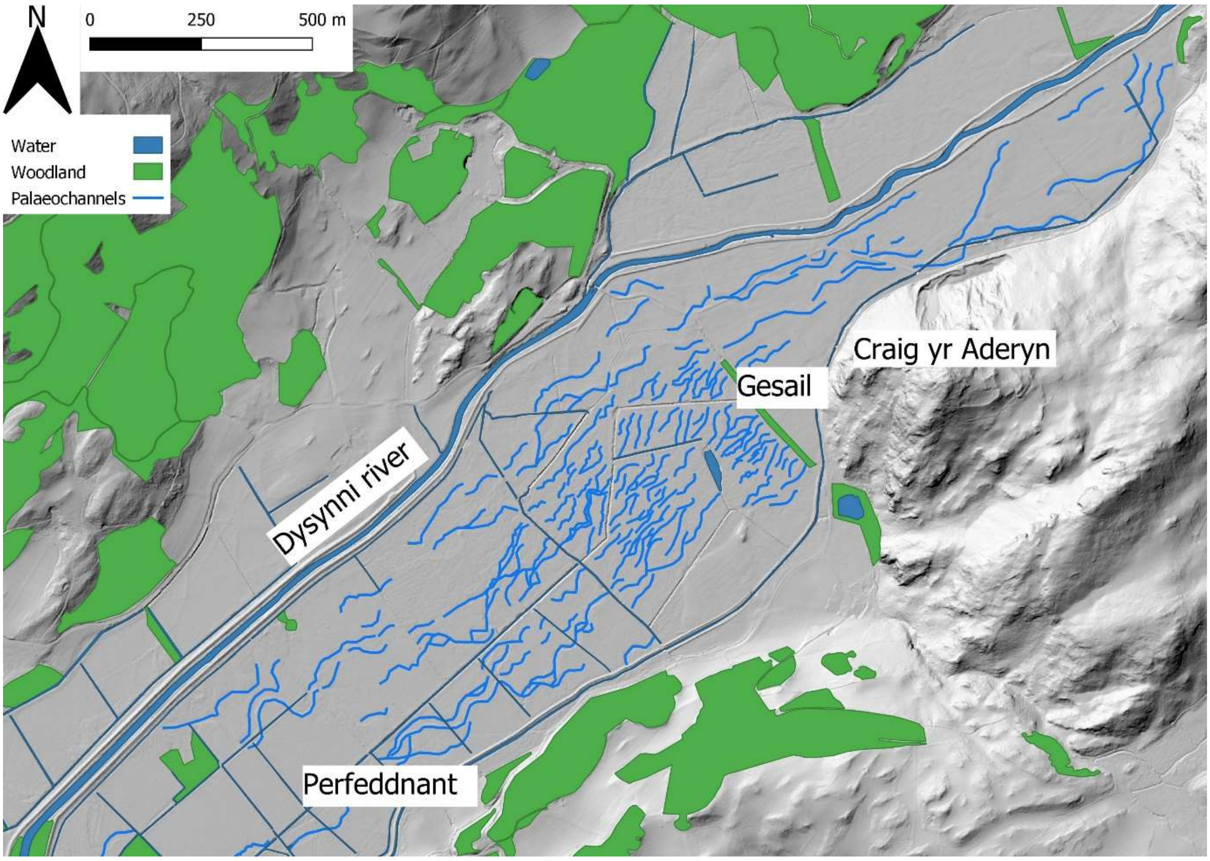


Figure 6.74: Location map of Gesail showing palaeochannels and locations mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved.

At Gesail the valley is predominantly improved grassland with tall, spiky grass in the damper areas, such as the palaeochannels. There are tall grass, reeds, and bracken in the ditches with some small bushes and trees such as oak and willow. There is a narrow strip of woodland across the valley which is predominantly coniferous. Around the field edges, where the ditch fill has been dumped, there is disturbed ground with docks, nettles, and thistles. There are some areas of woodland and individual trees by the road, predominantly mixed oak-birch woodland.

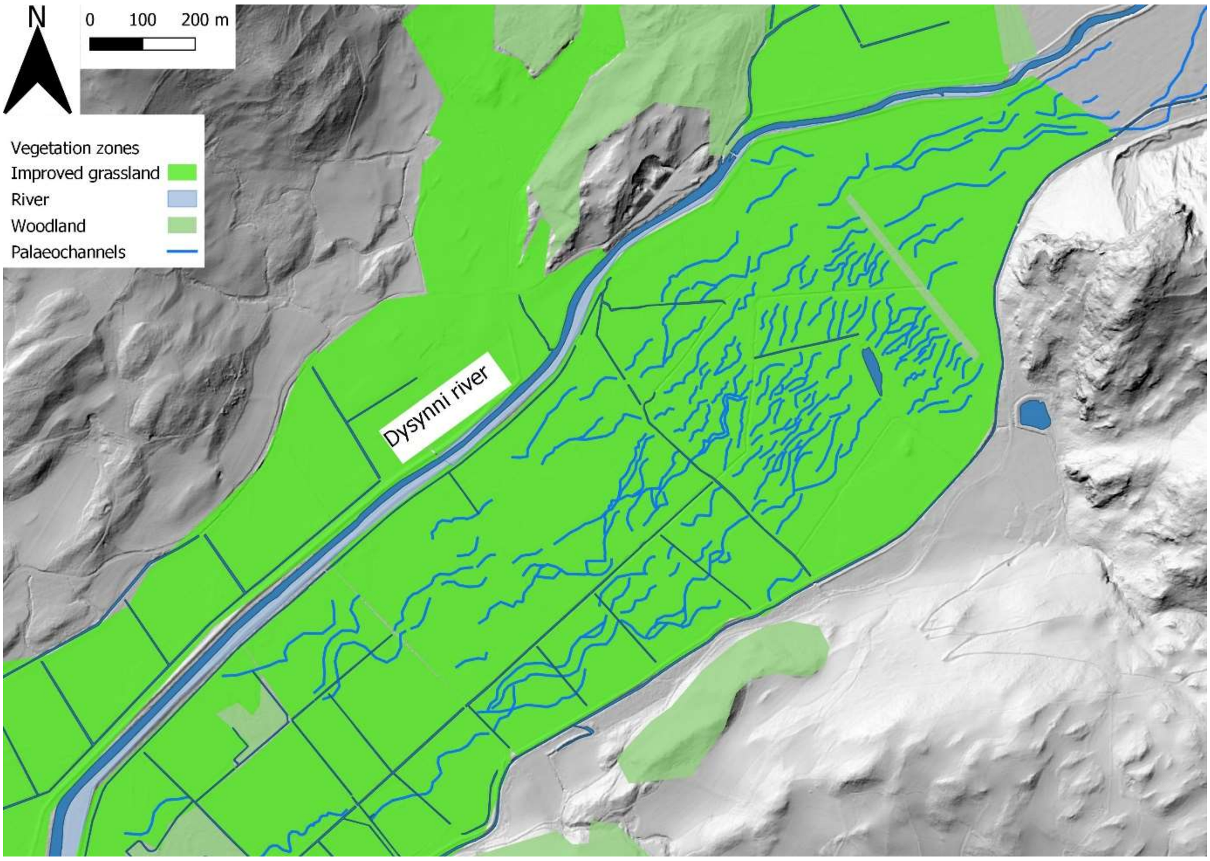


Figure 6.75: Map showing vegetation zones and palaeochannels at Gesail. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved.

6.1.2 Archaeological Evidence

The Iron Age hillfort Craig yr Aderyn, located towards the head of the valley is much larger than the smaller, coastal forts (Figure 6.76) (Chapter 4.1.1), with a greater proportion of stone used (Garrard and Dobson, 1974). Located on a rocky cliff-face, Craig yr Aderyn has elaborately in-turned entrances and a double line of defence (Garrard and Dobson, 1974). Romano British pottery was found with evidence of two distinct periods, with an initial earthen bank followed by an elaborate stone wall.

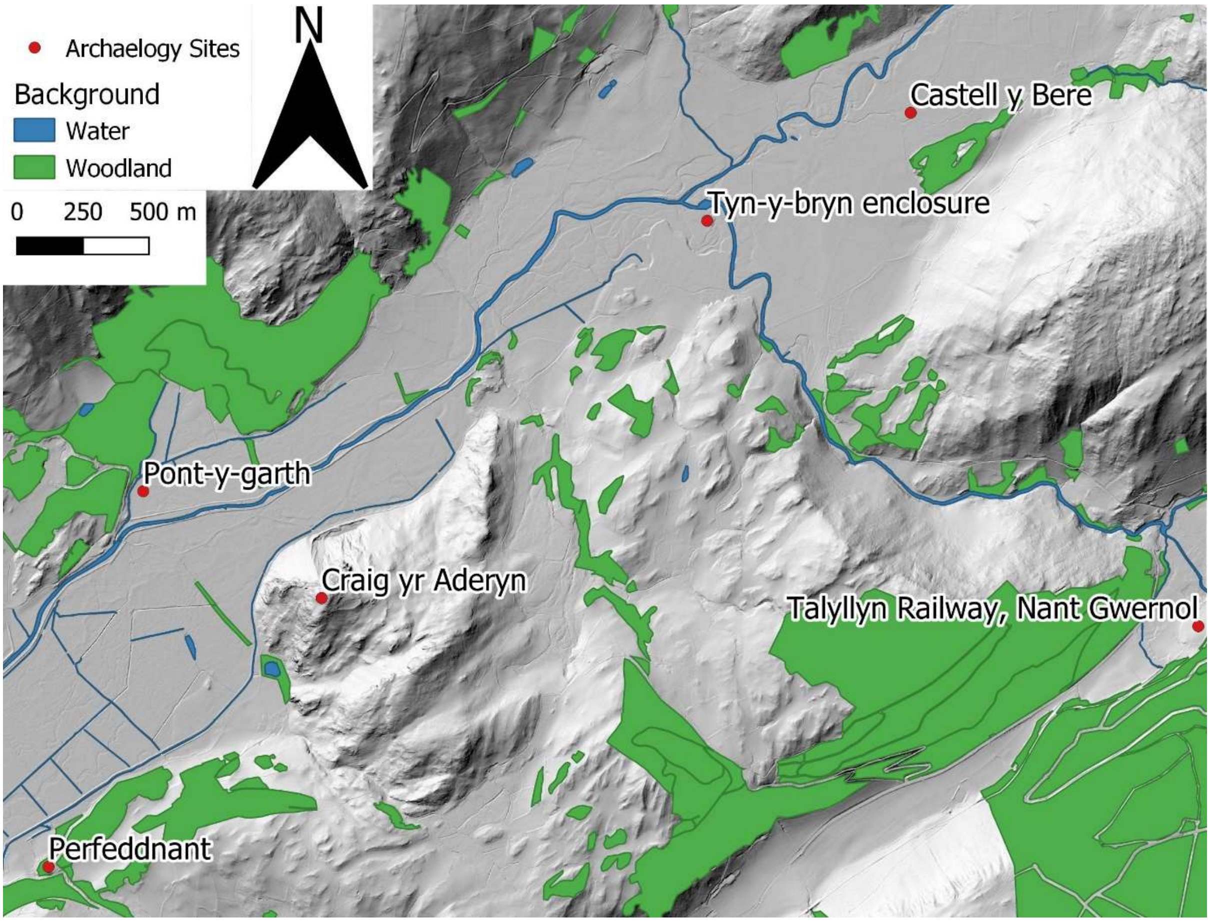


Figure 6.76: Archaeological evidence near Gesail, indicating sites mentioned in the text. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales and Archwilio HER information © Natural Resources Wales and Database

Few Roman finds have been recovered from the Dysynni Valley, with some Roman coins found at Castell y Bere, as well as a hoard of coins reported from Dollgellau in 1695 (Garrard and Dobson, 1974). A defended enclosure was found at Tyn-y-bryn (NPRN 423305). Cropmarks in the inner enclosure of a square stone building would be unusual in a prehistoric although it is unclear if it was Roman or medieval.

The medieval Welsh Castell y Bere (NPRN 93719) was built in 1221 at the head of the Dysynni valley, which was surrendered to Edward I in 1283, although the castle was abandoned in 1294-5 (Avent, 2004). There are record of an associated borough being built by the English, however no archaeological remains have been found.

6.2 Fieldwork

6.2.1 Site stratigraphy

A gouge survey was undertaken along the valley from southeast to northwest to the base of Craig yr Aderyn, where gravel reached the surface and could not be penetrated (Figure 6.77). All efforts were made to avoid coring the palaeochannels.

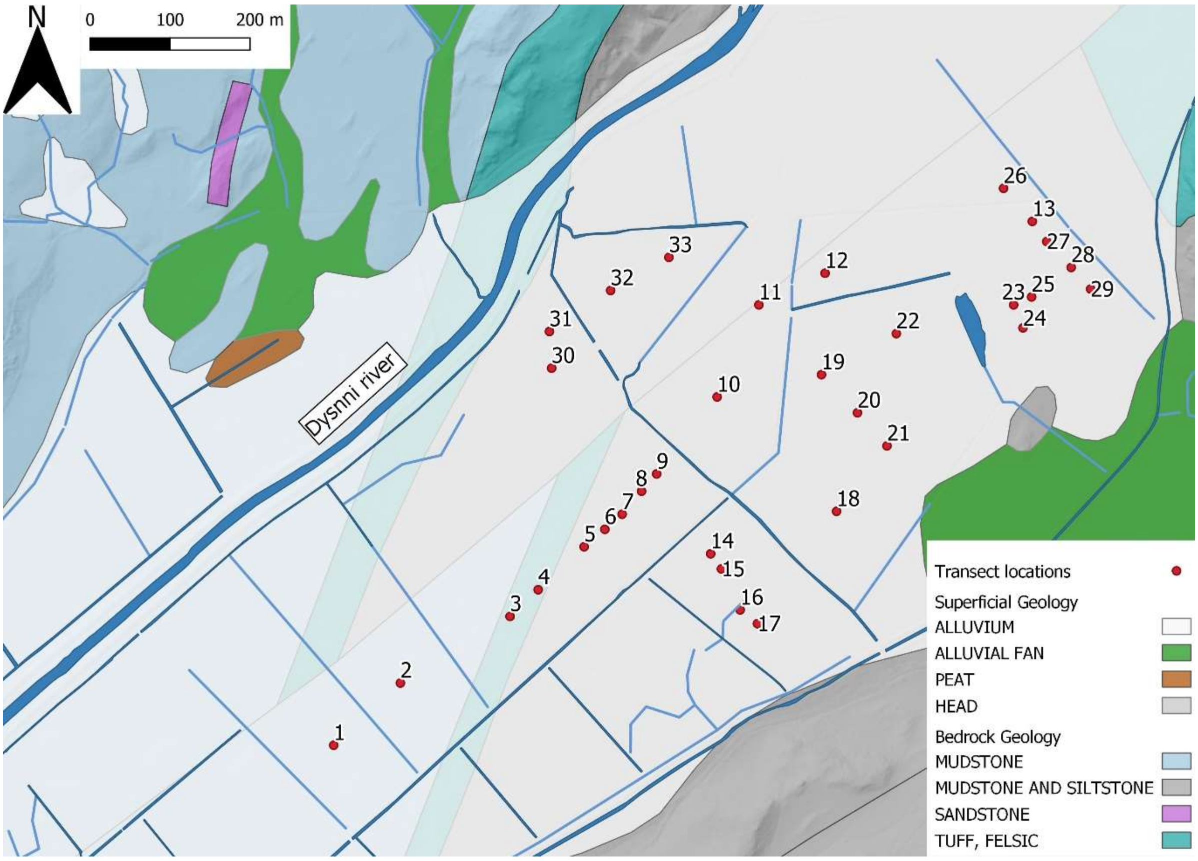


Figure 6.77: Map showing the bedrock and superficial geology with gouge survey locations at Gesail. © Crown copyright and database rights 2022 Ordnance Survey (100025252). Contains Natural Resources Wales information © Natural Resources Wales and Database Right. All rights Reserved. Geological Map Data BGS © UKRI 2022

A basal gravel of angular, poorly sorted shale with occasional quartz inclusions was found from -0.5m OD up to 4m OD near the base of Craig yr Aderyn. This gravel was overlain by grey clay with iron banding, which was soft and elastic towards the base and dry at the surface, possibly due to drainage. Lenses of sand and gravel were present, for example at locations 9 and 11, as well as organic silt layers such as at location 10 (Figures 6.78, 6.79, 6.80, 6.81)

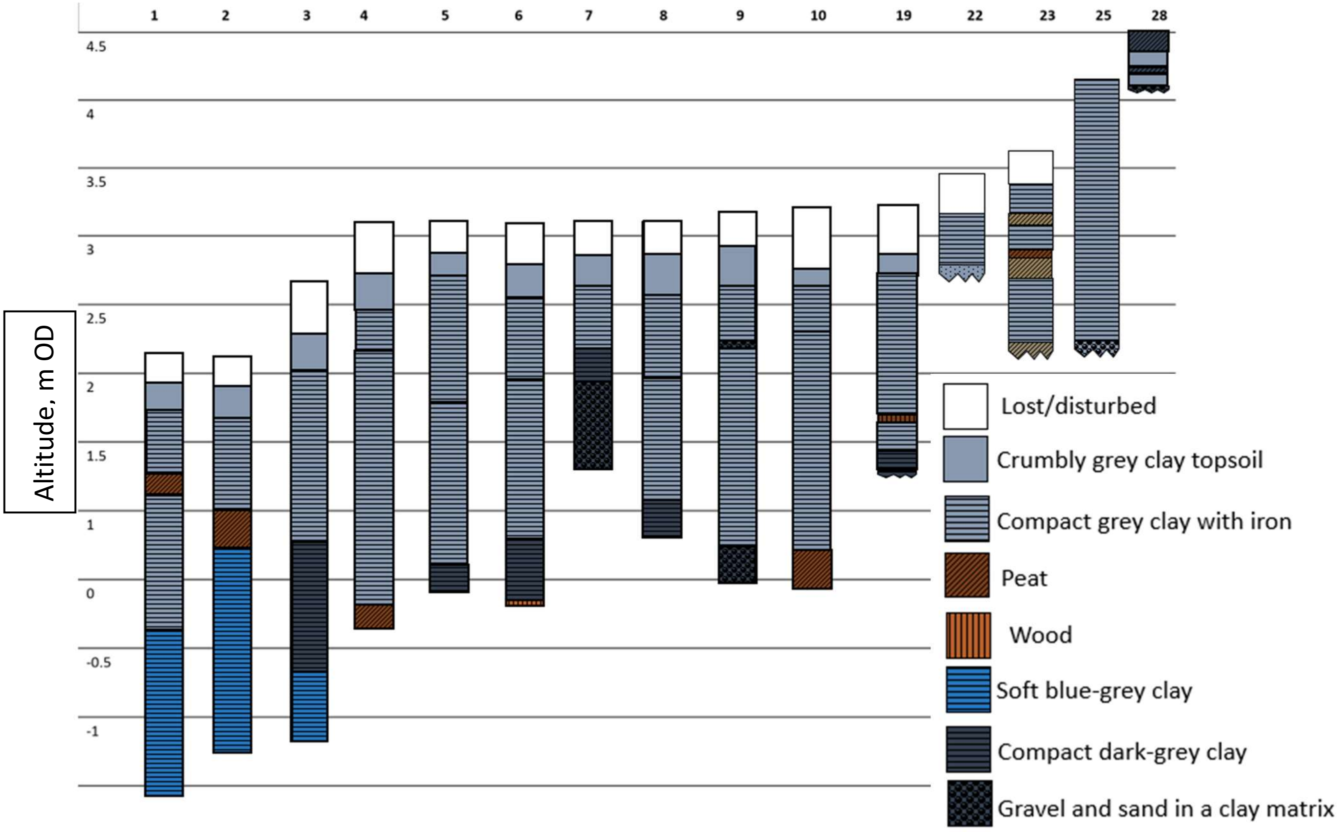


Figure 6.78: Stratigraphic sequence of the transect located southwest to northeast, boreholes 1 to 10, 19, 22, 23, 25 and 28

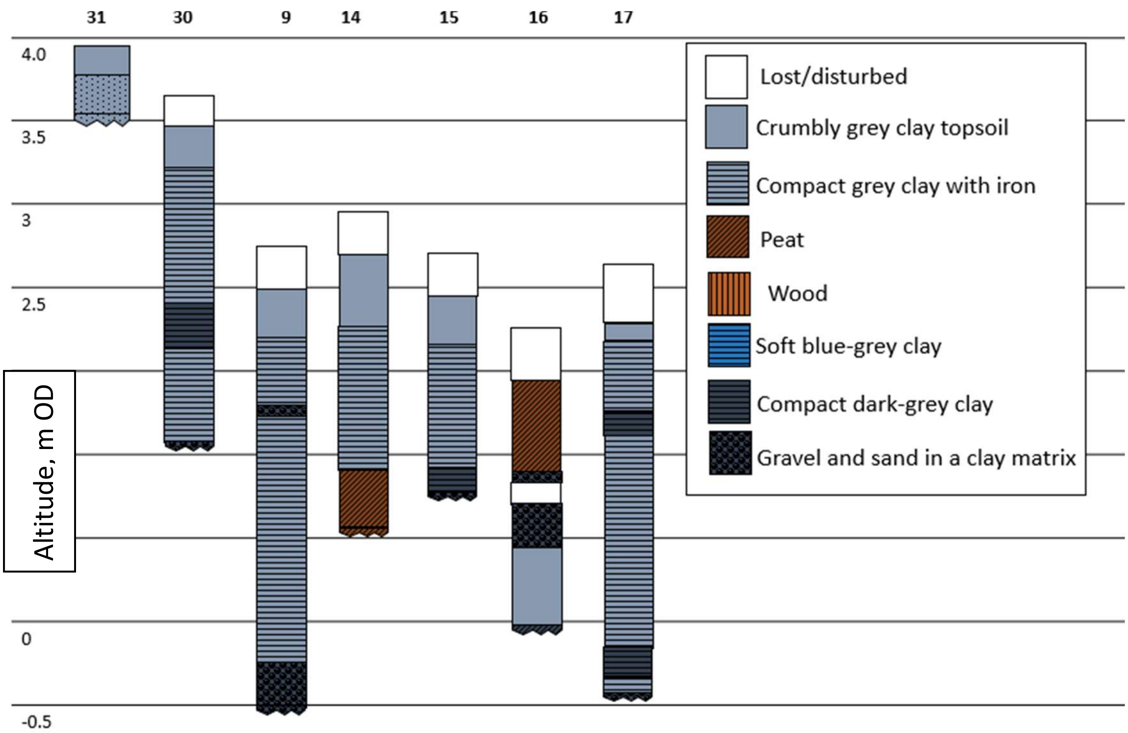


Figure 6.79: Stratigraphic sequence of the transect located northwest to southeast, boreholes 31, 30, 9, 14-17

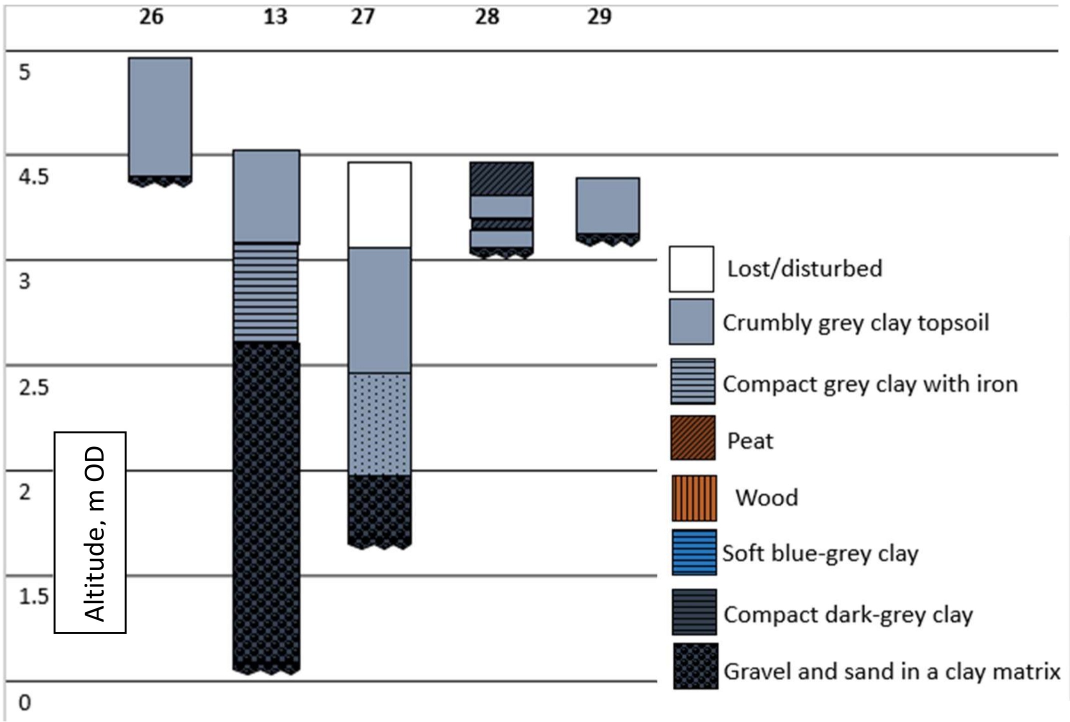


Figure 6.80: Stratigraphic sequence of the transect located northwest to southeast, boreholes 26, 13, 27-29

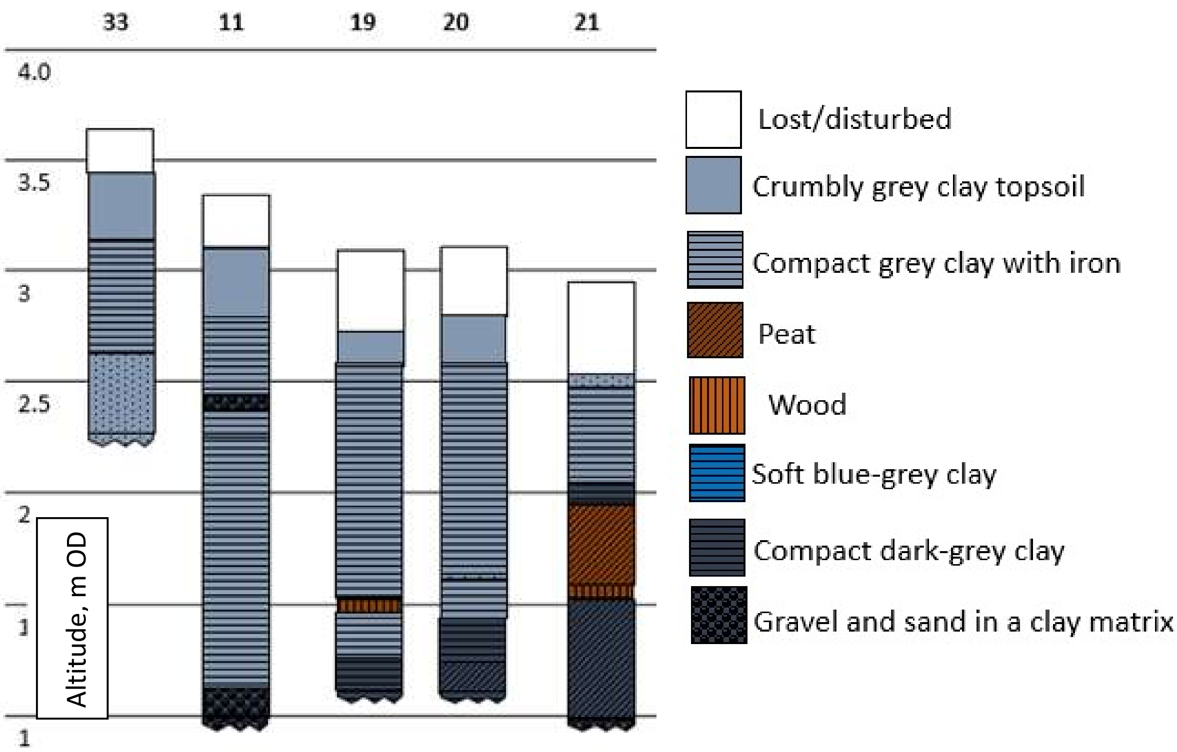


Figure 6.81: Stratigraphic sequence of the transect located northwest to southeast, boreholes 33, 11, 19-21

6.3 Laboratory work

A sample core was taken at location 23 as it represented a complete sequence of clay at the top of the valley, with organic horizons. Subsamples were taken for loss on ignition, particle size, diatom and foraminifera analyses. No radiocarbon dates were submitted as the sequence is freshwater and would not contribute to SLIPs. Pollen analysis was not undertaken due to time constraints caused by COVID-19.

6.3.1 Stratigraphy

Table 6.50: Description of Gesail stratigraphy following Tröels-Smith (1955). Nig=Nigro, Strf-Stratificatio, Elas = Elasticitas, Sicc = Siccitas, Strf = Structura, lim = Limes, Hum = Humositas. See methods section.

Depth (cm)	Altitude (m OD)	Description	Nig	Strf	Elas	Sicc	Colour	Stru	Lim	Hum	Composition
0 to 48	3.5 to 3.02	Clay topsoil, grey crumbly	1+	0	1	4	Grey	Homoge-neous	0	0	2 As, 1 Sh, 1 Th (roots).
48 to 76	3.02 to 2.74	Black peat	3	0	1+	2+	Black	Fibrous	1	2	2 Sh, 1 Ag 1 As plus Th (roots)
76 to 124	2.74 to 2.26	Grey clay	1	0	3	2+	Grey	Homoge-neous	0	0	4 As plus Ag
124 to 145	2.26 to 2.05	Dark grey peaty-organic clay with gravel and fine sand	3	0	1+	1	Dark grey	Homoge-neous	0	1	Sh 2, Ag 1 Ga 1 plus Gg(min) plus Gg(maj) and Th (not roots). Shale pieces/gravel up to 4mm across

6.3.2 Loss on Ignition

Samples for loss on ignition were taken every 4cm throughout the core. Carbonate content was negligible throughout the core (Figure 6.82). The gravel at the base of the core contained 10-30% organic content, falling to 10% throughout the grey silt. In the peat layer, organic content rose to 40-60%, falling again to between 10-20% to the surface. The basal gravel appears to be glacial in origin, similar to Penllyn and Perfeddnant, however it is interesting to note the organic content in the gravel, which could indicate a fluvial origin.

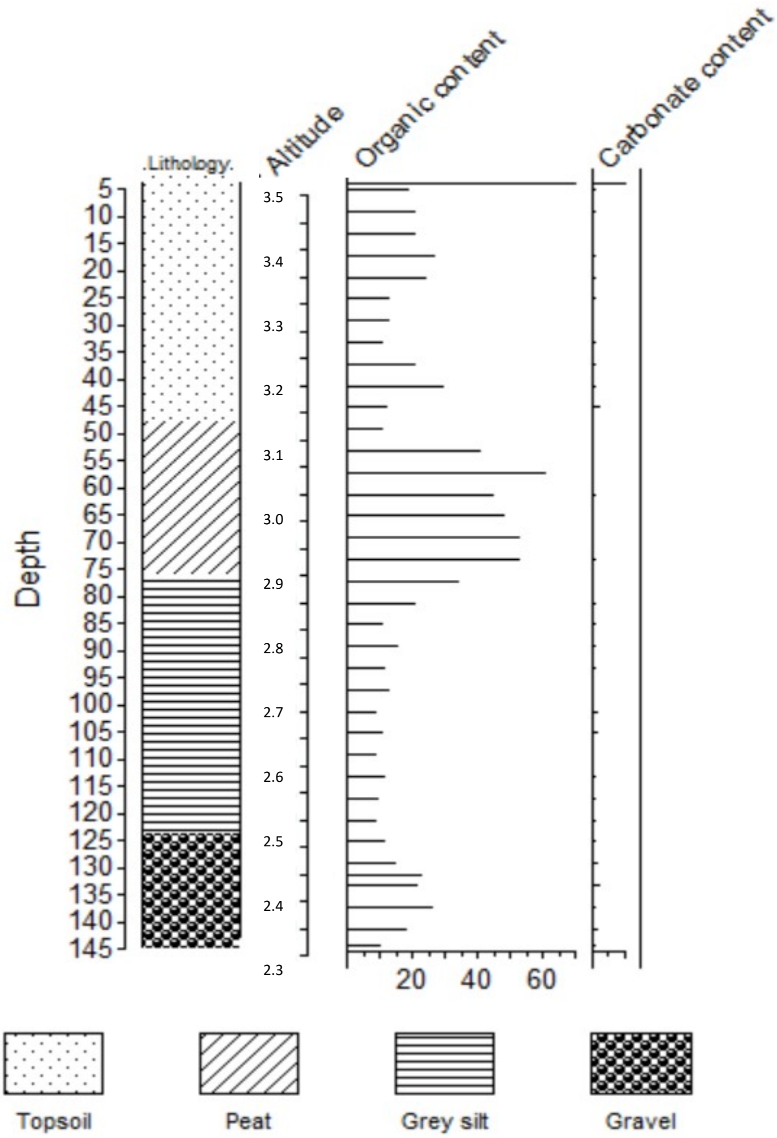


Figure 6.82: Graph showing organic carbon and carbonate content for the sample core at Gesail.

6.3.3 Particle Size

Subsamples for particle size were taken every 12cm, with samples taken every 2cm over the upper boundary from clay to peat in order to better define the changes occurring (Figure 6.83). The basal gravel was not sampled with the Russian-style corer, but the basal peat (124 to 145cm, 2.26 to 2.05m OD) contained a high percentage of sand. The sand content within the peat resembles the blown sand found in a machair type landscape often seen in Scotland (RSPB, 2014), where large grassland and peat are enriched with lime from sand from seashells at the coast (RSPB, 2014). Within the overlying grey silt (76 to 124cm, 2.74 to 2.26m OD) high levels of sand remained present, with increasing percentages of silt towards the surface. At the transition to peat at 76cm (2.74m OD) the percentage of silt declined and a peak in sand occurred, which in conjunction with higher organic content could indicate an inwash of sediment. Sand was present throughout the peat as well as silt. The overlying silt topsoil (0 to 48cm, 3.5 to 3.02m OD) contained predominantly silt with very little sand. The predominance of sand throughout the core indicates a high energy environment and implies that fluvial conditions dominated, which correlates with the ribboned paleochannels seen at the site.

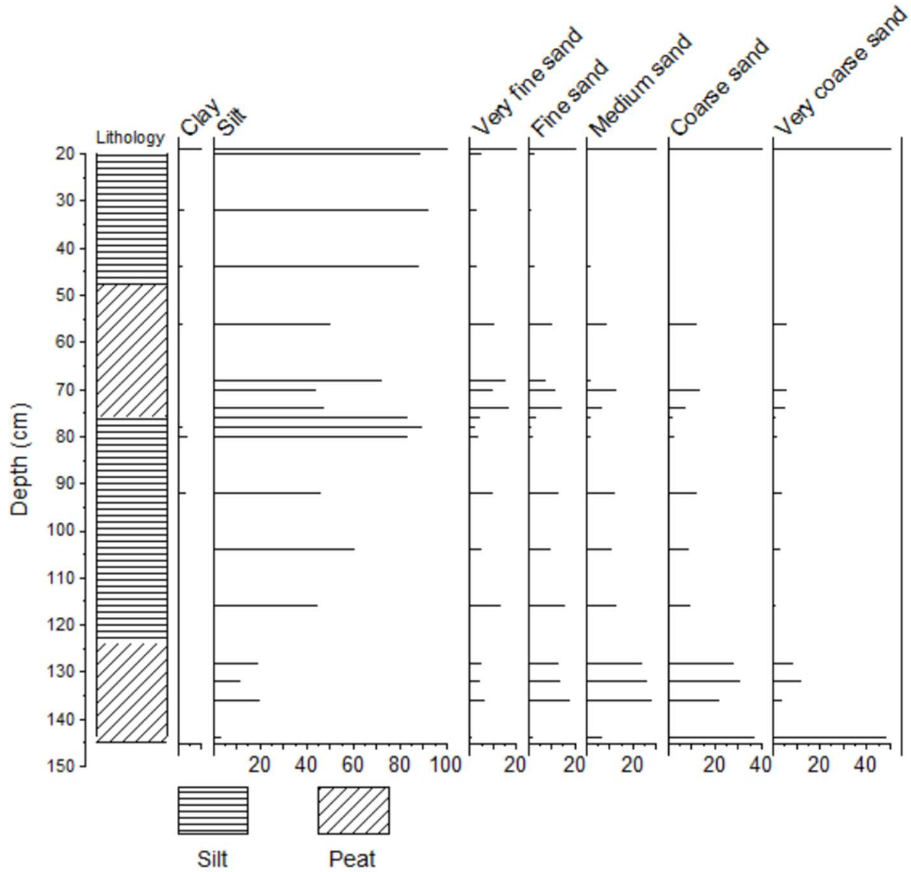


Figure 6.83: Graph showing particle size data for the sample core at Gesail.

6.3.4 Foraminifera analysis

Samples of 2-5cc of sediment were taken for foraminifera analysis at 4cm intervals throughout the core. No foraminifera were present, indicating no marine influence.

6.3.5 Diatom analysis

Samples for diatom analysis were taken at 12cm intervals throughout the core, and no diatoms were found above 70cm (Figure 6.84). Preservation was very poor with many broken valves, and several could not be identified to species level. The diatom assemblage at Gesail was largely homogeneous and predominantly terrestrial, dominated by the freshwater, epiphytic *Ctenophora pulchella* (Roe *et al.,* 2009; Williams and Round, 1986). *Eunotia sp.* was also present; *Eunotia praerupta* is epontic and stenohaline (Denys, 1991/2; Van dam *et al.,* 1994) and *Eunotia paludosa* is a terrestrial species which inhabits damp, mossy environments (Liu *et al.,* 2011; Van der vijver and Beyens, 1998). Other freshwater species were present in small amount such as *Tabbelaria flocculosa* (Knudson, 1954), *Stauroneis phonicentron* (Denys, 1994) and *Pinnularia* sp., particularly *P. lata* and *P. major* (Krammer, 2000; Noga *et al.,* 2014). Some broken fragments of single marine and brackish diatoms were recovered such as *Amphora* sp., *Diploneis* sp. and *Cosmioneis pusilla* (Denys, 1991/2), however these are likely allochthonous due to the damage to the valves and low frequencies.

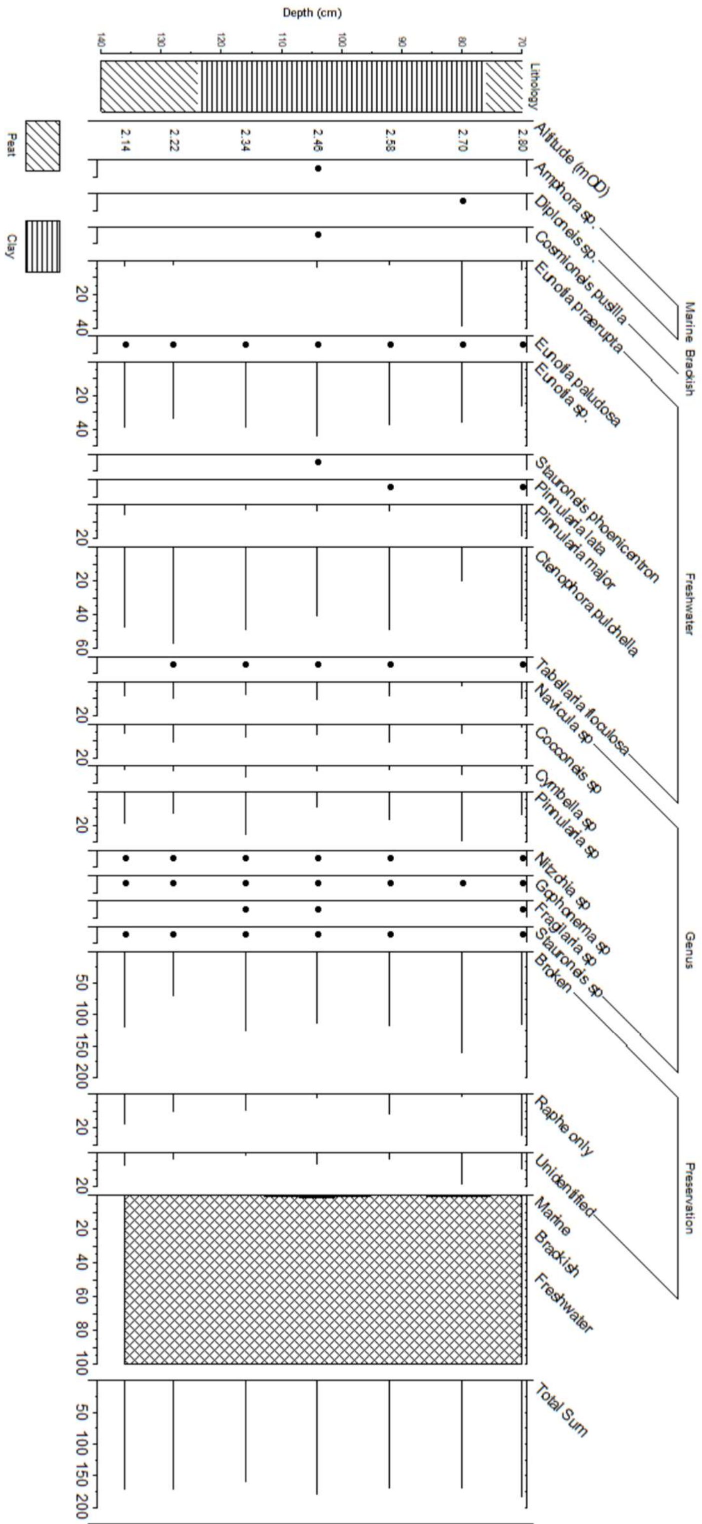


Figure 6.84: Diatom percentage diagram from Gesail showing stratigraphy and radiocarbon dates

6.4 Environmental change at Gesail

Towards the top of the valley at Gesail, the surface altitude rises to 4m OD, with poorly sorted, angular shale gravel with quartz inclusions outcropping at the surface. The stratigraphic sequence across Gesail is quite irregular, with pockets of organic silt, sand and gravel intercalated in grey clay. In many of the gouge cores a grey clay with iron banding was recovered, which was soft and elastic at the base and more compact towards the surface. The environment throughout the sample core was terrestrial with no foraminifera, dominated by freshwater and weakly brackish diatoms such as *Ctenophora pulchella*. There were three individual, broken valves of marine and brackish diatoms which were likely allochthonous. There were some layers of organic silty peat in the sample core, but overall, the organic content was below 20%. Sand was present throughout the sample core, indicating a high energy environment. This freshwater, high energy environment with poor diatom preservation correlates with the many ribboned palaeochannels across the top of the valley, indicating that fluvial conditions dominated the sequence. No evidence for marine influence has been found, although it is possible that erosion has occurred.

6.5 Conclusions

The gouge survey at Gesail revealed basal, glacial gravel overlain by clay with sand and organic horizons. A short sample core was taken, which contained no foraminifera and only freshwater diatoms, revealing the sediments were from high energy, fluvial environments. The landward extent of marine influence in the Dysynni valley was found at Perfeddnant and is unlikely to have exceeded past location 3 for Gesail, 7.76km from the coast, as it correlates with the most inland edge of the blue-grey silt. It is unlikely that marine inundation extended further inland. The gouge survey at Gesail established that sediments were freshwater at the base of Craig yr Aderyn, despite previous research (Pratt *et al.,* 1995) although care must be taken as some fluvial erosion may have taken place.

Section C: Conclusions of Research

Chapter 7: Discussion

This chapter discusses the results with respect to the original research aims of the thesis, section 1.2.

For reference, the original aims and objectives were:

1. To reconstruct the Holocene sea-level changes in the Dysynni Valley.
2. To recreate the environmental and vegetation changes in the Dysynni Valley in order to fully understand past changes in the coastal environment.
3. To examine the impacts that the changing coastal environment had on human settlement in the area.

7.1 Sea-level change in the Dysynni Valley

Reconstructing changing sea level is crucial in order to accurately interpret the dynamic coastal environment which impacted human populations in the past. Differential ice loading during the LGM resulted in isostatic uplift on the north coast of Scotland, and isostatic subsidence on the south coast of England (Shennan *et al.,* 2018). The valleys of northern Cardigan Bay, such as the Dysynni, are vital for sea-level research as they are unaffected by isostatic uplift or subsidence (Shennan *et al.,* 2012). Care must be taken, however, as although the current models indicate stability, it is possible the area may have been influenced by isostatic change in the past. GIA models (e.g., Peltier 1994, 2004; Peltier *et al.,* 2002) predicted a mid-Holocene high stand for the north coast of Wales, where past sea-level was higher than present. Recent GIA models such as Shennan *et al.* (2018) show no mid-Holocene high stand. More data are needed to confirm the presence or absence of a mid-Holocene high stand on the north of Wales.

Microfossil analyses, transfer functions and sea-level index points were applied to the sediments from the Dysynni Valley in order to establish sea-level change through the Holocene (Aim 1). Diatom and foraminifera analyses were undertaken at three sites in the Dysynni Valley: Penllyn at the coast near Tywyn, Perfeddnant 7km inland near Bryncrug, and Gesail, 9km inland at the base of Craig yr Aderyn. Transfer functions were established from regional training sets for diatom and foraminifera, as well as a multiproxy training set (Chapter 3.4.2). The diatom transfer function was the most precise, with good values of root mean square error of prediction (RMSEP) and coefficient of determination (r^2_{boot}), while the multiproxy training set had reasonable performance. In contrast, the foraminifera transfer function had poor precision, although pruning the dataset offered significant improvement. Errors were established following Hijma *et al.* (2015) and Smith *et al.* (2020) (Chapter 3.4).

At Penllyn (Chapter 4) a back-barrier saltmarsh was present in the basal peat, indicated by the diatoms *D. interrupta* and *C. pusila*, with some intertidal mudflats indicated by *D. didyma, N. peregrina, T. navicularis* and *S. turmida* (Vos and de Wolf, 1993). A monospecific *J. macrescens* assemblage suggested a high marsh environment on the very edge of marine influence (Scott and Medioli, 1978; Gehrels, 1999; Wright *et al.,* 2011; Horton and Edwards, 2006). At 500cm (-3.39m OD) the back barrier salt marsh gave way to an estuarine silt, with increasing salinity as *D. interrupta* and *C. pusilla* were replaced by *C. westii*, *T. navicularis* and *C. littoralis*. Foraminifera indicated an increase in marine conditions, with *T. inflata* and *M. fusca* appearing alongside *J. macrescens*. There was an increase in marine conditions from 480cm (-3.04m OD) indicated by low marsh foraminifera and intertidal, marine diatoms. A peak in salinity was indicated by an increase in *D. smithii* at 380cm (-2.04m OD). From 325 cm (-1.49m OD), marine conditions began to decline, with *J. macrescens* increasing alongside low marsh foraminifera. Between 276-223cm (-1.00 to -0.47m OD) diatom and foraminifera evidence indicated a return to a brackish saltmarsh environment frequently exposed to the air, with calm conditions as *D. interrupta* and *J. macrescens* became dominant. As the sedimentation rate overtook rising sea level, peat developed with evidence that brackish, high saltmarsh conditions persisted across the stratigraphic boundary. The peaty silt was overlain by the submerged forest peat dated to 6402-6297 cal. BP (D-AMS 028035). No diatoms were found, with low counts of monospecific *J. macrescens* and *M. fusca* up to 221cm (-0.45m OD), indicative of high marsh conditions.

The transfer functions were applied to the diatom and foraminifera data taken from the blue-grey estuarine silt. Following Watchman *et al.* (2013) the modern analogue technique was used in order to assess the reliability of the transfer functions. The diatom transfer function indicated a rise in sea level from -8.5 to -5.5m OD between 7787-7665 cal. BP and 6402-6297 cal. BP. Although the diatom transfer function was the most accurate model in terms of prediction statistics, it included only poor modern analogues throughout the core, making the data unusable. Local diatom communities are remarkably diverse, and the lack of a local training set is detrimental to the application of the diatom transfer function in the Dysynni Valley.

Similarly, the multiproxy transfer function produced poor analogues for samples with both diatom and foraminifera microfossils, due to a lack of a local multiproxy training set, therefore it could not be used. In contrast, the less precise foraminifera transfer function produced good modern analogues for the high marsh assemblages, with close and poor analogues for the calcareous, low marsh assemblages. This is due to the narrow tolerance ranges of high marsh foraminifera (Scott and Medioli, 1978; Gehrels, 1999; Wright *et al.,* 2011; Horton and Edwards, 2006), in contrast to the wide tolerance ranges of calcareous foraminifera, which are also known to dissolve upon burial (Gehrels *et al.,* 2001; Horton and Edwards, 2001, 2006). The foraminifera transfer function showed an overall rise in sea level from -6 to -1.5m OD from 7787-7665 to 6402-6297 cal. BP, which is much higher than the diatom transfer function.

In order to accurately assess the performance of a transfer function, results must also be compared to independent measurements (Watchman *et al.,* 2013). However, no independent observational record, such as tide gauge data, was available for the timeframe, so following Barlow *et al.* (2013) the transfer function was compared to the sediment lithology. The lithology and pollen evidence from this period record a transition from peat to blue-grey silt and saltmarsh, fen, and reed swamp.

The rate of sea-level rise declined upon the transition to peat, dated to 6402-6297 cal. BP, with the saltmarsh pollen diminishing. Alder developed, replaced with birch, suggesting a lowering of the water table and infilling of the estuary following a decrease in the rate of sea-level rise. Both the diatom and foraminifera transfer functions agree with the sediment lithology, making them equally likely to be accurate.

The transfer function findings were used to calculate the indicative meaning of two sea-level index points for Penllyn (Chapter 4.4.2). A third SLIP was calculated by establishing the indicative meaning from the microfossil evidence without the use of a transfer function. The diatom and foraminifera results indicated a high marsh environment for both dated horizons, which Hijma *et al.* (2015) placed between the highest astronomical tide (HAT) and mean tide level (MTL), using the equation (HAT+MTL)/2 (Chapter 4.4.2). The SLIPs from the foraminifera transfer function demonstrated a rise from -6.16m OD MHWST at 7787-7665 cal. BP to -2.88m OD MHWST at 6402-6297 cal. BP. In contrast, the diatom transfer function SLIPs plotted much lower, reflecting a rise from -8.42 OD MHWST at 7787-7665 cal. BP to -6.00m OD MHWST at 6402-6297 cal. BP. The indicative meaning calculated from the microfossil evidence revealed a rise from -7.82 m OD MHWST at 7787-7665 cal. BP to -4.91m OD MHWST at 6402-6297 cal. BP, similar to the diatom transfer function. Although the SLIP from the diatom transfer function is remarkably similar to the indicative meaning from the microfossil analyses, there were no modern analogues and therefore the data are not reliable. In contrast, the foraminifera transfer function had good modern analogues yet produced reconstructions which indicated higher sea level than all other SLIPs. Due to problems with both transfer functions, the SLIPs calculated from the microfossil evidence were used for Penllyn. The application of the transfer functions in this study has highlighted the need for local, multiproxy training sets, as well as the importance of detailed, multiproxy microfossil analyses alongside statistical analyses.

The site at Perfeddnant is located 7km from the coast, northeast of the alluvial fan of Bryncrug, with a glacial gravel buried by a freshwater peat dated to 9885-9547 cal. BP (D-AMS 031520) at 435cm (-2.61m OD). The peat was overlain by an estuarine silt, with microfossil evidence for a high-marsh saltmarsh that kept pace with sea-level rise. The silt was overlain by a woody, freshwater peat which was dated to 4782-4420 cal. BP (D-AMS 031519). The foraminifera transfer function had good modern analogues and showed a sea-level rise from -7.74 OD MHWST to -3.48m OD MHWST, however, these horizons did not produce reliable dates (Chapter 5.3.4), and therefore SLIPs could not be calculated. Diatom preservation was poor throughout the silt, with only five samples suitable for use with the transfer function and no modern analogues. The diatom samples showed a rise in sea level from -7.4 to -4.95m OD MHWST, which was much lower than the foraminifera transfer function, similar to Penllyn. Microfossil analyses at the dated horizons at Perfeddnant showed the site was freshwater at 9763-9551 cal. BP and again at 4567-4436 cal. BP. At the top of the valley, the core taken from Gesail showed layers of freshwater alluvium between +2.05 and +3.5m OD overlaying glacial gravel. There was no evidence of foraminifera and only freshwater diatoms, indicating that no marine influence was present.

The sea level evidence from the Dysynni Valley showed that sea level rose from -7.82m OD MHWST at 7787-7665 cal. BP up to -4.91m OD MHWST at 6402-6297 cal. BP, with sea level reaching to Perfeddnant. Marine inundation could not have extended further than 7.76km inland, where the leading edge of the estuarine blue-grey silt was found at SH 63503 06583. The core at Perfeddnant showed that the site was freshwater at 9763-9551 cal. BP and again at 4567-4436 cal. BP.

Shennan *et al.* (2018) compiled a sea-level database for Britain and Ireland with over 2100 datapoints from eighty-six regions. The database structure and errors followed Hijma *et al.* (2015), although for Wales no corrections were made for compaction or tidal range, with only tidal corrections made for the east coast of the UK using the work of Horton and Shennan (2009). The

SLIPs from Penllyn were plotted against mean tide level (MTL) to allow comparison with the database (Shennan *et al.,* 2018) (Figure 7.85, 7.86).

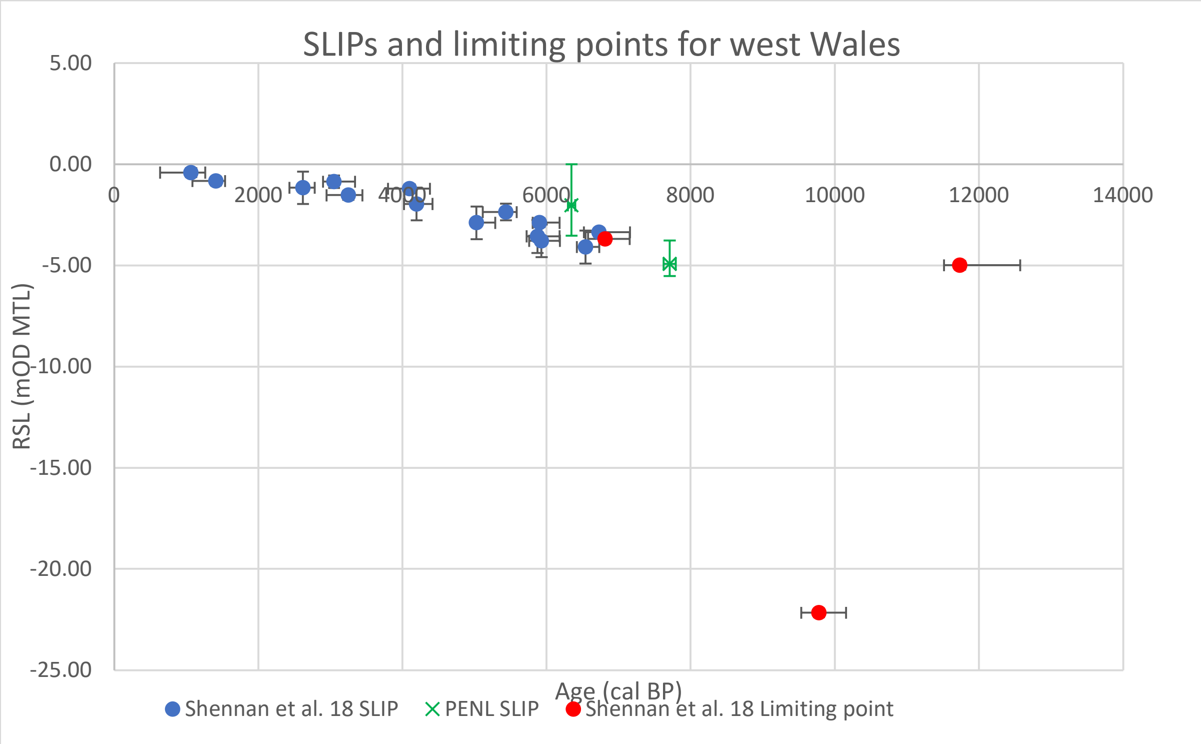


Figure 7.85: Sea-level curve for the west coast of Wales, with data from Shennan *et al.,* 2018 (blue: SLIP, red: limiting point) and Penllyn (green: SLIP)

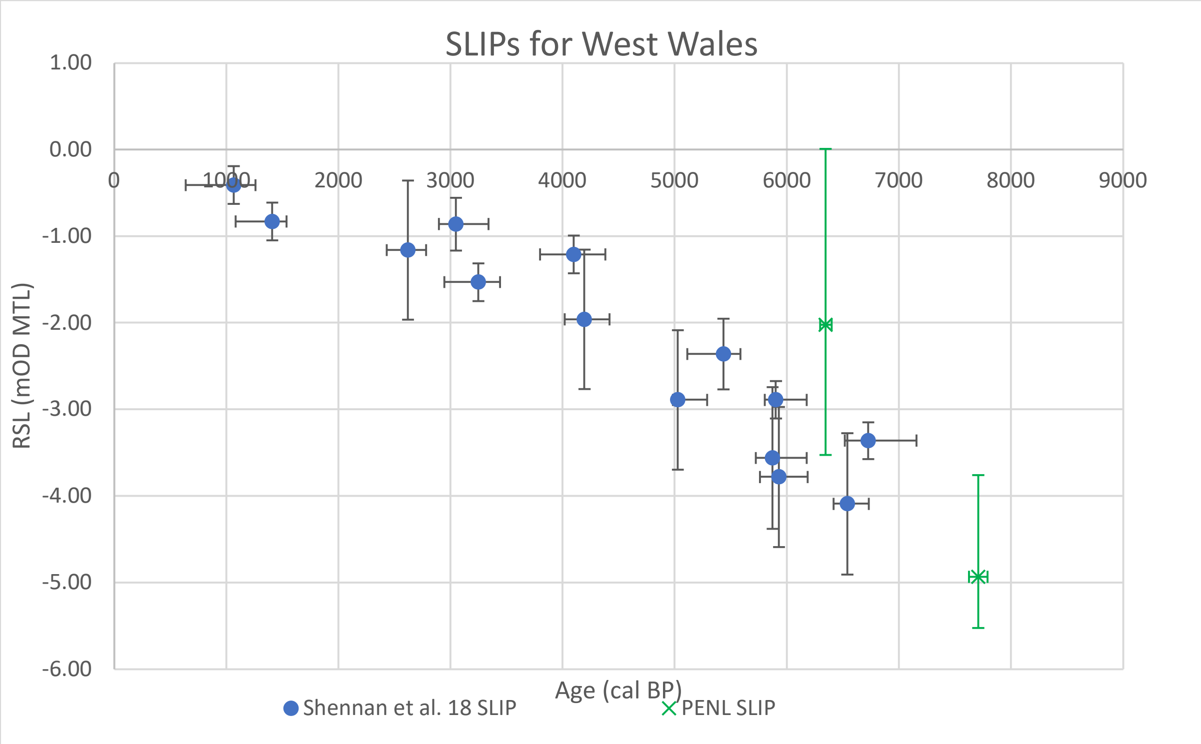


Figure 7.86: Sea-level curve for the west coast of Wales, with SLIP data from Shennan *et al.,* 2018 (blue: SLIP) and Penllyn (green: SLIP)

For west Wales, the data from Shennan *et al.,* (2018) show a rise in sea level from 6540 cal. BP at -4.09m OD MTL to 1066 cal. BP at -0.41m OD MTL (Figure 7.85 and 7.86). However, there are only limiting points between 9779 cal. BP (-22.16m OD MTL) to 6813 cal. BP (-3.69m OD MTL), which leaves the rapid sea-level rise through the early Holocene undefined. The SLIPs from Penllyn correlate well with the data for west Wales, showing a rise in sea level between -2.30m OD MTL (-7.82 m OD MHWST) at 7787-7665 cal. BP to -5.21m OD MTL (-4.91m OD MHWST) at 6402-6297 cal. BP. The Penllyn SLIPs have wider error bars compared to Shennan *et al.* (2018) due to the addition of errors such as compaction and tide range. The earlier SLIP at Penllyn fills an important gap in the data for west Wales, extending the sea-level curve to 7787-7665 cal. BP. The SLIP at 6402-6297 cal. BP plots slightly higher than the data for west Wales, but within the error ranges. It is possible the sea-level record for the Dysynni was affected by other, local factors, such as sedimentation rates, causing the SLIP at 6402-6297 cal. BP to plot higher. It may also be possible that there was a partial high stand in northwest Wales. More data is needed, however, particularly from the deep estuarine silts of the northern valleys such as the Mawddach in order to fully assess the sea-level record for the west coast of Wales.

Most of the sea-level data from the west coast come from three sites: Clarach (SN588 830) (Heyworth and Kidson, 1982; Heyworth *et al.,* 1985; Wilks, 1977, 1979; Taylor, 1973) as well as Borth and Ynyslas (SN 60655 92807) (Adam and Haynes, 1965; Godwin, 1943; Godwin and Newton, 1938; Godwin and Willis, 1961; Godwin and Willis, 1964; Haynes and Dobson, 1969; Prince, 1988; Taylor, 1973; Wilks, 1977, 1979). At Penllyn, glacial gravel was found across the site, with the upper surface between -4.3m OD and +1.5m OD in the gouge survey, with similar gravel found at Perfeddnant (-3.25 to -2.61m OD) as well as Gesail (-0.5m OD up to +4m OD). At Penllyn the gravel was overlain by a thin basal peat dated to 7787-7665 cal. BP (2.30m m OD) which produced a SLIP of -7.82 m OD MHWST. At Perfeddnant, in contrast, the basal peat was dated to 9763-9551 cal. BP and was freshwater. At Clarach, the glacial gravel was overlain by freshwater silt with late-glacial pollen dated to 12511-10780 cal. BP (BIRM 229) (Taylor, 1973; Heyworth *et al.,* 1985). There was no basal peat at Borth or Ynyslas, but at Clarach a forest peat on the foreshore was found to contain pollen of pine, alder, hazel, birch, and oak, dated to 7154 to 6563 cal. BP (NPL113) at -4.41m OD. At Penllyn, the basal peat showed a similar pollen spectrum with alder, birch, hazel, and oak dated to 7787-7665 cal. BP (-2.30m m OD). At Penllyn and Perfeddnant, the basal peat was overlain by blue-grey silt between -3.39m OD to -0.49m OD and -2.6m OD to -0.39m OD respectively. At Borth and Ynyslas, a similar blue-grey clay was found overlaying silty sand. Keeping (1878) discovered the marine shell *Scrobularia piperata* in the blue-grey clay, which was often described as “*Scrobularia*” or “OD” clay as it has been found between 0 and -1m OD across Wales (Prince, 1988; Godwin, 1948; Wilks, 1977; 1979). A similar estuarine sediment can also be seen in the Dysynni Valley between -1 and 0 m OD. At Clarach, the woody peat was overlain by a blue-grey silt at a similar elevation to the OD clay (-1.5 to -0.5m OD). However, it was a freshwater silt with coarse gravel, and therefore not the same as the *Scrobularia* clay.

At Penllyn, the blue-grey silt was overlain by a woody peat dated to 6402-6297 cal. BP at -0.49m OD, with alder woodland replaced by birch, transitioning to more open, disturbed habitats and reed swamp. At Perfeddnant, a similar woody peat was found which was dated to 4567-4436 cal. BP (D-AMS 031519, 219cm, -0.44m OD). The woody peat at Penllyn correlates to a submerged forest peat which was found at Borth and Ynyslas, as well as other sites across the west coast of Wales (Heyworth *et al.,* 1985; Prince, 1988; Wilks, 1977). Wilks (1977, 1979) dated the submerged forest to between 6179-5663 cal. BP (HAR1020, -0.25m OD) and 5586 to 5312 cal. BP (HAR 1019, +0.27m OD) with a pollen succession of alder, birch, and pine with a developing reed-swamp, similar to Penllyn. At Clarach, a woody peat was deposited with pollen evidence for alder woodland shifting to an open, disturbed habitat dominated by grass, sedges and other herbs such as ribwort plantain (Heyworth *et al.,* 1985).

At Penllyn, the woody peat extended to the surface with no evidence of further flooding episodes, implying no marine influence occurred after 6402-6297 cal. BP. In contrast, at Clarach the woody peat was overlain by silty peat which was dated to 2784-2379 cal. BP (HAR 1575) at 1.4m OD and 1261-927 cal. BP (HAR 1572) at 1.95m OD, with evidence of tidal flooding from freshwater and marine benthic diatoms (Heyworth *et al.,* 1985). There was also evidence for marine, brackish conditions at Ynyslas, dated to 4381-3894 cal. BP (HAR 1018, 1.15m OD). The saltmarsh peat at Ynyslas was overlain by peaty clay which contained *Juncus maritimus* (sea rush), high marsh foraminifera and freshwater diatoms as well as reed-swamp pollen, dated to 1538-1299 cal. BP (HAR 1016, 1.83m OD) (Adam and Haynes, 1965; Godwin, 1943; Wilks, 1977). A similar peat layer at Borth was dated to 3442-3040 cal. BP (HAR1017, 1.03m OD) with brackish diatoms and *Sphagnum* pollen indicating a regeneration of bog growth (Wilks 1977) (Figures 7.87,7.88).

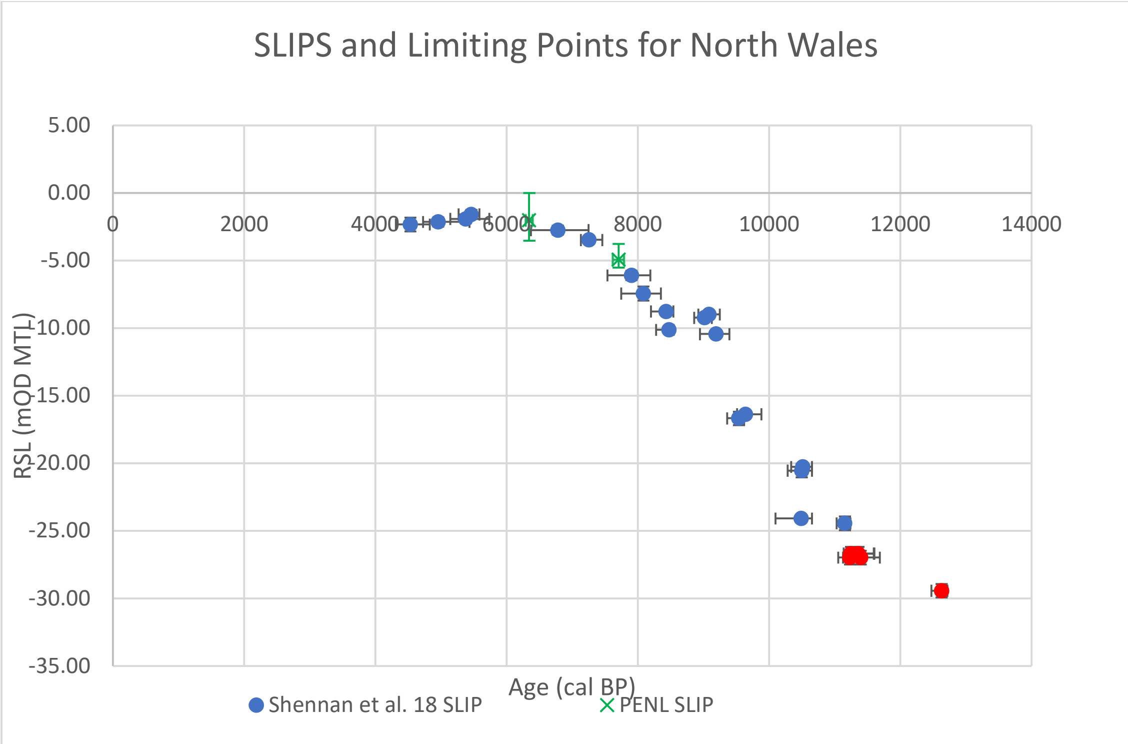


Figure 7.87: Sea-level curve for the north coast of Wales, with data from Shennan *et al.,* 2018 (blue: SLIP, red: limiting point) and Penllyn (green: SLIP)

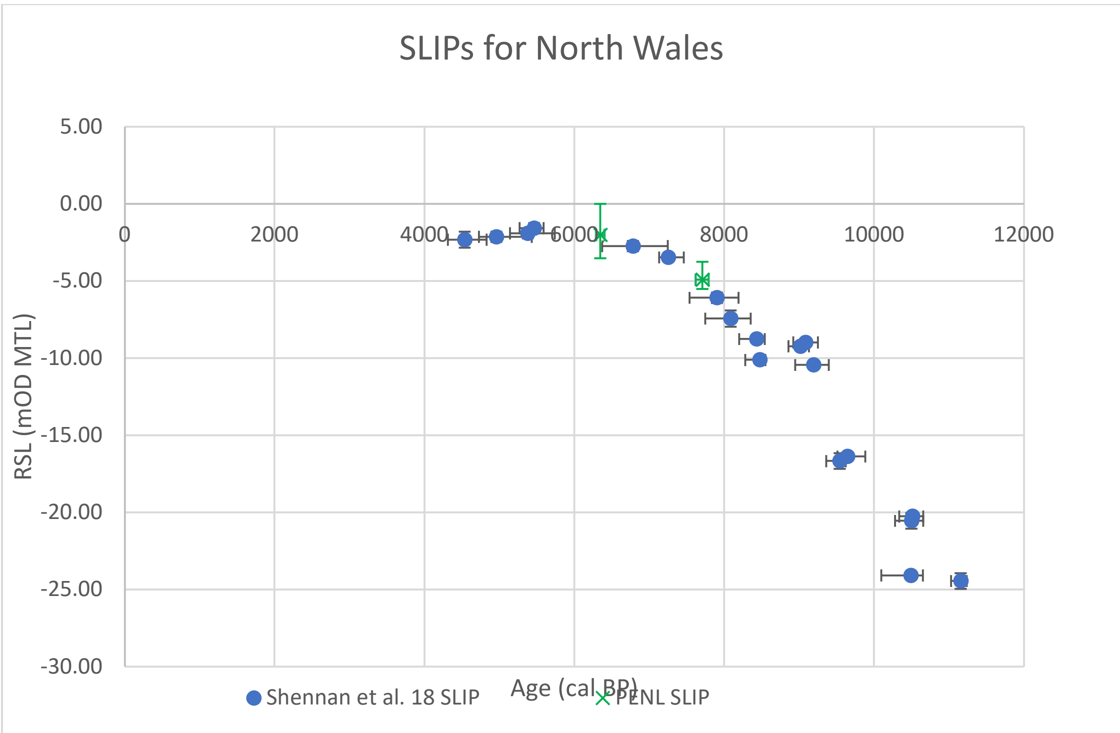


Figure 7.88: Sea-level curve for the north coast of Wales, with SLIP data from Shennan et al., 2018 (blue: SLIP) and Penllyn (green: SLIP), with the x axis truncated to 4000-12000 BP.

For north Wales, a rapid sea-level rise can be seen from 11155 cal. BP at -24.44m OD MTL to 7254 cal. BP at -3.47m OD MTL, followed by a more gradual rise to 4535 cal. BP at -2.33mOD MTL (Shennan *et al.,* 2018) (Figure 7.87). The sea-level curve between 7903 cal. BP (-6.09m OD MTL) and 5463 cal. BP (-1.58m OD MTL) is poorly defined, with a further gap between 11155 cal. BP (-24.44m OD MTL) and 9194 cal. BP (-10.44m OD MTL) (Figure 7.87 and 7.88). The SLIPs from Penllyn fit well into the north Wales data, filling two important gaps at 7787-7665 cal. BP (-2.30m OD MTL, -7.82 m OD MHWST) and at 6402-6297 cal. BP (-5.21m OD MTL, -4.91m OD MHWST). The data from Penllyn provide definition for the transition from the rapid sea-level rise of the early Holocene to the more gradual sea-level rise of the mid-Holocene, and confirm that a mid-Holocene high stand was unlikely for northwest Wales. More data is needed, however, as the SLIP from Penllyn at 6402-6297 cal. BP reaches almost to the surface if the error bars are taken into account, which would indicate the possibility of a small high stand. This also highlights the importance of taking into account error such as compaction which Shennan *et al.,* (2018) do not include.

Previous sea level research in north Wales has focussed on Anglesey and the Vale of Clwyd. In the Vale of Clwyd, Prince (1988) studied two cores at the sites of Plas Llywyd and Woodlands, obtaining many radiocarbon dates, although, the dates were rejected in future work (Bedlington, 1994; Roberts *et al.,* 2011; Shennan *et al.,* 2018). Bedlington (1994) undertook research at Hendre Fawr in the Clwyd plains, with 21 cores taken in a north-south transect, and stratigraphic analyses on two cores. At Hendre Fawr, Woodlands and Plas Llywyd, similar stratigraphy to Penllyn was found, with a basal, glacial gravel overlain by blue-grey clay (Bedlington, 1994; Prince, 1988). The microfossil evidence from the basal peat and blue-grey clay at Penllyn produced similar results to both Plas Llwyd and Woodlands, with foraminifera and diatom evidence for a back barrier saltmarsh transitioning to low marsh mudflats, and pollen evidence of mixed oak woodland, saltmarsh, reed swamp and fen. However, the date at Penllyn was much younger than Plas Llwyd and Woodlands (7787-7665 cal. BP, -2.30m OD), which reinforces the evidence that the Plas Llwyd and Woodland dates were too old (Bedlington, 1994; Prince, 1988).

At Hendre Fawr on the Clwyd Plain, Bedlington (1994) found a similar sequence to the nearby sites of Woodlands and Plas Llwyd (Prince, 1998). A basal peat at -2.48m OD was dated to 8192-7609 cal. BP, with freshwater diatoms giving way to upper and then lower marsh conditions in the overlying clay. A similar date was obtained from Llandudno station, where a basal peat, dated to 8542-8369 cal. BP (-5.21 to -4.16m OD) was overlain by estuarine clay and subsequently beach deposits (Harkness and Wilson, 1974). The dates from Hendre Fawr and Llandudno station are slightly older than the date of 7787-7665 cal. BP (-2.30m OD) found in the basal peat at Penllyn, although similar microfossil evidence for estuarine conditions were found. This may be due to local conditions in the Clwyd plain influencing early onset of marine environments.

At Penllyn, the blue-grey silt was overlain by woody peat at 214cm (-0.38m OD), however, at Hendre Fawr the blue-grey silt was overlain by fine silty clays with intercalated peats between -1.00m and +2.00m OD, dated to 7250–6320 cal. BP (Hv 17812) at +1.25m OD, with diatom evidence for a transition from upper marsh to fen environments (Bedlington, 1994). The date at Hendre Fawr correlated with the date of 7470–6910 cal. BP obtained from Morfa Penrhyn at -3.67m OD (Hv 17815) from blue grey clay with evidence for a marine brackish environment progressing to an upper saltmarsh and fen (Bedlington, 1994). At Penllyn, a similar date was obtained from the basal peat (7787-7665 cal. BP, -3.39m OD), with diatom and foraminifera evidence of a high saltmarsh progressing to a low marsh and estuarine, tidal mudflat environment. In contrast, at the site of Woodlands the blue-grey silt was overlain by a woody detrital peat between -12.95m to -13.07m OD, with pollen indicating an open habitat with sedges, grass and and goosefoot, which transitioned to an arboreal landscape with a saltmarsh pollen assemblage (Prince, 1988). Two dates were obtained (9400-9000 cal. BP, -12.95m OD SRR 2510; 9690-9330 cal. BP -13.07m OD SRR 2511), which are similar to the date of 9763-9551 cal. BP (-2.60m OD, D-AMS 031520) found at Perfeddnant. However, at Woodlands, these dates were considered too old (Prince, 1988).

At Penllyn, the core was dominated by peat from 6402-6297 cal. BP (D-AMS 028035) to the present, with pollen evidence for a mixed oak woodland and alder carr, correlating with the submerged forest found at Borth, Ynyslas and Clarach. However, at Hendre Fawr the peat was overlain by silty clay containing marine-brackish diatoms, dated to 5440–4520 cal. BP (+1.87m OD, Hv 17810) and 5730–4870 cal. BP (+1.7m OD Hv 17811) (Bedlington, 1994). A similar date of 5590–5310 cal. BP (-1.58m OD) was obtained by Tooley (1974, 1978) at Rhyl beach (SJ 0310 8261) from -1.58m OD from a regressive overlap of peat underlying silty clay. The Vale of Clwyd therefore shared some similarities with sites in the Dysynni Valley, with a basal peat overlain by estuarine blue-grey silt, however there was evidence for further episodes of marine inundation in the Vale of Clwyd, unlike at Penllyn and Perfeddnant.

For Anglesey, research has been undertaken at the Menai Straits (Roberts *et al.,* 2011) and the Malltraeth marshes (Prince, 1988; Bedlington, 1994). At Ty'n-y-pwll Farm on the Malltreath marshes, the basal gravel was overlain by a *Typha* layer between -9.35 to -9.50m OD, which produced two limiting points between 8979-8380 and 8378-8189 cal. BP (SRR 2507, SRR 2506) (Prince, 1988). At Tregarnedd Bach in the Malltreath marshes a basal peat was found between -4.42 and -3.15m OD, with inverted radiocarbon dates, however the pollen evidence for mixed-oak woodland indicated that the basal date of 8352-7839 at -4.11m OD was correct (Bedlington, 1994). At Penllyn, from 7787-7665 cal. BP (D-AMS 028034) a back-barrier saltmarsh, surrounded by mixed oak woodland, progressed to estuarine tidal flats then back to saltmarsh by 6402-6297 cal. BP (D-AMS 028035). A similar pattern was seen at the sites from the Malltreath marshes (Bont Farm, Tyn-y-pwll farm and Tregarnedd Bach), which indicated that the area was dominated by marsh and tidal flats throughout the Holocene, while marginal areas, such as Tregarnedd Bach, were more sensitive to sea-level fluctuations (Bedlington, 1994; Prince, 1998).

Roberts *et al.,* (2011) studied three cores from the Menai Straits (SH479 644), which ranged from the end of the LGM and the start of the Early Holocene (18000 to 10500 cal. BP). At Penllyn, the glacial gravel was found at 515cm (-3.39m OD), however in core CJSC1 of the Menai straits, basal diamict was found between -35.42 to -27.02m OD, overlain by angular gravels up to -23.62m OD (Roberts *et al.,* 2011). At Ty'n-y-pwll Farm on the Malltreath marshes, Anglesey the till surface was found at -11.94m OD overlain by sand and gravelly clay to -9.35m OD (Prince, 1988). Similarly, at Bont Farm, Malltreath marshes, the till surface was at -47.97m OD, overlain by gravel and then clay from -46.53 to -23.81m OD (Prince, 1988). A number of dates were obtained from Bont Farm; however, they were rejected as they were inverted (Prince, 1988; Bedlington, 1994).

In the Menai straits, deep late glacial and early Holocene sequences were found, for example in core CJSC1 light grey silts and sands overlaid the basal gravel to -4.12m OD, with marine molluscs of the genus *Mytilus*, *Cerastoderma, Tapes* and *Littorina*. Intercalated peats were found (1a and 1b), producing four limiting points between 11687-11246 and 11399-11099 cal. BP (SUERC2579, SUERC2578, SUERC2493, SUERC2496) as well as two SLIPs that indicated sea level rose from -24.44m OD to -24.08m OD between 11321-10825 cal. BP and 10652-10291 cal. BP (Roberts *et al.,* 2011).

Core CJSC2 in the Menai Straits was dominated by fine, light-grey and orange-brown silty clay containing sand and marine shells including *Mytilus*, but no gravel or diamict was recovered (Roberts *et al.,* 2011). Five intercalated organic horizons were found within the silty clay (2a-e), which produced a limiting point of 12950–12360 cal. BP (SUERC 2506) from a transgressive boundary at -29.4m OD, as well as two SLIPs tween 10654-10294 and 10655-10300 cal. BP when RSL was between -20.54 and -20.26m OD (Robert *et al.,* 2011). Two early Holocene SLIPs were also obtained of 9874-9543 cal. BP (SUERC2497, -12.97m OD) and 9622-9489 cal. BP (SUERC2498, -13.07m OD).

Core CJSC3 in the Menai Straits was taken from intertidal mudflats at the landward end of Bangor pier (SH5845 7335) and consisted of partially laminated silty clay with interbedded organics (peat beds 3a and 3b) overlain by silty sand and clay (Roberts *et al.,* 2011). Peat 3a was dated to between 9394-9031 and 8548-8407 cal. BP (-7.89 to -7.77mOD, SUERC2511 and SUERC2512), with foraminifera indicating upper marsh conditions. Peat 3b contained fresh and brackish aquatic macrofossils, dated to between 9250-9008 and 9129-8780 cal. BP (-6.68 to -6.63mOD, SUERC2509 and SUERC2510). At Perfeddnant, the basal peat was dated to 9885-9547 cal. BP (D-AMS 031520), with diatom evidence for freshwater conditions, similar to peat 3b in core CJSC3 of the Menai Straits.

The pattern of sea-level rise in north and west Wales were very similar, with rapid, early Holocene sea-level rise giving way to more gradual sea-level rise through the mid Holocene. In north Wales, there was a higher concentration of early Holocene and post LGM SLIPs, with the sea-level curve covering 11155 cal. BP to 4535 cal. BP (-24.44m OD MTL to -2.33mOD MTL). In contrast, there were more late Holocene dates for the west coast, where SLIPs spanned 6540 cal. BP to 1066 cal. BP (-4.09m OD MTL to -0.41m OD MTL). The data from Penllyn fill very important gaps in both data sets, helping to define the curve from the rapid early Holocene sea-level rise to the more gradual rise to the mid-Holocene. There has been some debate as to the presence of a mid-Holocene high stand in northwest Wales, with some models predicting that sea level would have been marginally higher than present (Peltier 1994; 2004; Shennan *et al.,* 2006; Brooks *et al.,* 2008). The data from Penllyn confirms the most recent models by Bradley *et al.* (2011) and Shennan *et al.* (2018), which predict no high stand in northwest Wales. Further data are needed, however, as the error range from the Penllyn SLIPs may indicate a small-scale high stand in north and west Wales.

There is more data for south Wales and Bristol Bay in comparison to the north and west coast, with very few gaps. The data from Shennan *et al.* (2018) demonstrate a rapid rise in sea level from 9330 cal. BP (-27.07m OD MTL) to 6804 cal. BP (-5.29m OD MTL) with a steadier rise to 685 BP (-0.08m OD MTL) (Figure 7.89 and 7.90). Although the overall pattern is similar to the northwest coast, the south Wales SLIPs plot much lower compared to the data from west and north Wales due to differences in isostatic uplift. Subsequently, the data from Penllyn plots consistently higher than the SLIPs from south Wales with a rise in sea level between -2.30m OD MTL (-7.82 m OD MHWST) at 7787-7665 cal. BP to -5.21m OD MTL (-4.91m OD MHWST) at 6402-6297 cal. BP.

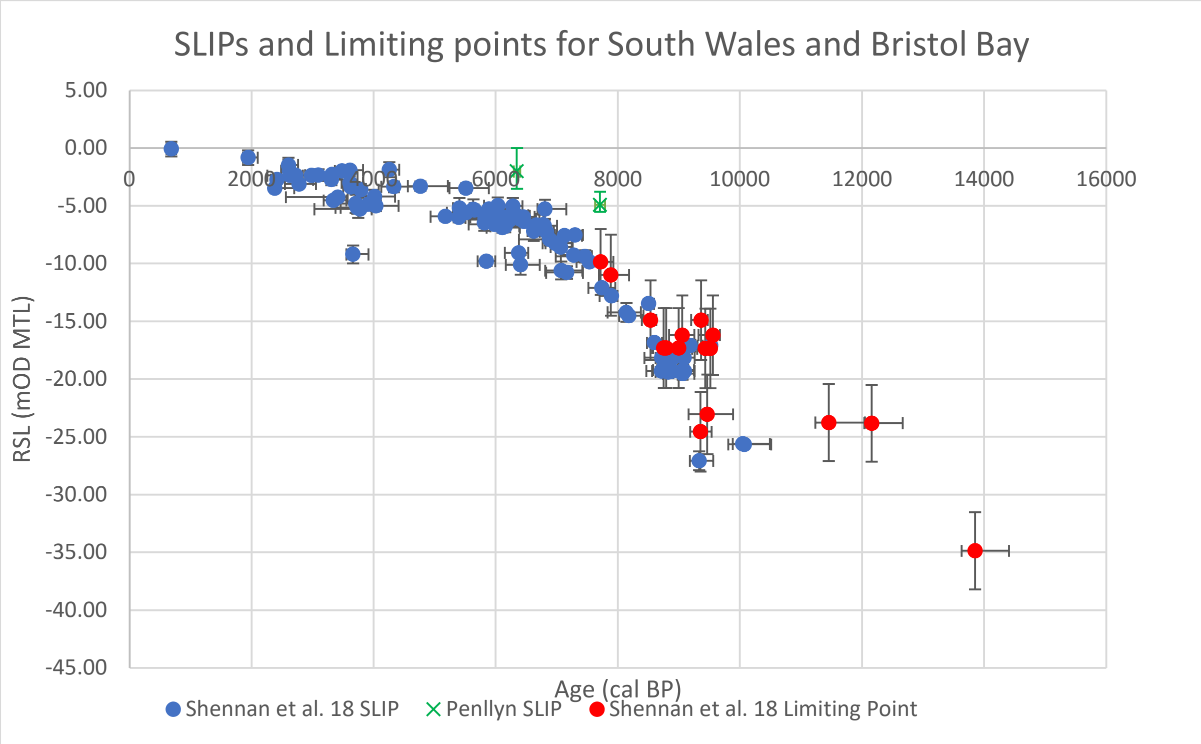


Figure 7.89: Sea-level curve for south Wales and Bristol Bay, with data from Shennan *et al.,* 2018 (blue: SLIP, red: limiting point) and Penllyn (green: SLIP)

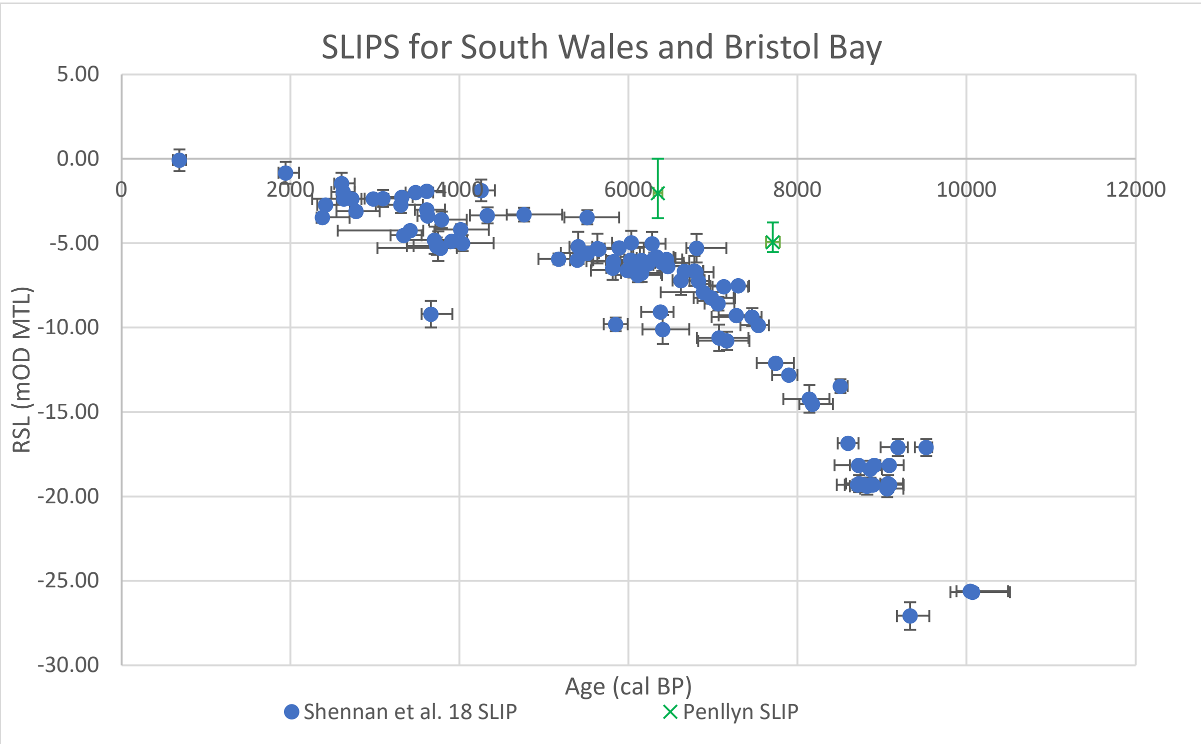


Figure 7.90: Sea-level curve for south Wales and Bristol Bay, with SLIP data from Shennan *et al.,* 2018 (blue: SLIP) and Penllyn (green: SLIP)

For Wales as a whole, the SLIPs and limiting points compiled by Shennan *et al.* (2018) exhibit an overall sea-level rise from 9330 cal. BP (-27.07m OD MTL) or 11155 cal. BP (-24.44m OD MTL) to 685 cal. BP (-0.08m OD MTL), with a wide range between the sea-level curves from north and south Wales (Figure 7.91 and 7.92). There was a wide range in the sea-level data, as the south coast SLIPs plot much lower than the north, due to differential isostatic recovery caused by patterns of ice loading during the LGM. The valleys of northern Cardigan Bay, such as the Dysynni and Mawddach, are vital for sea-level research as they are unaffected by isostatic uplift or subsidence (Shennan *et al.,* 2012). Furthermore, much of the data from south Wales comes from the Severn estuary. A large macrotidal estuary such as the Severn will produce very different results compared to small, over-deepened glacial valleys, which is why more research is needed from northern Cardigan Bay.

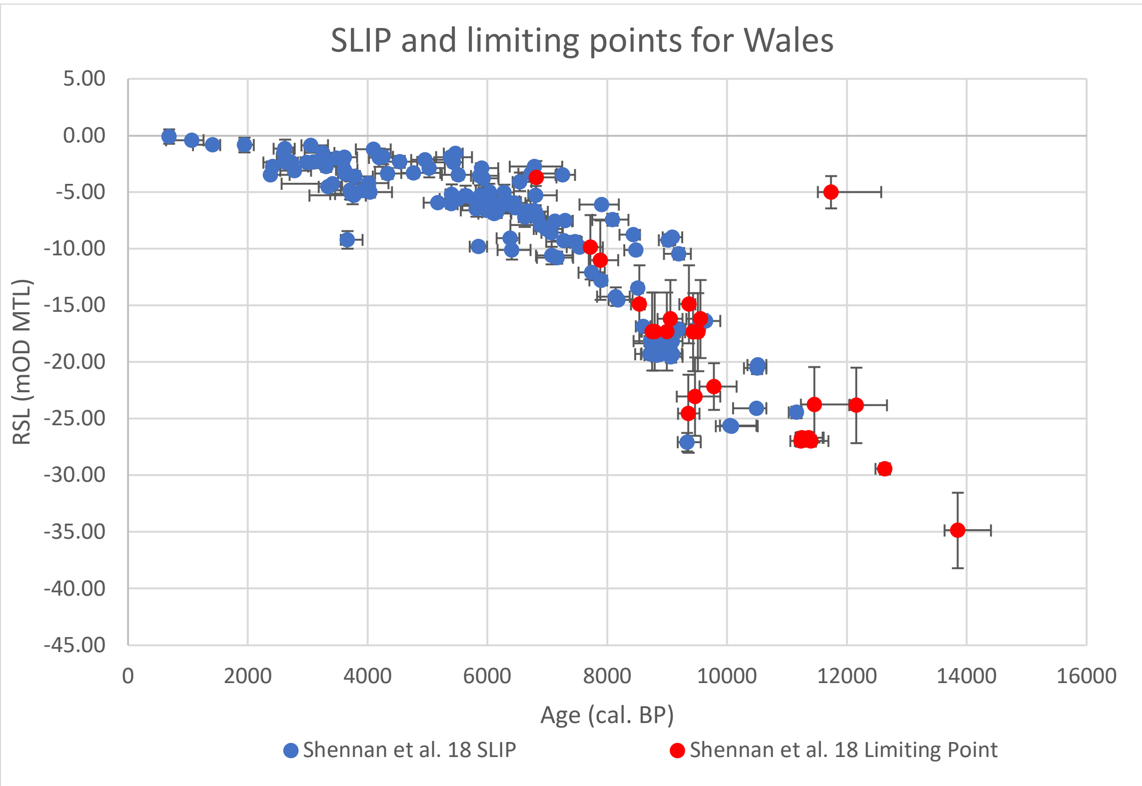


Figure 7.91: Sea-level curve for Wales, with data from Shennan *et al.,* 2018 (blue: SLIP, red: limiting point)

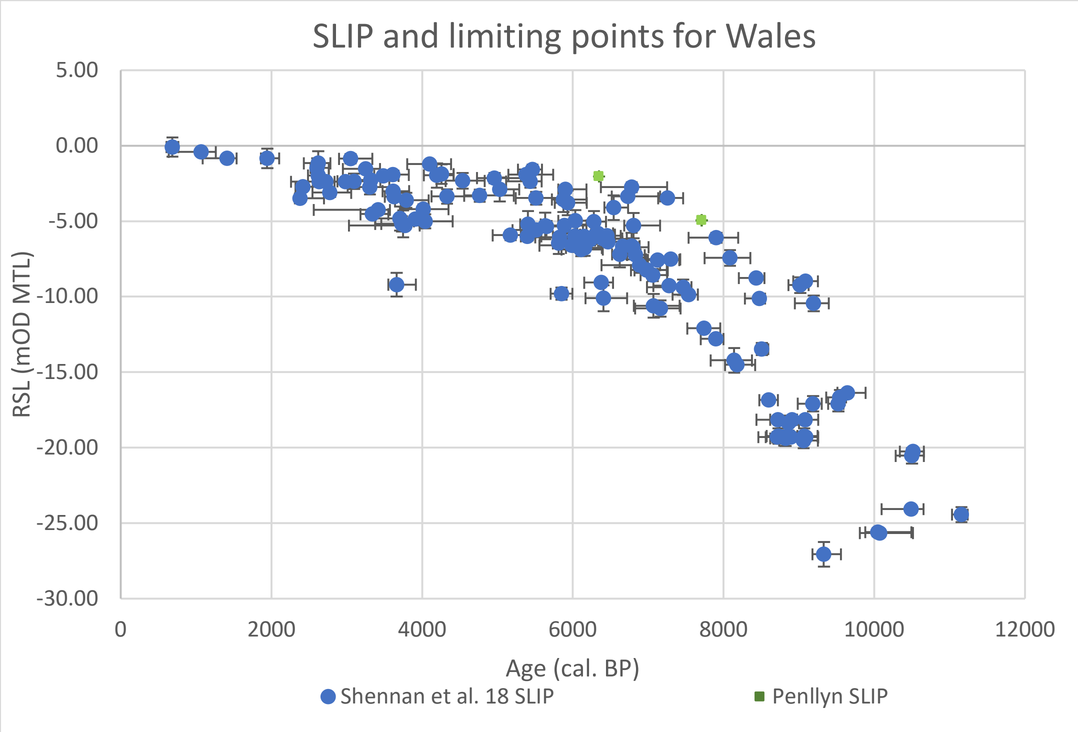


Figure 7.92: Sea-level curve for Wales, with SLIP data from Shennan *et al.,* 2018 (blue: SLIP) and Penllyn (green: SLIP)

7.2 Coastal and vegetation change in the Dysynni Valley

The following discussion reviews the coastal and vegetation changes reconstructed for the Dysynni valley (Aim 2). This is followed by a review of the environmental and vegetation reconstructions compared to the wider literature.

7.2.1 The Dysynni Valley

The Dysynni Valley is an over-deepened U-shaped glacial valley with steep sides and a broad floodplain. Much of the valley was reclaimed in the late eighteenth and early nineteenth century by the local gentry estates at Peniarth and Ynysmaengwyn (Frost, 2012).

Till, or boulder clay, was deposited during the last glacial period at the base of a glacier which extended the length of the Dysynni Valley (Pratt *et al.,* 1995). The town of Tywyn is situated on a large east-west facing drumlin of boulder clay up to 23m above sea level. Fieldwork at the site of

Penllyn for this study revealed the same gravel deposit, with the upper surface from +1.5m OD in the north to -4.3m OD in the south, representing the southern edge of the drumlin giving way to a basin of estuarine silt and peat (Chapter 4.2).

A prominent storm beach borders the coastline from Fairbourne in the north to Tywyn in the south. It is most well-developed at the mouth of the Dysynni, where it is 5m high and 120m wide (Pratt *et al.,* 1995). Fossil storm beaches can be found around the mouth of the Dysynni, consisting of diverse, well-rounded pebbles, 0.02-0.15m in diameter. Around Tywyn, a 500m stretch of degraded and vegetated sand dunes can be found behind the storm beach, best established at Broad Water (SH 5680 0280) where the dunes form arcuate ridges aligned parallel to the ancient storm beaches (Pratt *et al.,* 1995). The west of Tywyn lies on hummocky blown sand, and temporary excavations have exposed at least 1.25m of ochreous sand with thin fossil soil horizons. The gouge coring for this PhD project at Penllyn (Chapter 4.2) revealed blown sand from the west up to the dune and storm beach, which was covered by the modern caravan park.

Well-developed estuarine and marine alluvium has been found in the Mawddach, Dysynni and Dyfi estuaries. A borehole on Fairbourne spit (SH 6154 1486) revealed 48m of marine and estuarine sands, cobbles, and clays without encountering solid rock (Pratt *et al.,* 1995). Pratt *et al.* (1995) found predominantly sandy, estuarine alluvium underlying the well-drained land between Broadwater and Tywyn exposed in drains (SH 5876 0223) as well as an area of peat 7.5km^2 southeast of Broadwater intercalated with the estuarine alluvium. Fieldwork from the present study at Broadwater attempted to locate the estuarine alluvium and the intercalated peat, but coring attempts were not able to penetrate the surface sand. Two sites close to the river were also investigated (SH 024 593 and SH 039 593) which revealed layers of river alluvium, organic deposits, and gravel (Chapter 4.2). South of Tywyn at Penllyn farm (SN 5820 9930) Pratt *et al.* (1995) found 3.8m of blue-grey estuarine clay underlying 3-4m of peat with a basal layer of *in situ* and fallen tree stumps and roots, representing a submerged forest which is periodically exposed at low tide (Pratt *et al.,* 1995).

Bryncrug (SH 6050 0350) is a village situated on an alluvial fan of gravel originating from the mouth of the Tal-y-llyn valley (Snowdonia National Park Authority, 2014b). The largest area of peat in the Dysynni is situated upstream from Pont Dysynni behind the Bryncrug fan deposit. The wide floodplain at the mouth of the Dysynni narrows above Bryncrug to a steep-sided glacial valley with small alluvial fans. The river bottom is flat, with mounds of glacial till at different points in the valley, such as at Peniarth. The present study revealed blue-grey silt up to a depth of 7m (-5.1m OD) near Peniarth at SH 60915 04411, however the gravel could not be reached (Chapter 5.2). Blundell (1969) undertook a seismic survey in a transect across the valley near Peniarth and found bedrock at 152m, indicating that the valley was over deepened up to 30m. The soft, blue grey silt was found from the coast at Penllyn up the valley beyond the site of Perfeddnant to the southwestern end of the Gesail transect 7.76km inland (Chapter 6.2.1, location 3). The overlying woody peat was found up to location 14 at the site of Perfeddnant (Chapter 5.2.1), which was replaced by stiff, grey clay northeast up the valley to Gesail (Chapter 6). The sample core from Perfeddnant (SH 627 058) showed a basal gravel from 500 to 436cm (-3.25 to -2.61m OD), which consisted of poorly sorted angular shale up to 20mm long, with very coarse and coarse sand fractions, in a blue-black clay matrix, with some infrequent quartz inclusions. The gravel at Perfeddnant matched the gravel found at Penllyn, as well as the till described by the BGS surveys at Tywyn (BGS survey reports, 18/241).

At the foot of Craig yr Aderyn ribboned palaeochannels are visible in the valley bottom, as well as large boulders that are likely glacial in origin. Pratt *et al.* (1995) recorded evidence for estuarine alluvium in ditch cuttings at Craig yr Aderyn, which was overlain by river alluvium, but no coring or analysis was undertaken. Fieldwork for this study revealed stiff, grey silt with orange banding as well as lenses or organic material and gravel (Chapter 6.2). A sample core taken at Gesail near the base of Craig yr Aderyn, revealed a basal gravel of angular, poorly sorted shale with occasional quartz inclusions, which was found in gouge cores between -0.5m OD up to +4m OD. This gravel is similar to glacial sediments found at both Penllyn and Perfeddnant, and likely underlies the whole valley. The gravel was overlain by grey clay with iron banding, which was soft and elastic towards the base and dry at the surface, possibly due to drainage. Lenses of sand and gravel were present, for example at locations 9 and 11 as well as organic silt layers such as at location 10 (Chapter 6.2.1). The sample core at Gesail showed evidence for freshwater diatoms and no foraminifera, with a homogeneous assemblage dominated by the freshwater, epiphytic *Ctenophora pulchella* (Roe *et al.,* 2009; Williams and Round, 1986). Within the grey silt (124 to 76 cm, 2.26 to 2.74m OD) high levels of sand remained present, with increasing percentages of silt towards the surface. On the transition to peat at 76cm (2.74m OD) the percentage of silt declined and a peak in sand occurred, which in conjunction with higher organic content could indicate an inwash of sediment. Within the peat organic content rose to 40-60%, falling again to 10-20% up to the surface. The predominance of sand throughout the core indicates a high energy environment and implies that fluvial conditions dominated, which correlates with the ribboned paleochannels seen at the site. Therefore, the sediments deposited at the base of Craig yr Aderyn were freshwater, with no evidence of estuarine silt in contrast to the report by Pratt *et al.* (1995). It is possible, however, that erosion may have occurred, as the ribboned palaeochannels indicate the site would have been dynamic, with shifting river channels. The environment at the base of Craig yr Aderyn would have been a freshwater, dynamic, riverine environment with ribboned channels, however no estuarine or tidal action was present.

7.2.2 Environmental and vegetation change in the Dysynni valley and the wider region

The sample core from the coastal site at Penllyn (Chapter 4.3) revealed a basal peat between dated to 7787-7665 cal. BP (D-AMS 028034) with evidence for a back-barrier saltmarsh giving way to intertidal mudflats. Pollen indicated a diverse vegetation of saltmarsh, reed swamp and fen communities with nearby alder carr and diverse herb species as well as mixed-oak woodland in the surrounding valley. The core from Perfeddnant produced a thin, freshwater basal peat at Perfeddnant between 436 and 435cm (-2.61 to -2.60m OD), which was dated to 9763-9551 cal. BP

(D-AMS 031520). Freshwater conditions existed in the basal peat, continuing through the transition to blue-grey silt. From 418cm (-2.43m OD) a marginal marine saltmarsh developed, with both aerial exposure and flooding episodes from tidal action, river flooding or slope wash. From 354cm (-1.79m OD) up to 258cm (-0.83 m OD), high marsh foraminifera indicated a high saltmarsh, keeping pace with sea-level rise. The drumlin of Tywyn was likely an island surrounded by saltmarsh, reed-swamp, mudflats, and pockets of alder carr, with mixed-oak woodland on the upland slopes of the valley. Marsh extended up the valley to Perfeddnant, with the alluvial fan of Bryncrug and the moraine ridges found in the valley bottom, such as Peniarth, standing as islands in the woodland and marsh.

The pollen evidence for northwest Wales after the glacial period was characterised by the recolonisation of tree species, initially mixed woodland of oak and elm. The spread of alder and hazel followed, however the timing has been widely debated. The presence of alder at Penllyn at 7787-7665 cal. BP (D-AMS 028034) is unusual for the UK, however there is some evidence of very early alder in Wales compared to the rest of the UK. For example, alder was found at Moel y Gerddi by 8465 BP (Chambers and Price, 1985) and at at Cors Gyfelog by 8641-8400 cal. BP (Botterill, 1988). At Nant ffrancon the appearance of alder was dated to 9885- 9020 cal. BP (Hibburt and Swinstur, 1976), and at Melynllyn, the first evidence of alder was dated to 8518-7869 cal. BP (SRR639) (Watkins 1978). At Tregaron Bog, Dyfedd, alder was found at 8334-7622 cal. BP (Hibburt and Swinstur, 1976). The reason for an early spread of alder in Wales is unclear, with potential causes ranging from Mesolithic human activity, natural fires, beavers, or the creation of suitable habitats through floodplain development (Huntley and Birks, 1983; Smith, 1984; Chambers and Price, 1985; Brown, 1988; Chambers and Elliott, 1989; Bennett and Birks 1990; Edwards, 1990; Tallantire, 1993; Mighall and Chambers, 1995).

At Penllyn, the basal peat was overlain by blue-grey silt from 500cm (-3.24m OD) with pollen evidence for a local expansion of diverse saltmarsh and reed swamp communities as well as alder carr. Through the silt from 280cm (-1.04m OD), pollen analysis indicated that the valley remained largely occupied by mixed-oak woodland, while alder carr persisted with diverse woodland edge and disturbed habitat taxa. Reed-swamp herbs persisted, but saltmarsh and tall herb fen communities declined locally. There was possible evidence for disturbed habitats, and the presence of low amounts of microcharcoal, could be interpreted as evidence of anthropogenic impact.

Similar pollen evidence of disturbance and burning has been found between 8000 to 6000 BP across Wales as a whole, which has been explained as Mesolithic people intentionally manipulating the environment through the burning of woodland and coastal wetlands to promote grazing (Bell, 2007). For example, at Bryn-y-Castell there was circumstantial evidence for human impact before the elm decline including clearance episodes (Mighall and Chambers, 1995). There has been no archaeological evidence for agriculture before the elm decline in Wales (Groenman-van Waateringe, 1993; Edwards and Hirons, 1984; O'Connell, 1987), therefore the clearances were likely Mesolithic ecological disturbance such as burning (Mighall and Chambers, 1995). Similarly, at Porth Neigwl, Llŷn Peninsula, a Mesolithic and early Neolithic intertidal peat was found with evidence for disturbance from a decline in woodland cover and subsequent expansion of grass, nettles, docks, ribwort plantain and black nightshade (Smith *et al.*, 2016). Similar disturbances have been recorded in the pollen record at upland sites such as Moel y Gerddi (Chambers and Price, 1985; Chambers *et al.,* 1988a), the Berwyn mountains (Bostock, 1980), Waun-fignen-felen (Smith and Cloutman, 1988) as well as at Banc Wernwgan, Mynydd Du (Caseldine, 2013a), Clogwyngarreg (Grant 2012a) and Ffridd y Bwlch (Grant, 2012b; Caseldine *et al.,* 1990; Caseldine, 2017). Evidence for Mesolithic disturbance can be seen in other parts of Wales, for example at Goldcliff in the Severn Estuary, microcharcoal was interpreted as evidence of Mesolithic burning of local peat beds, which coincided with human occupation seen in the archaeological record (Bell, 2007; Dark and Allen, 2005).

The mid Holocene marked a period of dramatic coastal change across Wales, triggered by rapid sea-level rise. Submerged forests are found along the Welsh coastline, dated to between 9000-5500 BP (Bell *et al.*, 2007). The peaty silt at Penllyn was overlain by a woody forest peat dated to 6402-6297 cal. BP (D-AMS 028035). At Perfeddnant the woody peat overlying the blue-grey silt was dated to 4782-4420 cal. BP. The transition from silt to peat at Penllyn and Perfeddnant were both near 0m OD (-0.39m OD and -0.44m OD respectfully). A similar shift from minerogenic, estuarine silt to freshwater, terrestrial peat occurred at a similar altitude across the west coast of Wales. For example, at Borth, Wilks (1977, 1979) dated the transition from clay to peat at 6179-5663 cal. BP (HAR1020) and 5586 to 5312 cal. BP (HAR 1019). At Goldcliff, Severn estuary, the submerged forest was dated to 7682-7576 cal. BP (OxA-12359) (Bell *et al.*, 2007). The ubiquitous transition to freshwater peat has been correlated with the sedimentation rates overtaking the lower rate of sea-level rise across Wales, leading to coastal expansion (Wilks 1977, 1979). Similar estuarine silt near 0m OD has been seen in other parts of the UK as well, for example the carselands of the Forth Valley, where a mid-Holocene flood event was dated to 8,348–7,982 BP and occurred around −0.1 m OD (Jordan *et al.,* 2010; Smith *et al.,* 2010; Smith *et al.,* 2019).

Pollen evidence from Penllyn indicated alder encroached on the infilled marsh following the transition to woody peat between 220 to 140cm (-0.47 to 0.36m OD). Mixed oak woodland was present in the valley, with little evidence of saltmarsh, fen, or reed swamp. Some open or disturbed habitats were indicated by low amounts of ribwort plantain but even with the lack of marsh, the drumlin of Tywyn would likely have remained an island surrounded by woodland. At 140cm (0.36m OD) local alder carr was replaced by birch, implying conditions were drier. Regionally, mixed-oak woodland spread with many open, disturbed habitats as well as an expansion of reed-swamp and fen. At Penllyn, pollen analysis from the top of the core suggests that the landscape opened up, as most trees declined. There was an expansion of heather, likely in the surrounding uplands. Locally, the valley was dominated by grassland and reed-swamp, with some tall herb fen, and many open habitat or woodland edge taxa. An increase in anthropogenic impact may be interpreted from the combination of disturbance indicators such as ribwort plantain, a fall in tree pollen and an increase in microcharcoal.

A similar pattern can be seen in pollen sites across northwest Wales during the late Holocene, with marked anthropogenic as woodland clearance and agriculture transformed the Welsh landscape. The timing of these changes varied considerably, however. By the end of the Bronze Age at Bryn-y-Castell large scale woodland clearance produced an open landscape dominated by mire vegetation, which persisted to the present day (Mighall and Chambers, 1995). The permanent decline in woodland, affecting all tree species was dated to between 2992-2711 cal. BP (GrN-17583) and 2932-2750 cal. BP (GrN-17580) (Mighall and Chambers, 1995). At Tre'r Gof tree cover declined by 4406-3984 cal. BP (GrN-15521), which correlated with onset of the Bronze Age (Botterill, 1988). In the Aber valley on the north coast, however, woodland was widespread up to 3690-3467 cal. BP (BETA-246693) with woodland clearance indicated by a decline in tree pollen and expansion of grass and herbs by 897-673 cal. BP (BETA-246691) (Woodbridge *et al.*, 2014).

At Penllyn, the opening of the landscape is undated, and could correlate with later prehistory up to the post medieval period. It is possible that the pollen record of anthropogenic impact correlated with the post-medieval improvements of the valley in 1788 undertaken by Edward Corbert (Frost, 2012). It is likely the valley remained largely wooded and waterlogged until the post medieval period, with the drumlin of Tywyn, the alluvial fan at Bryncrug and the moraine ridges in the valley bottoms standing as islands in the marsh and woodland of the valley.

To conclude, sea level would have risen from -7.82m OD MHWST at 7787-7665 cal. BP up to -4.91m OD MHWST at 6402-6297 cal. BP. Marine influence was proved to extend inland to Perfeddnant, while the maximum horizontal inundation possible would be 7.76km inland, although it is unlikely the tidal influence reached to the base of Craig yr Aderyn. The valley would have resembled the Dyfi today with a wide, tidal estuary and saltmarsh extending up to Perfeddnant, with extensive reed swamp and tall herb fen as well as alder carr and mixed oak woodland. The alluvial fan at Bryncrug, the drumlin at Tywyn and other moraine ridges, such as Peniarth, would have been dry islands in a waterlogged environment of saltmarsh and estuarine channels for much of the Holocene.

Perfeddnant would have been freshwater from 4567-4436 cal. BP (D-AMS 031519, 219cm, -0.44m O). It is likely, however, that the valley would have remained waterlogged and heavily wooded until the drainage of 1788 by Edward Corbert (Frost, 2012).

7.3 Archaeology in the Dysynni Valley

The following discussion reviews the results of this study in relation to the archaeology of the Dysynni Valley and the surrounding areas in order to explore how past populations were impacted by coastal and environmental change (Aim 3). The section is split into four broad sections, firstly the period 11000-6000 BP where the Mesolithic response to sea-level change will be discussed as well as the lack of archaeological evidence. Secondly, the period 6000-3000 BP where changing prehistoric land use will be discussed in response to changing environmental conditions. Thirdly, the period 3000-1000 BP where the initial pollen evidence of human impact at Penllyn will be discussed. Finally, from 1000 BP onwards the evidence of large-scale farming in the pollen record will be discussed.

7.3.1 11000-6000 BP

There is currently little archaeological evidence for the Mesolithic on the west coast of Wales, with archaeological research focussed on the north and south coasts, with the closest sites to the Dysynni located on the Llŷn peninsula (Bell, 2007; Bell *et al.,* 2013). No Mesolithic finds have been recovered from the Dysynni Valley, but the nearby peat beds at Borth produced a Mesolithic flint pick, flint flakes, an antler tool and a hearth (Houlder, 1994; Sambrook and Williams, 1996; Smith *et al.,* 2002). Further to the south at Aberystwyth the flint production site at Penyranchor (Houlder, 1994) proves there was Mesolithic occupation of the west coast but further evidence is scarce. Arguments have been made that the lack of archaeological finds indicate that Mesolithic people were not inhabiting the west coast of Wales (Bell, 2007, p13). It is equally likely, however, that the lack of evidence is due to a dearth of archaeological excavation. It is vital to identify where Mesolithic sites are likely to be found in order to better elucidate future research on the prehistory of Wales. Understanding changing sea levels during the Mesolithic period is essential in locating potential archaeological sites by identifying the prehistoric coastline. Vegetation reconstructions from pollen analysis can reveal the presence of people within the landscape, where archaeological evidence for settlement may be less visible. When sea level and vegetation reconstructions are combined, the prehistoric landscape can be revealed as well as the response of past communities to the dynamic changing coastline in which they lived.

During the Early Holocene the Welsh coast was up to 35m lower than present (Smith *et al.,* 2002; Bell, 2007). As sea levels rose after deglaciation the deeply incised valleys of the Pleistocene would have become flooded estuaries, with diverse environments rich in resources (Bell, 2007). Around 9500 BP a third of Cardigan Bay would have been dry land, and by 7500 BP marine inundation extended significantly inland through river valleys across the Welsh coast (Bell, 2007; Robinson, 2019). Rising sea level would have had a dramatic impact on the size of Mesolithic territories, with the inundation of the Bristol channel, Severn estuary, Cardigan Bay, and Liverpool Bay leading to rivers shortening on average by 46% between 11500-6000 BP (Bell, 2007). It is likely, therefore, that many Mesolithic archaeological sites today would be submerged under Cardigan Bay, and more research is needed to identify potential areas of future research.

Bailey *et al.* (2020, p.214) identified important themes for future work on the archaeology of Britain's drowned Mesolithic landscapes. The impact of sea-level rise was highlighted, in particular the socio-economic, demographic, and psychological impact of coastal change which may be reflected in the patterns of exploitation, site distribution and settlement. Bailey *et al.* (2020) questioned whether changes in the late Upper Palaeolithic and Early Mesolithic archaeological record in Britain could be attributed to social impacts resulting from loss of territory on the retreating coastline. Such impacts would have depended on the density of populations as well as the magnitude and speed that land was lost through sea-level rise. If territory was lost rapidly on a large scale, it is possible that new economic and social arrangements were created. Conversely, a progressive, slow inundation of the coast could have facilitated population retreat without disrupting pre-existing social and economic composition. The loss of territory could have been offset by new environmental resources, such as productive fishing in the shallow seas formed as the continental shelf became inundated (Bailey *et al.,* 2020). In order to understand the relationship between prehistoric communities and the rapidly changing coastal environments, reconstructions of local coastal evolution and sea-level change is essential (Bell *et al.,* 2013). Sea level and vegetation reconstruction provides valuable information which can contextualise the archaeological record of the Mesolithic as well as identify areas for future work.

During the Mesolithic period estuaries and river valleys were important routes for transportation, with evidence of seasonal movement by Mesolithic people along the river valleys (Bell, 2007). Bell (2007) identified a pattern of Mesolithic occupation on many offshore islands, which would have been rich in resources, and may have had a more social significance. Clarke (1972) proposed a seasonal mobility model which suggested that Mesolithic people moved through a predictable seasonal round between small upland summer camps and sheltered lowland coastal sites (Preston and Kador 2018). Bell (2007) offered a similar model of seasonal movement in the Mesolithic, where communities occupied the coastal lowlands in winter, moving inland through the river valleys in summer with an annual range of perhaps 30 to 50km inland (Bell, 2007). The valleys of the west coast of Wales, including the Dysynni Valley, may well have formed similar routeways for Mesolithic people, transporting communities between upland summer camps to coastal winter settlements which would now be underwater. Future archaeological investigations will need to focus on the prehistoric Mesolithic coastline, much of which is now submerged under Cardigan Bay in order to identify the coastal Mesolithic sites. Environmental reconstruction will be essential in studying Mesolithic archaeology, with both offshore and onshore pollen cores needed as well as samples from archaeological sites in order to identify patterns of coastal and sea-level change as well as revealing potential anthropogenic impacts on the landscape.

At Penllyn, at 7787-7665 cal. BP a high marsh existed with mixed oak woodland, local alder carr and open, disturbed woodland edge herb taxa. There was extensive tall herb fen, saltmarsh, and reed swamp which would have produced a dynamic, complex mosaic of environments that would have been attractive to grazing animals and people. The diverse, woodland edge vegetation reconstructed at Penllyn during the Mesolithic period were similar to environments reconstructed from Goldcliff in the Severn estuary, with a mixed oak woodland, hazel scrub, bracken understory, saltmarsh and fen herb taxa (Bell 2000, 214) which produced extensive evidence for Mesolithic occupation. It is likely that the Dysynni Valley would have been similarly attractive to Mesolithic communities, which is in sharp contrast to the lack of Mesolithic archaeology, reinforcing the need for archaeological research into the Mesolithic of the west coast of Wales, particularly in the drowned landscapes of Cardigan Bay.

Pollen analysis from Goldcliff in the Severn estuary was able to identify Mesolithic land use, with microcharcoal evidence for fire interpreted as the local burning of reed beds (Bell, 2007). The evidence of burning at Goldcliff coincided with human occupation in the archaeological record (Dark and Allen, 2005). Similar pollen evidence of disturbance and burning has been found between 8000 to 6000 BP across Wales as a whole, which has been interpreted as Mesolithic people intentionally manipulating the environment through the burning of woodland and coastal wetlands to promote grazing (Bell, 2007). A similar pattern of burning can be seen at Penllyn, with an increase in microcharcoal at the transition from peat to blue-grey silt at -2.84m OD. The presence of disturbance indicators at Penllyn such as nettles, ribwort plantain, mugwort and docks alongside the increase in microcharcoal could be interpreted as an increase in human activity. Care must be taken, however, as the change from peat to silt at Penllyn represents a change in catchment area, as the estuarine silt would contain pollen from the whole river catchment while the peat reflects a more local pollen rain. It is possible the increase in microcharcoal in the silt represents natural fires in the whole river catchment. Furthermore, the lack of archaeological evidence for occupation makes it unclear if the pollen record represents anthropogenic or natural fires.

The evidence from the current study indicated that at Penllyn sea level rose from -7.82 m OD MHWST at 7787-7665 cal. BP to -4.91m OD MHWST at 6402-6297 cal. BP, with horizontal marine inundation extending 7.76km inland. The storm beach at Penllyn was likely breached and moved inland through the Holocene as the valley was inundated (Leng and Pratt, 1985), with diatom and foraminifera evidence indicating developing marine conditions as saltmarsh was replaced by low marsh, intertidal mudflats. A similar pattern has been seen across the UK, with rapidly rising sea levels through the Mesolithic period triggering abrupt periods of shoreline retreat as barriers of sand and gravel were inundated or breached (Bell, 2007). During the Mesolithic the Dysynni Valley would likely have resembled the Dyfi and Mawddach today with a wide, tidal estuary fringed by mudflats, marsh, and reed swamps. It is possible that Sarn y Bwch formed the edge of a large valley extending from the Dysynni Valley into Cardigan Bay, similar to the valley along Sarn Badric from Tremadoc Bay interpreted by Robinson (2013). The rapid shoreline retreat in the Dysynni likely forced Mesolithic communities to abandon coastal settlements, now submerged, in favour of sites further inland, such as the dry hills of Tywyn and Bryncrug which would have overlooked the marsh and intertidal mudflats which represented valuable resources.

7.3.2 6000-3000 BP

After the rate of sea-level rise at Penllyn declined from 6402-6297 cal. BP the landscape of the Dysynni Valley gradually transformed from a diverse wetland habitat with mudflats and saltmarsh, to a steadily drier catchment with birch, and mixed oak woodland. Pollen evidence showed that with the onset of peat, alder carr became dominant locally, with continued mixed oak woodland in the wider catchment. Although there was little evidence for fen, saltmarsh and reed swamp, with only a few disturbed areas indicated by ribwort plantain, alder dominated the profile and obscured an accurate picture of the environment.

Archaeological evidence for permanent settlement with durable domestic architecture in the Neolithic is relatively rare across the UK, although important discoveries of buildings have been made in northwest Wales (Sheridan, 2013). Ritual sites, such as chambered tombs, provide more numerous and tangible evidence of Early Neolithic activity in north Cardigan Bay. North of the Dysynni at Ardudwy, a group of six chambered tombs extend from the mouth of the Afon Arto to the north of Barmouth Bay (Bowen and Gresham, 1967; Cummings and Whittle; 2003, Robinson, 2016; 2019). Evidence for the Neolithic in the Dysynni Valley is limited to two stone axes found at Llanegrin (Frost, 2012). In contrast, evidence for Early Bronze Age land use in the Dysynni uplands is widespread, with many cairns and standing stones, likely built and used between 4200-3700 cal. BP (Smith, 2000). The range of upland cairns in the Dysynni reflects widespread anthropogenic impact in the uplands from the fifth millennium BP continuing throughout the fourth millennium BP (Johnston, 2021). The third millennium BP was marked by a change in land use across Wales, with evidence for settlement moving away from the uplands towards the more sheltered lowlands in response to a period of climatic deterioration (Burgess, 1985; Tipping, 2002; Tipping *et al.,* 2008; Davies and Lynch, 2000; Turney *et al.,* 2016). Turney *et al.* (2016) found dendrochronological evidence for extreme wet conditions in southwest Britain from 3100 BP, with Bronze age activity in the uplands ceasing by 2900 BP. Brown (2008) found a sharp increase in alluvial deposits and an increase in the peat water table across Britain from 2750-2600 to 2350 BP, with floods likely on the river floodplains (Waddington, 2013). Evidence has been found for a change in land use across upland areas of Britain in the third millennium BP, although the response was variable (Tipping, 2002; Tipping *et al.,* 2008; Dark, 2006). Tipping (2002, p17-20) reviewed pollen evidence for land use from pollen sites in northern Britain and found no evidence for landscape abandonment, instead highlighting a change in land-use towards more resilient farming practices such as pastoralism. Poorly defined chronologies and limited climate studies for northwest Wales makes an accurate assessment of climate driven landscape change difficult (Waddington, 2013). This is particularly true in the Dysynni Valley, where cropmark evidence for settlement at Bryncrug remains undated, and it is unclear if the site represents later prehistoric or early medieval settlement (Frost, 2012; Cook, 2019). It is interesting, however, that the settlement from both cropmarks as well as the early medieval church community at Tywyn and Croes Fan were focussed on the higher, dry hills of gravel such as the alluvial fan at Bryncrug and the drumlin of Tywyn. The focus on high ground implies that the valley bottom remained very wet until well into the medieval and post medieval periods, which would agree with the pollen record from Penllyn.

There was some evidence for climatic deterioration from the pollen profile at Penllyn. From 1m OD birch began to decline, replaced briefly by willow, which may indicate a period of wetter conditions, although the change is undated, it may correlate with a number of periods of climatic deterioration, such as recorded during the Bronze Age (e.g., Turney *et al.,* 2016), but this is unclear. The peak in willow coincides with the appearance of herbs such as ribwort plantain, daisy, buttercup, nettles, and dandelion, which may indicate a period of disturbance. Care must be taken, however, as these herb species may also occur in natural clearances (Behre, 2007; Simmons and Innes, 1987).

There was some evidence of human impact at Penllyn after 6402-6297 cal. BP from the presence of microcharcoal, and pollen evidence for open, disturbed habitats. Caseldine (in Smith, 2004) undertook a macrofossil assessment of the peats at Tywyn, and found evidence of charcoal fragments in the peat, which indicated woodland giving way to reed swamp, with local alder indicated by samples of identified alder wood. The presence of charcoal was interpreted as either natural fires nearby, or possible anthropogenic manipulation of the area (Smith 2004). While there was widespread upland land use from the early Bronze Age, it was not until the third millennium BP at the earliest that settlement shifted to the valley bottom, with a drier, although still heavily wooded landscape indicated by the pollen record.

7.3.3 3000-1000 BP

In the Dysynni Valley, it is unclear if settlement in the valley bottom developed in the medieval period, or if there was continuous occupation of the valley lowlands since the prehistoric period. At Penllyn, there was an increase in human activity at 0.36m OD indicated by a small increase in microcharcoal, which may indicate an increase in local burning, although it may also represent natural fires. At 0.36m OD the proportion of tree pollen fluctuated but remained high, indicating large scale clearances had not yet begun. No cereal pollen was found at Penllyn, however that does not prove the absence of cereal agriculture as cereals are self-pollinating, meaning that the pollen does not travel far (Bakels, 2000; Faegri *et al.,* 1964). For example, Bakels (2000) studied a peat core located next to an agricultural field with plough horizons which failed to find cereal pollen. The short-term human disturbance indicated in the pollen record at 0.36m OD may correlate to the undated settlement represented by the cropmarks at Bryncrug (Frost, 2012; Cook, 2019), but it is unclear if this represents continuous settlement from the Iron Age through to the medieval period, or if the pollen record represents the development of the early church community following the establishment of Tywyn (Smith *et al.,* 2001).

7.3.4 1000 BP and after.

It is unclear if large-scale farming in the valley bottom of the Dysynni began in prehistory, alongside evidence for upland grazing, or if the base of the valley was first farmed during post-medieval land improvements. Late eighteenth-century agricultural drainage saw large areas of 'turbay', or peat cutting, being ploughed for hay (Frost, 2012). The first large scale clearance and farming evidence from the pollen at Penllyn occurred between 1.27 and 1.76m OD. Pollen evidence indicated the opening of the landscape, with tree pollen declining, an expansion of grasslands and reed swamp, with an increase in heather. The presence of meadowsweet and bedstraw indicated the presence of tall herb fen, with many open habitat and woodland edge taxa such as ribwort plantain, daisy, mugwort and chamomile, as well as some buttercup, nettle, and pincushion flowers. A corresponding rise in microcharcoal indicates an increase in local burning, which combined with disturbance indicators and the fall in tree percentages indicated widespread clearance and farming in the valley. This was the first evidence of largescale human impact from Penllyn, and although undated it may well represent the medieval and post medieval improvements of the valley. Further pollen work and dating of the archaeological record will be needed in the Dysynni in order to fully understand the development of human occupation in the valley bottom.

7.4 Conclusions

The SLIPs from the Dysynni Valley showed that sea level rose from -7.82m OD MHWST at 7787-7665 cal. BP up to -4.91m OD MHWST at 6402-6297 cal. BP. Sea level reached no further than the site of Gesail, 7.76km inland, where the leading edge of the estuarine blue-grey silt was found at SH 63503 06583. The core at Perfeddnant showed that the site was freshwater at 9763-9551 cal. BP and again at 4567-4436 cal. BP. The SLIPs from Penllyn agree well with the sea-level curve produced by Shennan *et al.* (2018). The earlier SLIP at Penllyn fills an important gap in the data for west Wales, extending the sea-level curve to 7787-7665 cal. BP. The SLIP at 6402-6297 cal. BP plots slightly higher than the data for west Wales, but within the error ranges. The SLIPs from Penllyn also fill two important gaps in the north Wales data at 7787-7665 cal. BP (-2.30m OD MTL, -7.82 m OD MHWST) and at 6402-6297 cal. BP (-5.21m OD MTL, -4.91m OD MHWST). The data from Penllyn provide definition for the transition from the rapid sea-level rise of the early Holocene to the more gradual sea-level rise of the mid-Holocene, with little evidence of a mid-Holocene high stand. More data is needed, however, as the error range for the SLIP at 6402-6297 cal. BP extends to the surface, indicating a small high stand may have been possible.

The glacial gravel in the Dysynni Valley forms hills and ridges, such as the drumlin of Tywyn and moraine ridges in the valley bottom at Peniarth. A ubiquitous, estuarine silt was found throughout the Dysynni, with the upper level between -1 and 0m OD. At Penllyn a back-barrier saltmarsh existed from 7787-7665 cal. BP which gave way to intertidal mudflats with a high marsh saltmarsh extending up the valley 7km inland. Saltmarsh, fen, reed swamp and alder carr persisted in the valley, with Tywyn, Bryncrug and other ridges emerging as islands in a wide estuary with intertidal mudflats, saltmarsh, fen, reed swamp and alder carr along with mixed oak woodland for much of the Holocene.

Evidence for prehistoric archaeology is present in the Dysynni from the Neolithic period, with widespread Bronze Age land use in the uplands represented by cairns and standing stones. The initial evidence for settlement in the valley bottom is represented by the cropmarks such as at Bryncrug, which may reflect later prehistoric or early medieval settlement. It is important to note that all cropmarks exist on the higher, drier ridges of gravel, a pattern which continues into the development of the early church community on the drumlin of Tywyn. It is likely the settlements existed on the high ground standing as islands in the estuarine and marsh environments of the Dysynni. There is no evidence of large-scale agriculture in the Dysynni until the post medieval land improvements which is reflected in the Penllyn pollen record.

Chapter 8: Conclusions

8.1 Introduction

This section will outline the conclusions of the thesis. The results of the thesis will be summarised, and the research aims will be reviewed. The outcomes and implications of the research will be detailed, and finally, the recommendations for future work will be outlined. For reference, the aims and objectives were:

1. To reconstruct the Holocene sea-level changes in the Dysynni Valley.
2. To recreate the environmental and vegetation changes in the Dysynni Valley in order to fully understand past changes in the coastal environment.
3. To examine the impacts that the changing coastal environment had on human settlement in the area.

8.2 Thesis summary

Reconstructions of paleo sea level are essential in order to test the reliability of models of future sea-level change. GIA models require accurate sea-level reconstructions in order to assess their accuracy (see Shennan *et al.,* 2018). For example, some GIA models have previously predicted a mid Holocene high stand for the north coast of Wales (e.g., Peltier 1994, 2004; Peltier *et al.,* 2002), while others have not (e.g., Shennan *et al.,* 2018). Previously, the sea-level research on the west coast of Wales was limited to Borth, Ynyslas and Clarach, despite deep estuarine sediments in the valleys of northern Cardigan Bay (Blundell*,* 1969). The Dysynni and neighbouring valleys are vital for sea-level research as they are unaffected by isostatic uplift or subsidence (Shennan *et al.,* 2012). This PhD utilised microfossil analyses and transfer functions to produce sea-level index points (SLIPs) in order to contribute to the understanding of sea-level change in Wales.

The Dysynni Valley has a rich history from prehistory to modern day, but questions remain about the nature of settlement and land use in relation to the changing coastal environments. Much of the evidence for prehistoric land use is in the uplands. The valley bottom, however, is largely farmland, with settlement along the valley sides or situated on higher ground. The lowland field systems are more regular than the prehistoric, piecemeal enclosure of the uplands, implying that the present-day land use of the valley bottom is much more recent than the uplands. This thesis set out to apply environmental and sea-level reconstructions in the Dysynni Valley in order to better understand the history of the dynamic coastline in which people lived.

Fieldwork in the Dysynni Valley, revealed a ubiquitous blue-grey silt from the coast up to 7.76km inland, with the upper surface consistently between 0 and -1m OD. From Penllyn to Perfeddnant the blue-grey silt was overlain by woody peat. From Perfeddnant to the edge of Gesail the blue grey-silt was overlain by stiff, grey clay. From Gesail to the base of Craig yr Aderyn the sediments were dominated by grey silts, clays, and gravel with some organic horizons. A sample core was recovered from Penllyn at the coast, a second core from Perfeddnant 7km inland, and the final core at Gesail, at the base of Craig yr Aderyn. Loss on ignition and particle size analyses were undertaken on all three cores, followed by diatom and foraminifera analyses. Radiocarbon dates were taken from Penllyn and Perfeddnant, although only bulk dates were possible due to the lack of suitable macrofossils. The core at Penllyn was chosen to undertake pollen and microcharcoal analyses, as it represented the most complete sedimentary sequence. Limitations on lab access during the Covid pandemic prevented analyses of further proxies on the Perfeddnant and Gesail cores.

Transfer functions were applied to the cores at Penllyn and Perfeddnant. Two freshwater limiting points were produced from Perfeddnant and two SLIPs from Penllyn. The SLIPs from Penllyn indicate a rise in sea level from -7.82 m OD MHWST at 7787-7665 cal. BP to -4.91m OD MHWST at 6402-6297 cal. BP. The limiting points at Perfeddnant indicate that the site was freshwater at 9885-9547 cal. BP (D-AMS 031520) at -2.61m OD and 4782-4420 cal. BP (D-AMS 031519) at -0.44m OD. The sea-level reconstructions were compared to the regional work for Wales compiled by Shennan *et al,* (2018) and were found to correlate well. The SLIP at 7787-7665 cal. BP for Penllyn fills an important gap in the data for west Wales as previously there were only limiting points between 9779 cal. BP to 6813 cal. BP (Shennan *et al.,* 2018). Furthermore, the SLIPs from Penllyn fill important gaps in the data for north Wales between 7903 cal. BP (-6.09m OD MTL) to 5463 cal. BP (-1.58m OD MTL), helping to provide definition for the transition from the rapid sea-level rise of the early Holocene to the more gradual rise of the mid Holocene.

The coastal changes from the Dysynni Valley were reconstructed from the diatom, foraminifera, pollen, and sedimentological analyses. Microfossil analyses from Penllyn revealed a back barrier marsh existed at 7787-7665 cal. BP (D-AMS 028034, -3.39m OD), with intertidal mudflats, reed swamp, fen and alder carr. From the onset of blue-grey silt at -3.24m OD there was an expansion of intertidal mudflats with an increase in marine conditions and expansion of low marsh foraminifera. From -2.84m OD, alder carr developed in the surrounding area, with an expansion of reed swamp, saltmarsh, and tall herb fen. By -1.00m OD there was a return to brackish, high marsh conditions until 6402-6297 cal. BP (D-AMS 028035) at-0.48m OD which marked the end of marine conditions at Penllyn. Pollen analysis indicated that alder carr became dominant before being replaced by birch at 0.36m OD indicating drier conditions. The valley remained largely wooded with some evidence of disturbance indicators such as ribwort plantain. It was not until 1.27m OD, however, that widespread clearance and farming was seen in the pollen record, with trees declining, grass pollen increasing as well as many herb species such as ribwort plantain indicating disturbance.

The vegetation and sea-level reconstructions were compared to the present archaeological record, in order to further understand how people responded to sea-level change, and to inform future archaeological investigation in the area. There was possible evidence of Mesolithic burning at Penllyn, and the potential for future work on Mesolithic archaeology on the west coast was highlighted. Some small-scale clearance and anthropogenic impact became apparent at the top of the pollen profile while the valley remained wooded, followed by widespread clearance which likely represented land improvements in the late eighteenth century. It is unclear if there was continuous occupation of the valley from later prehistory or if there was a break in settlement until the establishment of the monastic community at Tywyn.

8.3 Achievement of research aims and objectives

8.3.1 Aim 1: To reconstruct the Holocene sea-level changes in the Dysynni Valley

This study has proven how valuable the estuarine sediments in Wales are for environmental and sea-level reconstructions. Paleoenvironmental analyses from three cores in the Dysynni Valley produced important sea-level reconstructions, which highlight the value of small-scale, valley wide sea-level reconstructions alongside regional sea-level work. This research also highlights the importance of multiproxy studies, with the diatom, foraminifera, and pollen vital to fully understand the sea-level changes in the Dysynni Valley. The study also highlights the value of using transfer functions although local training sets are required for the most accurate results.

Microfossil preservation was problematic, particularly for the Perfeddnant core, but the multiproxy analyses allowed a full assessment of the core despite issues with diatom preservation. Dating of estuarine sediments at Perfeddnant was challenging. Small*, in situ* microfossils would have been preferable for dating, but only large fragments of wood and reeds were present, which were not suitable for dating as they were not short-lived or *in situ*, and could easily have been transported as opposed to moss or fragile leaves (Turney *et al.,* 2000).

Transfer functions are very valuable for sea-level research, but a local training set is needed for more accurate reconstructions for the Dysynni Valley. The lack of local training sets made the application of transfer functions less effective, and in future more local training sets will be needed for sea-level research in Wales.

8.3.2 Aim 2: To recreate the environmental and vegetation changes in the Dysynni Valley in order to fully understand past changes in the coastal environment.

The multiproxy analyses completed for the three sites in the Dysynni Valley allowed the changes in the coastal and vegetation environment to be reconstructed for the past 9000 years. This produced a deeper understanding of vegetation and coastal changes that occurred throughout the Holocene.

Similar to the sea-level reconstruction, problems with dating and microfossil preservation occurred. For example, at Perfeddnant diatom preservation in the silt prevented analysis as only one sample could be counted to 250, and it was dominated by broken valves. Pollen, in particular, can be problematic due to differential pollen production by plants, as well as transport and preservation in estuarine sediments. For example, the over-representation of pine pollen in the estuarine silt at Penllyn represents buoyant pine pollen floating in the estuary, rather than local pine forest. Counting pollen from silt can be difficult when using HF preparation methods, however pollen preparation with heavy liquids, such as used in this thesis, allowed clean slides suitable for analysis.

8.3.3 Aim 3: To examine the impacts that the changing coastal environment had on human settlement in the area.

The sea level and vegetation reconstructions were compared to the archaeological record and allowed a more complete understanding of the occupation and land use history of the valley in relation to the changing environments. The sea level and vegetation reconstructions helped identify the nature and timing of coastal changes which will contribute to future archaeological research. The current study highlighted the need for future work in the excavation and dating of sites, particularly the cropmarks in the Dysynni Valley. It also highlighted the need for more extensive pollen work for future research into the archaeology of the Dysynni and northern Cardigan Bay area.

8.4 Implications of findings

The sea-level reconstruction from the Dysynni Valley contributes to the understanding of sea-level change through the Holocene for the UK. The SLIPs produced from Penllyn fill important gaps in the database for Wales, and help elucidate the change from rapid to moderate sea-level rise from the early to mid Holocene (Shennan *et al.,* 2018). Northern Cardigan Bay is an essential area for sea-level data as is affected by neither isostatic uplift nor subsidence (Shennan *et al.,* 2012). Previous GIA models (e.g., Peltier 1994, 2004; Peltier *et al.,* 2002) have predicted a mid Holocene high stand for north Wales, when past sea level extended higher than present. More recent models, such as Shennan *et al.,* (2018), however, show there was no mid Holocene high stand, which is supported by the sea-level reconstructions from the Dysynni Valley, although more data is needed to confirm it.

The environmental reconstructions from the Dysynni illustrate the dynamic coastal environments along northern Cardigan Bay. After rapid sea-level rise in the early Holocene, the diverse saltmarsh, fen and reed swamp environments gave way first to alder carr and then birch woodland. However, woodland and marshland persisted until the post medieval period in the Dysynni Valley. The environmental reconstructions aided the understanding of how past communities were likely to respond to changing environments, with a focus on upland pastoral land use until later prehistory or even the medieval period. There was unlikely to be widespread farming in the valley bottom until drainage in the late eighteenth century, with settlement prior to drainage focussing on the drier hills of gravel such as Tywyn and Bryncrug. The research also highlights the value of lowland estuarine sediments for multiproxy microfossil analyses, in particular for sea level and pollen investigations. Despite issues with dating, valuable diatom, foraminifera, and pollen evidence were recovered from the estuarine silts in the Dysynni Valley, and it is possible for further sea level and pollen research to be undertaken in similar sediments across the valleys of northern Cardigan Bay.

The research has also emphasised important questions that will direct future archaeological work in the Dysynni Valley. The potential for Mesolithic archaeological research has been highlighted, with possible evidence for Mesolithic burning of reedbeds. There is also potential for pollen and archaeological investigation into the cropmark settlement evidence in the valley bottom. From the pollen evidence anthropogenic impact did not occur until the top of the pollen core, but it is unclear if the cropmarks represent later prehistoric settlement through to the medieval period, or rather the settlement of the early medieval church community upon the establishment of Tywyn.

This study has also stressed the importance of multiproxy analyses within environmental archaeology. In particular, the use of sea-level research and analyses alongside pollen, microcharcoal and vegetation reconstruction in order to understand past populations responses to sea-level change.

8.5 Future work

There is potential for additional research into the coastal and environmental changes of the Dysynni Valley. Further sediment samples could be taken from the leading edge of the blue-grey silt in order to confirm the extent and timing of the horizontal and vertical sea-level changes in the Dysynni Valley. More dating evidence will be needed from *in situ* macrofossils as well as further pollen work, from both the valley bottom and uplands in order to establish the regional and local vegetation changes.

More sea-level data from northwest Wales is needed in order to better understand sea-level change in Wales and the UK as a whole. There are many estuarine silts present in the valleys in northern Cardigan Bay such as the Mawddach, and Tremadoc Bay, which have potential for sea-level research. Temporally, more data are needed prior to 6000 BP in order to define the rapidly rising sea levels of the early Holocene. There is very little natural saltmarsh left on the west coast, with the most well-preserved marsh located at Borth, which has been the focus of extensive sea-level research (Peck, 2018; Wilks 1977, 1979; Heyworth and Kidston, 1987). However, the deep estuarine sediments of northern Cardigan Bay are valuable resources for sea-level research which span into the Holocene. Furthermore, for future sea-level research into the west coast of Wales, more local, multiproxy training sets are needed, particularly for diatoms which are very diverse and form distinct local communities.

There is very little archaeological evidence for the Mesolithic on the west coast of Wales, despite deep Holocene sediments with a high potential for preservation of Mesolithic archaeology. The sea-level reconstructions of the present study have highlighted that sea levels rose 2.91 m between 7787-7665 cal. BP to 6402-6297 cal. BP. It is likely, therefore, that potential Mesolithic sites will now be submerged off the coast. Future work is needed to investigate the prehistoric coastline, with archaeological investigation paired with pollen analysis in order to better examine potential land use.

Further archaeological investigation is also needed on the cropmarks seen in the Dysynni Valley in order to properly date the onset of settlement. As well as excavation and dating of the archaeological sites, pollen analysis will be needed. Pollen cores from the peats and estuarine sediments of the valley bottom will need to be compared to cores from the uplands to identify local and regional pollen signals. Cores will also be needed from the archaeological sites themselves in order to identify local changes in land use. For example, cereal pollen does not transport very far and small agricultural fields are unlikely to be represented in pollen cores from a great distance.

The primary aim of the research was to reconstruct sea-level change through the Holocene in the Dysynni Valley, which has succeeded. Sea-level rose from -7.82 to -4.91m OD MHWST between 7787-7665 and 6402-6297 cal. BP, extending no further than 7.76km inland. This expands the understanding of sea-level and coastal change on the west coast of Wales, and proves the valleys of northern Cardigan Bay, such as the Mawddach, are valuable for sea-level and environmental research. The research has also demonstrated the value of sea-level research as part of environmental archaeology, helping to elucidate the nature of settlement changes in the Dysynni Valley and contributing to future archaeological research.

Bibliography

Abrantes, F., Gil, I., Lopes, C. and Castro, M., (2005) Quantitative diatom analyses—a faster cleaning procedure. *Deep Sea Research Part I: Oceanographic Research Papers*, 52, 189-198.

Adam, P. (1993) *Saltmarsh Ecology*. Cambridge University Press,

Adams, T.D., Haynes, J.R. & Walker, C.T. (1965) Boron in Holocene illites of the dovey estuary, wales, and its relationship to palaeosalinity in cyclothems. *Sedimentology*, 4, 189-195.

Allen, J.R.L. (1987) Late Flandrian shoreline oscillations in the Severn Estuary: the Rumney Formation at its typesite (Cardiff area). *Philosophical Transactions of the Royal Society of London B: Biological Sciences*, 315, 157-184.

Allen, J.R.L. (1990) The Severn Estuary in southwest Britain: its retreat under marine transgression, and fine-sediment regime. *Sedimentary Geology*, 66, 13-28.

Allen, J.R.L. (2000) Holocene coastal lowlands in NW Europe: autocompaction and the uncertain ground. *Geological Society, London, Special Publications*, 175, 239-252.

Allen, J.R.L. & Rae, J.E. (1987) Late Flandrian shoreline oscillations in the Severn Estuary: a geomorphological and stratigraphical reconnaissance. *Philosophical Transactions of the Royal Society of London. Series B, Biological Sciences*, 185-230.

Allen, J.R.L. & Thornley, D.M. (2004) Laser granulometry of Holocene estuarine silts: effects of hydrogen peroxide treatment. *The Holocene*, 14, 290-295.

Allen, J.R.L. & Haslett, S.K. (2007) The Holocene estuarine sequence at Redwick, Welsh Severn Estuary Levels, UK: the character and role of silts. *Proceedings of the Geologists' Association*, 118, 157-174.

Allen, J.R.L., Scourse, J.D., Hall, A.R. and Coope, G.R. (2009) Palaeoenvironmental context of the Late-glacial woolly mammoth (Mammuthus primigenius) discoveries at Condover, Shropshire, UK. *Geological Journal*, 44, 414-446.

Allen, P.M., Jackson, A.A. & Dunkley, P.N. (1985) *Geology of the country around Harlech*. Natural Environment Research Council: H.M.S.O., London.

Allen, M.M., Taratoot, M. & Adams, P.W. (1999) Soil compaction and disturbance from skyline and mechanized partial cuttings for multiple resource objectives in western and northeastern Oregon, USA. 107-117.

Amesbury, M.J., Charman, D.J., Fyfe, R.M., Langdon, P.G. and West, S. (2008) Bronze Age upland settlement decline in southwest England: Testing the climate change hypothesis. *Journal of Archaeological Science*, 35, 87-98.

Andersen, S.T. (1979) Identification of wild grass and cereal pollen [fossil pollen, Annulus diameter, surface sculpturing]. *Aarbog. Danmarks Geologiske Undersoegelse (Denmark)*,

Anderson, J.G.C. & Blundell, C.R.K. (1965) The sub-drift rock-surface and buried valleys of the Cardiff district. *Proceedings of the Geologists' Association*, 76, 367-IN2.

Austin, W.E.N. (1991) *Late Quaternary benthonic Foraminiferal stratigraphy the western U.K. Continental Shelf.* Prifysgol Bangor University.

Avnaim-Katav, S., Gehrels, W.R., Brown, L.N., Fard, E. and MacDonald, G.M., (2017) Distributions of salt-marsh foraminifera along the coast of SW California, USA: Implications for sea-level reconstructions. *Marine Micropaleontology*, 131, 25-43.

Bakels, C.C. (2000) Pollen diagrams and prehistoric fields: the case of Bronze Age Haarlem, the Netherlands. *Review of Palaeobotany and Palynology*, 109, 205-218.

Ballantyne, C.K. (2012) Chronology of glaciation and deglaciation during the Loch Lomond (Younger Dryas) Stade in the Scottish Highlands: implications of recalibrated 10 Be exposure ages. *Boreas*, 41, 513-526.

Barber, K. (1982) peat-bog stratigraphy as a proxy climate record. In *Climatic Change in Later Prehistory,* (Ed, Harding, A.F.) Edinburgh University Press, Edinburgh, 3-113

Bard, E., Arnold, M., Hamelin, B., Tisnerat-Laborde, N. and Cabioch, G., (1990) Radiocarbon calibration by means of mass 230Th/234U and 14C ages of corals: an updated database including samples from Barbados, Mururoa and Tahiti. *Radiocarbon*, 40, 1085-1092.

Barlow, N.L., Shennan, I., Long, A.J., Gehrels, W.R., Saher, M.H., Woodroffe, S.A. and Hillier, C., (2013) Salt marshes as late Holocene tide gauges. *Global and planetary change*, 106, 90-110.

Barlow, N.L., Long, A.J., Saher, M.H., Gehrels, W.R., Garnett, M.H. and Scaife, R.G., (2014) Salt-marsh reconstructions of relative sea-level change in the North Atlantic during the last 2000 years. *Quaternary science reviews*, 99, 1-16.

Battarbee, R.W., Jones, V.J., Flower, R.J., Cameron, N.G., Bennion, H., Carvalho, L. and Juggins, S. (2001) Diatoms. In *Tracking Environmental Change Using Lake Sediments: Terrestrial, Algal and Siliceous Indicators,* (Eds, Smol, J.P., Birks, H.J. and Last, W.M.) Kluwer Academic Publishers, Dordrecht, pp. 155-202.

Beaudoin, A. (2003) A comparison of two methods for estimating the organic content of sediments. *Journal of Paleolimnology*, 29(3), 387-390.

Bedlington, D.J. (1994) *Holocene sea-level changes and crustal movements in North Wales and Wirral.* Durham University.

Behre, K.-E. (2007) Evidence for Mesolithic agriculture in and around central Europe? *Vegetation History and Archaeobotany*, 16, 203-219.

Bell, M. (2007) *Prehistoric coastal communities: the Mesolithic in Western Britain*. Council for British Archaeology, York.

Bell, M. & Neumann, H. (1997) Prehistoric intertidal archaeology and environments in the Severn Estuary, Wales. *World Archaeology*, 29, 95-113.

Bell, M., Caseldine, A.E. & Neumann, H. (2002) *Prehistoric intertidal archaeology in the Welsh Severn Estuary*. Archaeology Data Service,

Bendle, J.M. & Glasser, N.F. (2012) Palaeoclimatic reconstruction from Lateglacial (Younger Dryas Chronozone) cirque glaciers in Snowdonia, North Wales. *Proceedings of the Geologists' Association*, 123, 130-145.

Bennett, K.D. & Birks, H.J.B. (1990) Postglacial history of alder (Alnus glutinosa (L.) Gaertn.) in the British Isles. *Journal of Quaternary Science*, 5, 123-133.

Bennett, M.R. & Boulton, G.S. (1993) Deglaciation of the Younger Dryas or Loch Lomond Stadial ice-field in the northern Highlands, Scotland. *Journal of Quaternary Science*, 8, 133-145.

Berglund, B.E. (2003) Human impact and climate changes—synchronous events and a causal link? *Quaternary International*, 105, 7-12.

Best, L.A. (2016) *Late Holocene relative sea-level change and the implications for the groundwater resource, Humber Estuary, UK*. University of York.

Beuselinck, L., Govers, G., Poesen, J., Degraer, G. and Froyen, L. (1998) Grain-size analysis by laser diffractometry: comparison with the sieve-pipette method. *Catena*, 32, 193-208.

Beyens, L. (1982) Problems in diatom analysis of deposits: allochtonous valves and fragmentation. *Geologie en Mijnbouw,* 61(2), 159-162.

Bickerdike, H.L., Evans, D.J.A., Ó Cofaigh, C. and Stokes, C.R. (2016) The glacial geomorphology of the Loch Lomond Stadial in Britain: a map and geographic information system resource of published evidence. *Journal of Maps*, 12, 1178-1186.

Bird, M.I., Fifield, L.K., Chua, S. and Goh, B. (2004) Calculating Sediment Compaction for Radiocarbon Dating of Intertidal Sediments. *Radiocarbon*, 46, 421-435.

Birks, H.J.B. (1989) Holocene Isochrone Maps and Patterns of Tree-Spreading in the British Isles. *Journal of Biogeography*, 16, 503-540.

Birks, H.J.B. (1995) Quantitative Palaeoenvironmental Reconstructions. In *Statistical Modelling of Quaternary Science Data, Technical Guide 5,* (Ed, Maddy, D.) Quaternary Research Association, Cambridge, pp. 161-255.

Birks, H.J.B. (2010) Numerical methods for the analysis of diatom assemblage data. *The diatoms: applications for the environmental and earth sciences, 2nd edn. Cambridge University Press, Cambridge*, 23-54.

Birks, H.J.B., Braak, C.T., Line, J.M., Juggins, S. and Stevenson, A.C. (1990) Diatoms and pH reconstruction. *Philosophical transactions of the royal society of London. B, Biological Sciences*, 327, 263-278.

Birks, H.J.B., Mackay, A., Battarbee, R.W., Birks, J. and Oldfield, F. (2014) Quantitative palaeoenvironmental reconstructions from Holocene biological data. *Global change in the Holocene,* Routledge, 123-139.

Björck, S. *et al.* (1998) A high-resolution14C dated sediment sequence from southwest Sweden: age comparisons between different components of the sediment. *Journal of Quaternary Science: Published for the Quaternary Research Association*, 13, 85-89.

Björck, S., Persson, T. & Kristersson, I. (1978) Comparison of two concentration methods for pollen in minerogenic sediments. *Geologiska Föreningen i Stockholm Förhandlingar*, 100, 107-111.

Blackford, J.J. & Chambers, F.M. (1991) Proxy records of climate from blanket mires: evidence for a Dark Age (1400 BP) climatic deterioration in the British Isles. *The Holocene*, 1, 63-67.

Blott, S.J. & Pye, K. (2006) Particle size distribution analysis of sand-sized particles by laser diffraction: an experimental investigation of instrument sensitivity and the effects of particle shape. *Sedimentology*, 53, 671-685.

Blundell, D.J., Griffiths, D.H. & King, R.F. (1969) Geophysical investigations of buried river valleys around Cardigan Bay. *Geological Journal*, 6, 161-180.

Bostock, J.L. (1980) *The history of the vegetation of the Berwyn mountains, North Wales, with emphasis on the development of the blanket mire*. University of Manchester.

Boulton, G.S. & Worsley, P. (1965) Late Weichselian Glaciation in the Cheshire-Shropshire Basin. *Nature*, 207, 704-706.

Botterill, E.M. (1988) *A palaeoecological study of Cors Gyfelog and Tre'r Gof: lowland mines of north west Wales*. Keele University.

Bowen, D.Q. (1981) The 'South Wales end-moraine': fifty years after. *The Quaternary in Britain*, 60-67.

Bowen, E.G. & Gresham, C.A. (1967) *History of Merioneth. Vol. 1: From the earliest times to the age of the native princes*. Merioneth Historical and Record Society, Dolgellau.

Boyle, J. (2004) A comparison of two methods for estimating the organic matter content of sediments. *Journal of Paleolimnology*, 31, 125-127.

Bradley, S.L., Milne, G.A. & FN, T. (2009) Glacial isostatic adjustment of the British Isles: new constraints from GPS measurements of crustal motion. *Geophysical Journal International*, 178, 14-22.

Bradley, S.L., Milne, G.A., Shennan, I. and Edwards, R. (2011) An improved glacial isostatic adjustment model for the British Isles. *Journal of Quaternary Science*, 26, 541-552.

Brain, M.J., Long, A.J., Woodroffe, S.A., Petley, D.N., Milledge, D.G. and Parnell, A.C. (2012) Modelling the effects of sediment compaction on salt marsh reconstructions of recent sea-level rise. *Earth and Planetary Science Letters*, 345-348, 180-193.

Brain, M.J., Kemp, A.C., Horton, B.P., Culver, S.J., Parnell, A.C. and Cahill, N. (2015) Quantifying the contribution of sediment compaction to late Holocene salt-marsh sea-level reconstructions, North Carolina, USA. *Quaternary Research*, 83, 41-51.

Brain, M.J., Kemp, A.C., Hawkes, A.D., Engelhart, S.E., Vane, C.H., Cahill, N., Hill, T.D., Donnelly, J.P. and Horton, B.P. (2017) Exploring mechanisms of compaction in salt-marsh sediments using Common Era relative sea-level reconstructions. *Quaternary Science Reviews*, 167, 96-111.

Brock, F., Lee, S., Housley, R.A. and Ramsey, C.B. (2011) Variation in the radiocarbon age of different fractions of peat: A case study from Ahrenshöft, northern Germany. *Quaternary Geochronology*, 6, 550-555.

Bronk Ramsay, C. (2001) Development of the radiocarbon calibration program. *Radiocarbon*, 43, 355-363.

Bronk Ramsay, C. (2005) *OxCal version 3.10*. Oxford, UK, University of Oxford, downloaded in November,

Bronk Ramsay, C. (2011) Development of the radiocarbon calibration program Oxcal. *Radiocarbon*, 43, 255-363.

Bronk Ramsay, C. (2017) Methods for Summarizing Radiocarbon Datasets. *Radiocarbon*, 59, 1809-1833.

Bronk Ramsay, C. & Lee, S. (2013) Recent and planned developments of the program OxCal. *Radiocarbon*, 55, 720-730.

Bronk Ramsey, C., Scott, M. & van der Plicht, H. (2013) Calibration for archaeological and environmental terrestrial samples in the time range 26–50 ka cal BP. *Radiocarbon*, 55, 2021-2027.

Brooks, A.J., Bradley, S.L., Edwards, R.J., Milne, G.A., Horton, B. and Shennan, I. (2008) Postglacial relative sea-level observations from Ireland and their role in glacial rebound modelling. *Journal of Quaternary Science*, 23, 175-192.

Brown (1988) The palaeoecology of Alnus (alder) and the Postglacial history of floodplain vegetation. Pollen percentage and influx data from the West Midlands, United Kingdom. *New Phytologist*, 110, 425-436.

Brown, V.H., Evans, D.J.A. & Evans, I.S. (2011) The Glacial Geomorphology and Surficial Geology of the South-West English Lake District. *Journal of Maps*, 7, 221-243.

Brown, V.H., Evans, D.J., Vieli, A. and Evans, I.S. (2013) The Younger Dryas in the English Lake District: reconciling geomorphological evidence with numerical model outputs. *Boreas*, 42, 1022-1042.

Cadw (2022) Registered historic landscapes | Cadw. Available at: https://cadw.gov.wales/advice-support/historic-assets/conservation-areas-and-other-historic-assets/other-historic-assets-0 (Accessed: 28 November 2022).

Caffrey, M.A. & Horn, S.P. (2013) The use of lithium heteropolytungstate in the heavy liquid separation of samples which are sparse in pollen. *Palynology*, 37, 143-150.

Cahoon, D.R. (2006) A review of major storm impacts on coastal wetland elevations. *Estuaries and Coasts*, 29, 889-898.

Cahoon, D.R., Lynch, J.C., Perez, B.C., Segura, B., Holland, R.D., Stelly, C., Stephenson, G. and Hensel, P. (2002) High-Precision Measurements of Wetland Sediment Elevation: I. Recent Improvements to the Sedimentation-Erosion Table. *Journal of Sedimentary Research*, 72, 730-733.

Campbell, J.F.E., Fletcher, W.J., Hughes, P.D. and Shuttleworth, E.L. (2016) A comparison of pollen extraction methods confirms dense-media separation as a reliable method of pollen preparation. *Journal of Quaternary Science*, 31, 631-640.

Campbell, S. & Bowen, D.Q. (1989) *Quaternary of Wales*. Nature Conservancy Council,

Cann, J.H., Belperio, A.P. & Murray-Wallace, C.V. (2000) Late Quaternary Paleosealevels and Paleoenvironments inferred from foraminifera, northern spencer gulf, south Australia. *Journal of Foraminiferal Research*, 30, 29-53.

Caseldine, A.E. (2017) Refresh of the research agenda for Wales: palaeoenvironments. Available at: https://www.archaeoleg.org.uk/pdf/review2017/palaeoenvreview2017.pdf (Accessed: 28 November 2022).

Caseldine, A., Evans, J.G. & Evans, J.G. (1990) *Environmental archaeology in Wales*. Department of Archaeology Saint David's University College, Caseldine, A.E., Griffiths, C.J., Jones, S., Peck, I. and Smith, G. (2016) A Late Post-Glacial environmental record and Early Neolithic intertidal peat land surface at Porth Neigwl, Llŷn Peninsula, Gwynedd. *Studia Celtica*, 50, 1-18.

Chambers, F.M. (1998) The palynological investigations. In *The Graeanog Ridge: the Evolution of a Farming Landscape and its Settlements in North-West Wales,* Cambrian Archaeological Monographs 6, pp. 52-63.

Chambers, F. & Price, S. (1985) Paleoecology of Alnus (alder) - Early Post-Glacial Rise in a Valley Mire, Northwest Wales. *New Phytologist*, 101, 333-344.

Chambers, F.M. & Price, S.M. (1988) The Environmental Setting of Erw-wen and Moel y Gerddi : Prehistoric Enclosures in Upland Ardudwy, North Wales. *The Environmental Setting of Erw-wen and Moel y Gerddi : Prehistoric Enclosures in Upland Ardudwy, North Wales*, 54, 93-100.

Chambers, F.M. & Elliott, L. (1989) Spread and Expansion of Alnus Mill. In the British Isles: Timing, Agencies and Possible Vectors. *Journal of Biogeography*, 16, 541-550.

Charlesworth, J.K. (1929) The South Wales End-Moraine. *Quarterly Journal of the Geological Society*, 85, 335-358.

Chiverrell, R.C. & Thomas, G.S.P. (2010) Extent and timing of the Last Glacial Maximum (LGM) in Britain and Ireland: a review. *Journal of Quaternary Science*, 25, 535-549.

Clark, C.D., Ely, J.C., Greenwood, S.L., Hughes, A.L., Meehan, R., Barr, I.D., Bateman, M.D., Bradwell, T., Doole, J., Evans, D.J. and Jordan, C.J. (2018) BRITICE Glacial Map, version 2: a map and GIS database of glacial landforms of the last British–Irish Ice Sheet. *Boreas*, 47, 11-e8.

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

Clark, S.H.E. & Edwards, K.J. (2004) Elm bark beetle in Holocene peat deposits and the northwest European elm decline. *Journal of Quaternary Science*, 19, 525-528.

CLIMAP Project Members, (1976) The Surface of the Ice-Age Earth: Quantitative geologic evidence is used to reconstruct boundary conditions for the climate 18,000 years ago. *Science*, 191, 1131-1137.

Cook, I. (2019) *Climate Change and Cultural Heritage: developing a landscape-scale vulnerability framework to measure and manage the impact of climate change on coastal historic landscapes*. PhD Thesis, University of Sheffield.

Cooper, S.R. (1999) Estuarine paleoenvironmental reconstructions using diatoms. In *The Diatoms: Applications for the Environmental and Earth Sciences,* (Eds, Stoermer, E.F. & Smol, J.P.) Cambridge University Press, Cambridge, pp. 352-373.

Crimes, T.P., Chester, D.K. & Thomas, G.S.P. (1992) Exploration of sand and gravel resources by geomorphological analysis in the glacial sediments of the Eastern Lleyn Peninsula, Gwynedd, North Wales. *Engineering Geology*, 32, 137-156.

Culver, S.J. & Banner, F.T. (1978) Foraminiferal assemblages as flandrian palaeo-environmental indicators. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 24, 53-72.

Dackombe, R. & Gardiner, V. (2020) *Geomorphological Field Manual*. Routledge,

de Groot, A.V., Veeneklaas, R.M. & Bakker, J.P. (2011) Sand in the salt marsh: Contribution of high-energy conditions to salt-marsh accretion. *Marine Geology*, 282, 240-254.

de Rijk, S. & Troelstra, S. (1999) The application of a foraminiferal actuo-facies model to salt-marsh cores. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 149, 59-66.

de Vries, H., Barendsen, G.W. & Waterbolk, H.T. (1958) Groningen radiocarbon dates II. *Science*, 127, 129-137.

De Vries, H. (1958) Radiocarbon dates for Upper Eem and Würm-interstadial samples. *age*, 39800, 1000.

Denys, L. & de Wolf, H. (1999) Diatoms as indicators of coastal paleoenvironments and relative sea-level change. In *The Diatoms: Applications for the Environmental and Earth Sciences,* (Eds, Stoermer, E.F. & Smol, J.P.) Cambridge University Press, Cambridge, pp. 277-297.

Denys, L. (1991/2) A check list of the diatoms in the Holocene deposits of the western Belgian coastal plain, with a survey of their apparent ecological requirements. I. Introduction, ecological code and complete list.

Denys, L. (1994) Diatom assemblages along a former intertidal gradient: A palaeoecological study of a subboreal clay layer (Western coastal plain, Belgium). *Netherland Journal of Aquatic Ecology*, 28, 85-96.

Devoy, R.J.N. (1979) Flandrian sea level changes and vegetational history of the lower Thames estuary. *Phil. Trans. R. Soc. Lond. B*, 285, 355-407.

dos Santos-Fischer, C.B., Corrêa, I.C.S., Weschenfelder, J., Torgan, L.C. and Stone, J.R. (2016) Paleoenvironmental insights into the Quaternary evolution of the southern Brazilian coast based on fossil and modern diatom assemblages. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 446, 108-124.

Druce, D. (2000) *Mesolithic to Romano-British archaeology and environmental change of the Severn Estuary, England*. University of Bristol.

Edwards, K.J. & Hirons, K.R. (1984) Cereal pollen grains in pre-elm decline deposits: Implications for the earliest agriculture in Britain and Ireland. *Journal of Archaeological Science*, 11, 71-80.

Edwards, R.J. (2006) Mid- to late-Holocene relative sea-level change in southwest Britain and the influence of sediment compaction. *The Holocene*, 16, 575-587.

Edwards, R.J. & Horton, B.P. (2000) Reconstructing relative sea-level change using UK salt-marsh foraminifera. *Marine Geology*, 169, 41-56.

Edwards, R.J., Van De Plassche, O., Gehrels, W.R. and Wright, A.J. (2004) Assessing sea-level data from Connecticut, USA, using a foraminiferal transfer function for tide level. *Marine Micropaleontology*, 51, 239-255.

Elliott, E. (2015) *Holocene sea-level change at the Steart Peninsula, Somerset: development and application of a multi-proxy sea-level transfer function for the Severn Estuary region*. PhD thesis, University of the West of England, Bristol.

Engelhart, S.E. & Horton, B.P. (2012) Holocene sea level database for the Atlantic coast of the United States. *Quaternary Science Review*, 54, 12-25.

Etienne, J.L., Jansson, K.N., Glasser, N.F., Hambrey, M.J., Davies, J.R., Waters, R.A., Maltman, A.J. and Wilby, P.R. (2006) Palaeoenvironmental interpretation of an ice-contact glacial lake succession: an example from the late Devensian of southwest Wales, UK. *Quaternary Science Reviews*, 25, 739-762.

Evans, D.J. & Thompson, M.S. (1979) The geology of the central Bristol Channel and the Lundy area, South Western Approaches, British Isles. *Proceedings of the Geologists' Association*, 90, 1-14.

Evans, D.J.A., Clark, C.D. & Mitchell, W.A. (2005) The last British Ice Sheet: A review of the evidence utilised in the compilation of the Glacial Map of Britain. *Earth-Science Reviews*, 70, 253-312.

Faegri, K., Kaland, P.E. and Krzywinski, K. (1964) *Textbook of pollen analysis*. John Wiley & Sons Ltd.,

Faegri, K., Inversen &, K. (1989) *Textbook of pollen analysis.* John Wiley & Sons Ltd., Fisher, P., Aumann, C., Chia, K., O'Halloran, N. and Chandra, S. (2017) Adequacy of laser diffraction for soil particle size analysis. *PloS one*, 12, e0176510.

Fletcher, W.J. & Hughes, P.D. (2017) Anthropogenic trigger for late Holocene soil erosion in the Jebel toubkal, high Atlas, Morocco. *Catena*, 149, 713-726.

Flower, R.J. (1993) Diatom preservation: experiments and observations on dissolution and breakage in modern and fossil material. *Developments in Hydrobiology*, 473-484.

Forster, M. & Flenley, J.R. (1993) Pollen purification and fractionation by equilibrium density gradient centrifugation. *Palynology*, 17, 137-155.

Foster, H.D. (1968) *The glaciation of the Harlech Dome.* University College London (University of London).

Foster, H.D. (1970) Establishing the Age and Geo Morphological Significance of Sorted Stone-Stripes in the Rhinog Mountains, North Wales. *Geografiska Annaler: Series A, Physical Geography*, 52, 96-102.

French, P.W. (1996) Long-term Temporal Variability of Copper, Lead, and Zinc in Salt Marsh Sediments of the Severn Estuary, UK. *Mangroves and Salt Marshes*, 1, 59-68.

Frost (2012) Proposed Ysgol Bro Dysynni, Llaegrin, Merionnydd, SH59787 05100: Archaeological Desk-Based Assessment, February 2012. *Casterling Archaeology*, Report no. 375,

Gale, S.J. & Hoare, P.G. (1991) *Quaternary sediments: petrographic methods for the study of unlithified rocks*. Belhaven Press; Halsted Press, London : New York N.Y.

Garrard, R.A. & Dobson, M.R. (1974) The nature and maximum extent of glacial sediments off the west coast of Wales. *Marine Geology*, 16, 31-44.

Garrett, E. (2013) *Biostratigraphic constraints on megathrust earthquake deformation history in south central Chile*. Durham University.

Gasse, F., Juggins, S. & Khelifa, L.B. (1995) Diatom-based transfer functions for inferring past hydrochemical characteristics of African lakes. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 117, 31-54.

Gee, G.W. & Bauder, J.W. (1986) Particle-size analysis. *Methods of soil analysis: Part 1 Physical and mineralogical methods*, 5, 383-411.

Gehrels, W.R. (1999) Middle and late Holocene sea-level changes in eastern Maine reconstructed from foraminiferal saltmarsh stratigraphy and AMS 14C dates on basal peat. *Quaternary Research*, 52, 350-359.

Gehrels, W.R. (2000) Using foraminiferal transfer functions to produce high-resolution sea-level records from salt-marsh deposits, Maine, USA. *The Holocene*, 10, 367-376.

Gehrels, W.R. (2007) Microfossil Reconstructions. In *Encyclopedia of Quaternary Science,* (Ed, Elias, S.) Elsevier, Oxford, pp. 3015-3024.

Gehrels, W. R. (2009) Rising sea levels as an indicator of global change. In *Climate Change,* Elsevier, pp. 325-336.

Gehrels, W.R. (2010) Late Holocene land- and sea-level changes in the British Isles: implications for future sea-level predictions. *Quaternary science reviews*, 29, 1648-1669.

Gehrels, W.R., Roe, H.M. and Charman, D.J., (2001) Foraminifera, testate amoebae and diatoms as sea-level indicators in UK saltmarshes: a quantitative multiproxy approach. *Journal of Quaternary Science*, 16, 201-220.

Gehrels, W.R., Milne, G.A., Kirby, J.R., Patterson, R.T. and Belknap, D.F. (2004) Late Holocene sea-level changes and isostatic crustal movements in Atlantic Canada. *Quaternary International*, 120, 79-89.

Gehrels, W.R., Kirby, J.R., Prokoph, A., Newnham, R.M., Achterberg, E.P., Evans, H., Black, S. and Scott, D.B. (2005) Onset of recent rapid sea-level rise in the western Atlantic Ocean. *Quaternary science reviews*, 24, 2083-2100.

Gehrels, W.R., Szkornik, K., Bartholdy, J., Kirby, J.R., Bradley, S.L., Marshall, W.A., Heinemeier, J. and Pedersen, J.B. (2006) Late Holocene sea-level changes and isostasy in western Denmark. *Quaternary Research*, 66, 288-302.

Gehrels, W.R. & Long, A.J. (2008) Sea level is not level: the case for a new approach to predicting UK sea-level rise. *Geography*, 93, 11-16.

Glasser, N.F., Hambrey, M.J., Huddart, D., Gonzalez, S., Crawford, K.R. and Maltman, A.J. (2001) Terrestrial glacial sedimentation on the eastern margin of the Irish Sea basin: Thurstaston, Wirral. *Proceedings of the Geologists' Association*, 112, 131-146.

Glasser, N.F., Hughes, P.D., Fenton, C., Schnabel, C. and Rother, H. (2012) ^{10}Be and ^{26}Al exposure-age dating of bedrock surfaces on the Aran ridge, Wales: evidence for a thick Welsh Ice Cap at the Last Glacial Maximum. *Journal of Quaternary Science*, 27, 97-104.

Godwin, H. (1943) Coastal peat beds of the British Isles and North Sea: presidential address to the British Ecological Society 1943. *Journal of Ecology*, 31, 199-247.

Godwin, H. (1958) Radiocarbon dating of the eustatic rise in ocean-level. *Nature*, 181, 1518-1519.

Godwin, H. (1962) Half-life of radiocarbon. *Nature*, 195, 984-984.

Godwin, H. & Newton, L. (1938) The submerged forest at Borth and Ynyslas, Cardiganshire. *New Phytologist*, 37, 333-344.

Godwin, H. & Willis, E.H. (1959) Cambridge University Natural Radiocarbon Measurements I. *Radiocarbon*, 1, 63-75.

Godwin, H. & Willis, E. (1961) Cambridge University natural radiocarbon measurements III. *Radiocarbon*, 3, 60-76.

Godwin, H. & Willis, E. (1964) Cambridge University natural radiocarbon measurements IV. *Radiocarbon*, 6, 116-137.

Godwin, H. & Willis, E.H. (1964) Cambridge University Natural Radiocarbon Measurements VI. *Radiocarbon*, 6, 116-137.

Goldstein, S.T. & Watkins, G.T. (1999) Taphonomy of salt marsh foraminifera: an example from coastal Georgia. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 149, 103-114.

Golledge, N.R. (2010) Glaciation of Scotland during the Younger Dryas stadial: a review. *Journal of Quaternary Science: Published for the Quaternary Research Association*, 25, 550-566.

Graham, A.G.C., Lonergan, L. & Stoker, M.S. (2007) Evidence for Late Pleistocene ice stream activity in the Witch Ground Basin, central North Sea, from 3D seismic reflection data. *Quaternary Science Reviews*, 26, 627-643.

Gray, J.M. & Lowe, J.J. (1982) Problems in the interpretation of small-scale erosional forms on glaciated bedrock surfaces: examples from Snowdonia, North Wales. *Proceedings of the Geologists' Association*, 93, 403-414.

Griffiths, H., Yorke, L. & Griffiths, S. (2013) Historical channel change and sediment dynamics in a heavily managed and protected small catchment: The Afon Dysynni, north Wales. *8th IAG International Conference on Geomorphology*,

Griffiths, S., Sturt, F., Dix, J.K., Gearey, B. and Grant, M.J. (2015) Chronology and palaeoenvironmental reconstruction in the sub-tidal zone: a case study from Hinkley Point. *Journal of Archaeological Science*, 54, 237-253.

Grimm, E. C. (2004) Tilia and TGView 2.0. 2. *Illinois state museum, research and collections center, Springfield*,

Grimm, E.C. (2018) Tilia software v. 2.0. 60. *Illinois State Museum, Springf*,

Groenman-van Waateringe, W. (1993) The effects of grazing on the pollen production of grasses. *Vegetation History and Archaeobotany*, 2, 157-162.

Hageman, P.I.B. (1969) Development of the western part of the Netherlands during the Holocene. *Geologie en Mijinbouw*, 48, 373-388.

Hambrey, M.J., Davies, J.R., Glasser, N.F., Waters, R.A., Dowdeswell, J.A., Wilby, P.R., Wilson, D. and Etienne, J.L. (2001) Devensian glacigenic sedimentation and landscape evolution in the Cardigan area of southwest Wales. *Journal of Quaternary Science*, 16, 455-482.

Hamilton, S. & Shennan, I. (2005) Late Holocene relative sea-level changes and the earthquake deformation cycle around upper Cook Inlet, Alaska. *Quaternary science reviews*, 24, 1479-1498.

Handa, S. & Moore, P.D. (1976) Studies in the Vegetational History of Mid Wales. *New Phytologist*, 77, 205-225.

Harkness, D.D. & Wilson, H.W. (1974) Scottish Universities Research and Reactor Centre Radiocarbon Measurements II. *Radiocarbon*, 16, 238-251.

Hart, T.J. (1957) Notes on practical methods for the study of marine diatoms. *Journal of the Marine Biological Association of the United Kingdom*, 36, 593-597.

Hartley, B. (1996) *An atlas of British diatoms*. Balogh Scientific Books,

Haslett, S.K., Davies, P. & Strawbridge, F. (1998) Reconstructing Holocene sea-level change in the Severn Estuary and Somerset Levels: The foraminifera connection. *Archaeology in the Severn Estuary*, 8, 29-40.

Haslett, S.K., Howard, K.L., Margetts, A.J. and Davies, P. (2001a) Holocene stratigraphy and evolution of the northern coastal plain of the Somerset Levels, UK. *Proceedings of the Cotteswold Naturalists' Field Club*, 42, 78-88.

Haslett, S.K., Strawbridge, F., Martin, N.A. and Davies, C.F.C. (2001b) Vertical saltmarsh accretion and its relationship to sea-level in the Severn Estuary, UK: an investigation using foraminifera as tidal indicators. *Estuarine, Coastal and Shelf Science*, 52, 143-153.

Hatte, C. & Jull, A.J.T. (2007) Radiocarbon Dating: Plant Macrofossils. In *S.A. Elias, (ed.) Encylopedia of Quaternary Science,* Elsevier Scientific Publishing, Amsterdam, pp. 2958-2965.

Hawkins, A.B. (1971) Sea-level changes around south-west England. *Colston papers*, 23, 67-87.

Haynes, J. & Dobson, M. (1968) Physiography, foraminifera and sedimentation in the Dovey Estuary (Wales). *Geological Journal*,

Heiri, O., Lotter, A.F. & Lemcke, G. (2001) Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. *Journal of Paleolimnology*, 25, 101-110.

Heiri, O., Brooks, S.J., Birks, H.J.B. and Lotter, A.F. (2011) A 274-lake calibration data-set and inference model for chironomid-based summer air temperature reconstruction in Europe. *Quaternary Science Reviews*, 30, 3445-3456.

Hendey, N.I. (1964) An introductory account of the smaller algae of British coastal waters. *Fishery Investigation Series. Pt. 5: Bacillariophyceae (Diatoms)*, 317,

Hewlett, R. & Bimie, J. (1996) Holocene environmental change in the inner Severn estuary, UK: an example of the response of estuarine sedimentation to relative sea-level change. *The Holocene*, 6, 49-61.

Heyworth, A. & Kidson, C. (1982) Sea-level changes in southwest England and Wales. *Proceedings of the Geologists' Association*, 93, 91-111.

Heyworth, A., Kidson, C. & Wilks, P. (1985) Late-glacial and Holocene sediments at Clarach Bay, near Aberystwyth. *The Journal of Ecology*, 459-480.

Hibbert, F.A. & Switsur, V.R. (1976) Radiocarbon Dating of Flandrian Pollen Zones in Wales and Northern England. *New Phytologist*, 77, 793-807.

Hiemstra, J.F., Evans, D.J., Scourse, J.D., McCarroll, D., Furze, M.F. and Rhodes, E. (2006) New evidence for a grounded Irish Sea glaciation of the Isles of Scilly, UK. *Quaternary Science Reviews*, 25, 299-309.

Hijma, M.P., Engelhart, S.E., Törnqvist, T.E., Horton, B.P., Hu, P. and Hill, D.F. (2015) A protocol for a geological sea-level database. *Handbook of Sea-Level Research, edited by: Shennan, I., Long, AJ, and Horton, BP, Wiley Blackwell*, 536-553.

Hill, T.C., Woodland, W.A., Spencer, C.D. and Marriott, S.B. (2007) Holocene sea-level change in the Severn Estuary, southwest England: a diatom-based sea-level transfer function for macrotidal settings. *The Holocene*, 17, 639-648.

Hillam, J., Groves, C.M., Brown, D.M., Baillie, M.G., Coles, J.M. and Coles, B.J., (1990) Dendrochronology of the English Neolithic. *Antiquity*, 64, pp. 210-220.

Hinton, A.C. (1996) Tides in the northeast Atlantic: considerations for modelling water depth changes. *Quaternary Science Reviews*, 15, 873-894.

Horton, B.P., Edwards, R.J. & Lloyd, J.M. (1999) UK intertidal foraminiferal distributions: implications for sea-level studies. *Marine Micropaleontology*, 36, 205-223.

Horton, B.P., Edwards, R.J. & Lloyd, J.M. (2000) Implications of a microfossil transfer function in Holocene sea-level studies. In *Holocene land–ocean interaction and environmental change around the western North Sea, Special Publication,* (Eds, Shennan, I. & Andrews, J.E.) Geological Society, London, pp. 41-54.

Horton, B.P., Engelhart, S.E., Hill, D.F., Kemp, A.C., Nikitina, D., Miller, K.G. and Peltier, W.R. (2013) Influence of tidal-range change and sediment compaction on Holocene relative sea-level change in New Jersey, USA. *Journal of Quaternary Science*, 28, 403-411.

Horton, B.P. & Edwards, R.J. (2005) The application of local and regional transfer functions to the reconstruction of Holocene sea levels, north Norfolk, England. *The Holocene*, 15, 216-228.

Horton, B.P. & Edwards, R.J. (2006) Quantifying Holocene Sea Level Change Using Intertidal Foraminifera: Lessons from the British Isles. *Cushman Foundation for Foraminiferal Research, Special Publication*, 40, 95.

Horton, B. & Sawai, Y. (2010) Diatoms as indicators of former sea levels, earthquakes, tsunamis, and hurricanes. *The diatoms: applications for the Environmental and Earth Sciences*, 2, 357-372.

Hosfield, R. (2007) Terrestrial implications for the maritime geoarchaeological resource: A view from the Lower Palaeolithic. *Journal of Maritime Archaeology*, 2, 4-23.

Houlder (1994) The Stone Age. In *Cardiganshire County History Vol 1 – From the Earliest Times to the Coming of the Normans,* Cardiganshire Antiquatarian Society, Cardiganshire, pp. 107-123.

Housley, R.A. (1988) The environmental context of the Glastonbury lake village. *Somerset Levels Papers*, 14, 63-82.

Housley, R.A., Straker, V., Chambers, F.M. and Lageard, J.G. (2007) An Ecological Context for the Post-Roman Archaeology of the Somerset Moors (South West England, UK). *Journal of Wetland Archaeology*, 7, 1-22.

Howard, A.J., Gearey, B.R., Hill, T., Fletcher, W. and Marshall, P. (2009) Fluvial sediments, correlations and palaeoenvironmental reconstruction: the development of robust radiocarbon chronologies. *Journal of Archaeological Science*, 36, 2680-2688.

Hu, (2010) *Developing a quality-controlled postglacial sea-level database for coastal Louisiana to assess conflicting hypotheses of Gulf Coast sea-level change*. Tulane University.

Hubbard, A., Bradwell, T., Golledge, N., Hall, A., Patton, H., Sugden, D., Cooper, R. and Stoker, M. (2009) Dynamic cycles, ice streams and their impact on the extent, chronology and deglaciation of the British–Irish ice sheet. *Quaternary Science Reviews*, 28, 758-776.

Hughes, A.L.C., Clark, C.D. & Jordan, C.J. (2010) Subglacial bedforms of the last British Ice Sheet. *Journal of Maps*, 6, 543-563.

Hughes, A.L.C., Clark, C.D. & Jordan, C.J. (2014) Flow-pattern evolution of the last British Ice Sheet. *Quaternary Science Reviews*, 89, 148-168.

Hughes, P.D. (2002) Loch Lomond Stadial glaciers in the Aran and Arenig Mountains, North Wales, Great Britain. *Geological Journal*, 37, 9-15.

Hughes, P.D. (2009) Loch Lomond Stadial (Younger Dryas) glaciers and climate in Wales. *Geological Journal*, 44, 375-391.

Hughes, P.D.M. & Dumayne-Peaty, L. (2002) Testing Theories of Mire Development Using Multiple Successions at Crymlyn Bog, West Glamorgan, South Wales, UK. *Journal of Ecology*, 90, 456-471.

Hughes, P.D., Glasser, N.F. & Fink, D. (2016) Rapid thinning of the Welsh Ice Cap at 20–19ka based on 10Be ages. *Quaternary Research*, 85, 107-117.

Hustedt, F. (1953) Die Systematik der Diatomeen in ihren Bezeihungen zur Geologic und Okologie nebst einer Revision des Halobien-systems. *Svensk Botanisk Tidskrift*, 47, 509-519.

Hustedt, F. (1957) Die Diatomeenflora des Fluss-systems der Weser im Gebeit der Hansestadt Bremen. *Abhandlungen Naturwissenschaftlichen Verein, Bremen*, 34, 181-440.

Hutton, J. (1795) *Theory of the earth: With proofs and illustrations*. Library of Alexandria,

Ince, J. (1983) Two postglacial pollen profiles from the uplands of Snowdonia, Gwynedd, North Wales. *New Phytologist*, 95, 159-172.

Jackson, S.T. & Williams, J.W. (2004) Modern analogs in Quaternary paleoecology: here today, gone yesterday, gone tomorrow? *Annual Review of Earth and Planetary Sciences*, 32,

Jansson, K.N. & Glasser, N.F. (2005) Palaeoglaciology of the Welsh sector of the British–Irish Ice Sheet. *Journal of the Geological Society*, 162, 25-37.

Jennings, S., Orford, J.D., Canti, M., Devoy, R.J.N. and Straker, V. (1998) The role of relative sea-level rise and changing sediment supply on Holocene gravel barrier development: the example of Porlock, Somerset, UK. *The Holocene*, 8, 165-181.

Jones, A.M. (2016) *Preserved in the peat: an extraordinary Bronze Age burial on Whitehorse Hill, Dartmoor, and its wider context*. Oxbow Books, Oxford; Philadelphia.

Jones, J., Tinsley, H., McDonnell, R., Cameron, N., Haslett, S. and Smith, D. (2005) Mid Holocene coastal environments from Minehead Beach, Somerset, UK. *Archaeology in the Severn Estuary 2004*, 15, 49-69.

Jordan, J.T., Smith, D.E., Dawson, S. and Dawson, A.G., (2010) Holocene relative sea-level changes in Harris, Outer Hebrides, Scotland, UK. *Journal of Quaternary Science*, *25*(2), pp.115-134.

Jowsey, P.C. (1966) An improved peat sampler. *New Phytologist,* 65, 245–248.

Juggins, S. (2003) C2 user guide: Software for ecological and palaeoecological data analysis and visualization. *University of Newcastle, Newcastle upon Tyne, UK*, 69.

Juggins, S. & Ter Braak, C.J.F. (1993) CALIBRATE-unpublished computer program. *Environmental Change Research Centre, University College, London*,

Juggins, S. & Birks, H.J.B. (2012) Quantitative environmental reconstructions from biological data. In *Tracking environmental change using lake sediments,* Springer, pp. 431-494.

Keeping, W. (1878) II.—Notes on the Geology of the Neighbourhood of Aberystwyth. *Geological Magazine*, 5, 532-547.

Kemp, A.C., Horton, B.P., Corbett, D.R., Culver, S.J., Edwards, R.J. and van de Plassche, O. (2009) The relative utility of foraminifera and diatoms for reconstructing late Holocene sea-level change in North Carolina, USA. *Quaternary Research*, 71, 9-21.

Kemp, A.C., Horton, B.P., Vann, D.R., Engelhart, S.E., Pre, C.A.G., Vane, C.H., Nikitina, D. and Anisfeld, S.C. (2012) Quantitative vertical zonation of salt-marsh foraminifera for reconstructing former sea level; an example from New Jersey, USA. *Quaternary Science Reviews*, 54, 26-39.

Kemp, A.C., Telford, R.J., Horton, B.P., Anisfeld, S.C. and Sommerfield, C.K. (2013) Reconstructing Holocene sea level using salt-marsh foraminifera and transfer functions: lessons from New Jersey, USA. *Journal of Quaternary Science*, 28, 617-629.

Kemp, A.C., Hawkes, A.D., Donnelly, J.P., Vane, C.H., Horton, B.P., Hill, T.D., Anisfeld, S.C., Parnell, A.C. and Cahill, N. (2015) Relative sea-level change in Connecticut (USA) during the last 2200 yrs. *Earth and Planetary Science Letters*, 428, 217-229.

Kemp, A.C. and Telford, R.J. (2015). Transfer functions. *Handbook of sea-level research*, pp.470-499.

Khan, N.S., Ashe, E., Shaw, T.A., Vacchi, M., Walker, J., Peltier, W.R., Kopp, R.E. and Horton, B.P. (2015) Holocene relative sea-level changes from near-, intermediate-, and far-field locations. *Current Climate Change Reports*, 1, 247.

Kidson, C. & Heyworth, A. (1972) The Flandrian sea-level rise in the Bristol Channel. *Proceedings of the Ussher Society*, 2, 565-584.

Kidson, C. & Heyworth, A. (1978) Holocene eustatic sea level change, *Nature*, 273(5665), pp.748-750.

Kitely (1975) *Recent and Postglacial Foraminifera in Cores from Cardigan Bay*. Aberystwyth.

Knudson, B.M. (1954) The Ecology of the Diatom Genus Tabellaria in the English Lake District. *Journal of Ecology*, 42, 345-358.

Kociolek, J.P. (2018) A worldwide listing and biogeography of freshwater diatom genera: a phylogenetic perspective. *Diatom Research*, 33, 509-534.

Krammer, K. & Lange-Bertalot, H. (1991) Bacillariophyceae, Teil 3. centrales, fragilariaceae, eunotiaceae, achnanthaceae. *Süßwasserflora von Mitteleuropa; Pascher, A., Ettl, H., Gerloff, J., Heynig, H., Mollenhauer, D., Eds*, 576.

Proudman Oceanographic Laboratory (2018) Highest & lowest predicted tides at Avonmouth, National Tidal and Sea Level Facility. Available at: http://www.ntslf.org/tides/hilo?port=Avonmouth (Accessed: 4 May 2018).

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

Lambeck, K. (1996) Sea-level change and shore-line evolution in Aegean Greece since Upper Palaeolithic time. *Antiquity*, 70, 588-611.

Lambeck, K. (1993) Glacial rebound of the British Isles—II. A high-resolution, high-precision model. *Geophysical Journal International*, 115, 960-990.

Lambeck, K. (1996) Glaciation and sea-level change for Ireland and the Irish Sea since Late Devensian/Midlandian time. *Journal of the Geological Society*, 153, 853-872.

Lambeck, K., Rouby, H. & Purcell, A. (2014) Sea level and global ice volumes from the Last Glacial Maximum to the Holocene. *Proceedings of the Natural Academy of Science*, 111, 15296-15303.

Larcombe, P. & Jago, C.F. (1994) The late devensian and holocene evolution of Barmouth Bay, Wales. *Sedimentary Geology*, 89, 163-180.

Lawrence, T. (2010) Holocene sea-level changes in the Bristol Channel - Evidence from Porlock, Somerset, UK. *The Plymouth Student Scientist*, 3, 233-280.

Leipe, C., Kobe, F. & Müller, S. (2019) Testing the performance of sodium polytungstate and lithium heteropolytungstate as non-toxic dense media for pollen extraction from lake and peat sediment samples. *Quaternary International*, 516, 207-214.

Leng & Pratt (1987) Survey South of Penllyn Farm. *BGS open file report SN59NE/1-23*,

Leorri, E. & Cearreta, A. (2009) Recent sea-level changes in the southern Bay of Biscay: transfer function reconstructions from salt-marshes compared with instrumental data. *Scientia Marina*, 73, 287-296.

Leorri, E., Horton, B.P. & Cearreta, A. (2008) Development of a foraminifera-based transfer function in the Basque marshes, N. Spain: Implications for sea-level studies in the Bay of Biscay. *Marine Geology*, 251, 60-74.

Leorri, E., Fatela, F., Cearreta, A., Moreno, J., Antunes, C. and Drago, T. (2011) Assessing the performance of a foraminifera-based transfer function to estimate sea-level changes in northern Portugal. *Quaternary Research*, 75, 278-287.

Libby, W.F. (1952) Chicago radiocarbon dates, III. *Science*, 116, 673-681.

Libby, W.F. (1955) *Radiocarbon dating*. Chicago University Press, Chicago.

Liu, Y., Wang, Q. & Fu, C. (2011) Taxonomy and distribution of diatoms in the genus Eunotia from the Da'erbin Lake and Surrounding Bogs in the Great Xing'an Mountains, China. *Nova Hedwigia*, 92, 205-232.

Long, A.J., Innes, J.B., Kirby, J.R., Lloyd, J.M., Rutherford, M.M., Shennan, I. and Tooley, M.J. (1998) Holocene sea-level change and coastal evolution in the Humber Estuary, eastern England: an assessment of rapid coastal change. *The Holocene*, 8, 229-247.

Long, A.J., Scaife, R.G. & Edwards, R.J. (1999) Pine pollen in intertidal sediments from Poole Harbour, UK; implications for late-Holocene sediment accretion rates and sea-level rise. *Quaternary International*, 55, 3-16.

Long, A.J., Waller, M.P. & Stupples, P. (2006) Driving mechanisms of coastal change: Peat compaction and the destruction of late Holocene coastal wetlands. *Marine Geology*, 225, 63-84.

Long, A.J., Woodroffe, S.A., Milne, G.A., Bryant, C.L. and Wake, L.M. (2010) Relative sea level change in west Greenland during the last millennium. *Quaternary Science Reviews*, 29, 367-383.

Lotter, A.F. (1991) Absolute dating of the late-glacial period in Switzerland using annually laminated sediments. *Quaternary Research*, 35, 321-330.

Lotter, A.F. (1999) Late-glacial and Holocene vegetation history and dynamics as shown by pollen and plant macrofossil analyses in annually laminated sediments from Soppensee, central Switzerland. *Vegetation History and Archaeobotany*, 8, 165-184.

Lowe, J.J. & Walker, M.J.C. (2014) *Reconstructing quaternary environments*. Routledge,

Lowe, S. (1994) *The Devensian Lateglacial and early Flandrian stratigraphy of Southern Snowdonia, North Wales,* (Doctoral dissertation, Queen Mary University of London).

Lukas, S. & Bradwell, T. (2010) Reconstruction of a Lateglacial (Younger Dryas) mountain ice field in Sutherland, northwestern Scotland, and its palaeoclimatic implications. *Journal of Quaternary Science*, 25, 567-580.

Massey, A.C., Gehrels, W.R., Charman, D.J. and White, S.V. (2006a) An intertidal foraminifera-based transfer function for reconstructing Holocene sea-level change in southwest England. *Journal of Foraminiferal Research*, 36, 215-232.

Massey, A.C., Paul, M.A., Gehrels, W.R. and Charman, D.J. (2006b) Autocompaction in Holocene coastal back-barrier sediments from south Devon, southwest England, UK. *Marine Geology*, 226, 225-241.

McCarroll, D. & Ballantyne, C.K. (2000) The last ice sheet in Snowdonia. *Journal of Quaternary Science*, 15, 765-778.

McDougall, D. (2013) Glaciation style and the geomorphological record: evidence for Younger Dryas glaciers in the eastern Lake District, northwest England. *Quaternary Science Reviews*, 73, 48-58.

Metcalfe, S.E., O'Hara, S.L., Caballero, M. and Davies, S.J. (2000) Records of Late Pleistocene–Holocene climatic change in Mexico—a review. *Quaternary Science Reviews*, 19, 699-721.

Mighall, T.M. & Chambers, F.M. (1995) Holocene vegetation history and human impact at Bryn y Castell, Snowdonia, north Wales. *New Phytologist*, 130, 299-321.

Milker, Y., Weinkauf, M.F., Titschack, J., Freiwald, A., Krüger, S., Jorissen, F.J. and Schmiedl, G. (2017) Testing the applicability of a benthic foraminiferal-based transfer function for the reconstruction of paleowater depth changes in Rhodes (Greece) during the early Pleistocene. *PloS one*, 12, e0188447.

Mills, H., Kirby, J., Holgate, S. and Plater, A. (2013) The Distribution of Contemporary Saltmarsh Foraminifera in a Macrotidal Estuary: an Assessment of Their Viability for Sea-Level Studies. *Journal of Ecosystem and Ecography*, 3, 131.

Milne, G.A. & Shennan, I. (2007) Sea Level Studies: Isostasy. In *Encyclopedia of Quaternary Sciences,* (Ed, Elias, S.) Elsevier, pp. 3043-3051.

Milne, G.A. & Mitrovica, J.X. (2008) Searching for eustasy in deglacial sea-level histories. *Quaternary Science Reviews*, 27, 2292-2302.

Mitrovica, J.X. & Peltier, W.R. (1993a) A new formalism for inferring mantle viscosity based on estimates of post glacial decay Times: A lication to RSL variations in N.E. Hudson Bay. *Geophysics Res Letters*, 20, 2183-2186.

Mitrovica, J.X. & Peltier, W.R. (1991) On postglacial geoid subsidence over the equatorial oceans. *Journal of Geophysical Research: Solid Earth*, 96, 20053-20071.

Mitrovica, J.X. & Peltier, W.R. (1993b) The inference of mantle viscosity from an inversion of the Fennoscandian relaxation spectrum. *Geophysical Journal International*, 114, 45-62.

Mitrovica, J.X. & Milne, G.A. (2002) On the origin of late Holocene sea-level highstands within equatorial ocean basins. *Quaternary Science Reviews*, 21, 2179-2190.

Mix, A.C., Bard, E. & Schneider, R. (2001) Environmental processes of the ice age: land, oceans, glaciers (EPILOG). *Quaternary Science Reviews*, 20, 627-657.

Mook, D.H. & Hoskin, C.M. (1982) Organic determinations by ignition: caution advised. *Estuarine, Coastal and Shelf Science*, 15, 697-699.

Moore, P.D., Webb, J.A. & Collinson, M.E. (1991) *Pollen analysis*. Blackwell Scientific Publications, Oxford.

Muller, R.A. (1977) Radioisotope Dating with a Cyclotron: The sensitivity of radioisotope dating is improved by counting atoms rather than decays. *Science*, 196, 489-494.

Munsterman, D. & Kerstholt, S. (1996) Sodium polytungstate, a new non-toxic alternative to bromoform in heavy liquid separation. *Review of Palaeobotany and Palynology*, 91, 417-422.

Murray-Wallace, C.V. (2007) Eustatic Sea-Level Changes Since the Last Glaciation. In *Encyclopedia of Quaternary Sciences,* (Ed, Elias, S.) Elsevier, pp. 3034-3043.

Murray-Wallace, C.V. (2013) Sea level studies: eustatic sea-level changes, glacial-interglacial cycles. In *Encyclopedia of Quaternary science,* (Ed, Elias, S.) pp. 427-438.

Murray, J.W. (1971) Living foraminiferids of tidal marshes: a review. *Journal of Foraminiferal Research*, 153-161.

Murray, J.W. (1979) *British Nearshore Foraminiferids: Keys and Notes for the Identification of the Species*. Linnean Society of London and the Estuarine and Brackish-water Sciences Association,

Murray, J.W. (2000) The enigma of the continued use of total assemblages in ecological studies of benthic foraminifera. *Journal of Foraminiferal Research*, 30, 244-245.

Murray, J.W. & Hawkins, A.B. (1976) Sediment transport in the Severn Estuary during the past 8000–9000 years. *Journal of the Geological Society*, 132, 385-398.

Murton, D.K. & Murton, J.B. (2012) Middle and Late Pleistocene glacial lakes of lowland Britain and the southern North Sea Basin. *Quaternary International*, 260, 115-142.

Nakagawa, T., Brugiapaglia, E., Digerfeldt, G., Reille, M., Beaulieu, J.L.D. and Yasuda, Y. (1998) Dense-media separation as a more efficient pollen extraction method for use with organic sediment/deposit samples: comparison with the conventional method. *Boreas*, 27, 15-24.

Neumann, B.S. (1967) Thermal techniques. In *Physical methods in determinative mineralogy,* Academic Press, London,

Nilsson, M., Klarqvist, M., Bohlin, E. and Possnert, G. (2001) Variation in 14C age of macrofossils and different fractions of minute peat samples dated by AMS. *The Holocene*, 11, 579-586.

Noga, T., Kochman, N., Peszek, Ł., Stanek-Tarkowska, J. and Pajączek, A. (2014) Diatoms (Bacillariophyceae) in rivers and streams and on cultivated soils of the Podkarpacie Region in the years 2007–2011. *Journal of Ecological Engineering*, 15, 6-25.

O'Connell, M. (1987) Early cereal-type pollen records from Connemara, western Ireland and their possible significance. *Pollen et spores*,

Overpeck, J.T., Webb, T.I.I.I. & Prentice, I.C. (1985) Quantitative interpretation of fossil pollen spectra: dissimilarity coefficients and the method of modern analogs. *Quaternary Research*, 23, 87-108.

Palmer, A.J.M. & Abbott, W.H. (1986) Diatoms as indicators of sea-level change. In *Sea-Level Research,* Springer, pp. 457-487.

Patton, H. & Hambrey, M.J. (2009) Ice-marginal sedimentation associated with the Late Devensian Welsh Ice Cap and the Irish Sea Ice Stream: Tonfanau, West Wales. *Proceedings of the Geologists' Association*, 120, 256-274.

Patton, H., Hubbard, A., Bradwell, T., Glasser, N.F., Hambrey, M.J. and Clark, C.D. (2013a) Rapid marine deglaciation: asynchronous retreat dynamics between the Irish Sea Ice Stream and terrestrial outlet glaciers. *Earth Surface Dynamics*, 1, 53.

Patton, H., Hubbard, A., Glasser, N.F., Bradwell, T. and Golledge, N.R. (2013b) The last Welsh Ice Cap: Part 2 – Dynamics of a topographically controlled icecap. *Boreas*, 42, 491-510.

Paul, A.M. & Little, A.J. (1991) Geotechnical properties of glacial deposits in lowland Britain. *Geotechnical properties of glacial deposits in lowland Britain*, 389-404.

Paul, M.A. (1983) The supraglacial land system. *Glacial Geology*, 71-90.

Paul, M.A. & Barras, B.F. (1998) A geotechnical correction for post-depositional sediment compression: examples from the Forth valley, Scotland. *Journal of Quaternary Science: Published for the Quaternary Research Association*, 13, 171-176.

Peck, A. (2018) An investigation into Late Holocene relative sea level change of the Dovey Estuary, Wales: A multi-proxy based analysis, MSc thesis.

Peltier, W.R. (1994) Ice Age Paleotopography. *Science*, 265, 195-201.

Peltier, W.R. (2002) On eustatic sea level history: Last Glacial Maximum to Holocene. *Quaternary Science Review*, 21, 377-396.

Peltier, W.R. (2004) Global glacial isostasy and the surface of the ice-age earth: The ICE-5G (VM2) model and GRACE. *Annual Review Earth Planet Science*, 32, 111-149.

Peltier, W.R. & Fairbanks, R.G. (2006) Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. *Quaternary science reviews*, 25, 3322-3337.

Phillips, E., Everest, J. & Diaz-Doce, D. (2010) Bedrock controls on subglacial landform distribution and geomorphological processes: Evidence from the Late Devensian Irish Sea Ice Stream. *Sedimentary Geology*, 232, 98-118.

Phillips, E., Lee, J.R., Riding, J.B., Kendall, R. and Hughes, L. (2013) Periglacial disruption and subsequent glacitectonic deformation of bedrock: an example from Anglesey, North Wales, UK. *Proceedings of the Geologists' Association*, 124, 802-817.

Phillips, F.M. (1994) Surface Exposure Dating of Glacial Features in Great Britain Using Cosmogenic Chlorine-36: Preliminary Results. *Mineralogical Magazine*, 58A, 722-723.

Phipps, D. & Playford, G. (1984) Laboratory techniques for extraction of palynomorphs from sediments.

Pizzuto, J.E. & Schwendt, A.E. (1997) Mathematical modeling of autocompaction of a Holocene transgressive valley-fill deposit, Wolfe Glade, Delaware. *Geology*, 25, 57-60.

Pratt, W.T., Woodhall, D.G. and Howells, M.F. (1995) *Geology of the country around Cadair Idris*. HM Stationery Office,

Preuss, H. (1979) Progress in computer evaluation of sea level data within the IGCP Project no. 61. 104-134.

Prince, H.E. (1988) *Late-glacial and Post-glacial sea-level movements in North Wales with particular reference to the techniques for the analysis and interpretation of unconsolidated estuarine sediments*. PhD thesis, University of Wales, Aberystwyth.

Reimer, P.J., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Ramsey, C.B., Buck, C.E., Cheng, H., Edwards, R.L., Friedrich, M. and Grootes, P.M. (2013) IntCal13 and Marine13 radiocarbon age calibration curves 0–50,000 years cal BP. *Radiocarbon*, 55, 1869-1887.

Ridgway, J., Andrews, J.E., Ellis, S., Horton, B.P., Innes, J.B., Knox, R.O.B., McArthur, J.J., Maher, B.A., Metcalfe, S.E., Mitlehner, A. and Parkes, A. (2000) Analysis and interpretation of Holocene sedimentary sequences in the Humber Estuary. *Geological Society, London, Special Publications*, 166, 9-39.

Roberts, M.J., Scourse, J.D., Bennell, J.D., Huws, D.G., Jago, C.F. and Long, B.T. (2011) Late Devensian and Holocene relative sea-level change in North Wales, UK. *Journal of Quaternary Science*, 26, 141-155.

Robinson, G. (2013) The excavation of a multi period rock-shelter at Garreg Hylldrem, Llanfrothen 2011-2012. *Archaeology in Wales*, 52, 3-10.

Roe, H.M., Doherty, C.T., Patterson, R.T. and Swindles, G.T. (2009) Contemporary distributions of saltmarsh diatoms in the Seymour–Belize Inlet Complex, British Columbia, Canada: Implications for studies of sea-level change. *Marine Micropaleontology*, 70, 134-150.

Rossi, V., Horton, B.P., Corbett, D.R., Leorri, E., Perez-Belmonte, L. and Douglas, B.C. (2011) The application of foraminifera to reconstruct the rate of 20th century sea level rise, Morbihan Golfe, Brittany, France. *Quaternary Research*, 75, 24-35.

Round, F.E., Crawford, R.M. & Mann, D.G. (1990) *The Diatoms: Biology and Morphology of the Genera*. Cambridge University Press, Cambridge.

Rovere, A., Stocchi, P. & Vacchi, M. (2016) Eustatic and Relative Sea Level Changes. *Current Climate Change Reports*, 1-11.

RSPB (2014). 'Machair Grassland Landscape Scale Conservation Project'. Accessed 8 April 2023. https://www.rspb.org.uk/our-work/conservation/landscape-scale-conservation/sites/machair-grassland/.

Rushby, G.T., Richards, G.T., Gehrels, W.R., Anderson Jr, W.P., Bateman, M.D. and Blake, W.H. (2019) Testing the mid-Holocene relative sea-level highstand hypothesis in North Wales, UK. *The Holocene*, 29, 1491-1502.

Rybczyk, J.M., Callaway, J.C. & Day Jr, J.W. (1998) A relative elevation model for a subsiding coastal forested wetland receiving wastewater effluent. *Ecological Modelling*, 112, 23-44.

Sahlin, E.A., Glasser, N.F., Jansson, K.N. and Hambrey, M.J. (2009) Connectivity analyses of valley patterns indicate preservation of a preglacial fluvial valley system in the Dyfi basin, Wales. *Proceedings of the Geologists' Association*, 120, 245-255.

Sambrook, R.P. & Williams, G. (1996) *Cardigan Bay Coastal Survey.* Dyfed Archaeological Trust, Rep. No. PRN 30751.

Santisteban, J.I., Mediavilla, R., Lopez-Pamo, E., Dabrio, C.J., Zapata, M.B.R., García, M.J.G., Castano, S. and Martínez-Alfaro, P.E. (2004) Loss on ignition: a qualitative or quantitative method for organic matter and carbonate mineral content in sediments? *Journal of Paleolimnology*, 32, 287-299.

Sawai, Y. (2001) Distribution of living and dead diatoms in tidal wetlands of northern Japan: relations to taphonomy. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 173, 125-141.

Sawai, Y., Horton, B.P. & Nagumo, T. (2004) The development of a diatom-based transfer function along the Pacific coast of eastern Hokkaido, northern Japan—an aid in paleoseismic studies of the Kuril subduction zone. *Quaternary Science Reviews*, 23, 2467-2483.

Scaife, R. & Long, A. (1995) Evidence for Holocene sea-level changes at Caldicot Pill. *Archaeology in the Severn Estuary*, 8, 81-86.

Scheder, J., Frenzel, P., Bungenstock, F., Engel, M., Brueckner, H. and Anna, P.I.N.T. (2019) Vertical and lateral distribution of Foraminifera and Ostracoda in the East Frisian Wadden Sea – developing a transfer function for relative sea-level change. *Geologica Belgica*,

Scott-Jackson, J.E. & Walkington, H. (2005) Methodological issues raised by laser particle size analysis of deposits mapped as Clay-with-flints from the Palaeolithic site of Dickett's Field, Yarnhams Farm, Hampshire, UK. *Journal of Archaeological Science*, 32, 969-980.

Scott, D.S. & Medioli, F.S. (1978) Vertical zonations of marsh foraminifera as accurate indicators of former sea-levels. *Nature*, 272, 528-531.

Scott, D.BS & Medioli, F.S. (1980) Living vs. total foraminiferal populations: their relative usefulness in paleoecology. *Journal of Paleontology*, 814-831.

Scourse, J.D. & Furze, M.F.A. (2001) A critical review of the glaciomarine model for Irish Sea deglaciation: evidence from southern Britain, the Celtic shelf and adjacent continental slope. *Journal of Quaternary Science: Published for the Quaternary Research Association*, 16, 419-434.

Sejrup, H.P., Haflidason, H., Aarseth, I., King, E., Forsberg, C.F., Long, D. and Rokoengen, K., (1994) Late Weichselian glaciation history of the northern North Sea. *Boreas*, 23, 1-13.

Shennan, I. (1982a) Interpretation of Flandrian sea-level data from the Fenland, England. *Proceedings of the Geologists' Association*, 93, 53-63.

Shennan, I. (1982b) Problems of correlating Flandrian sea-level changes and climate. *Climatic change in later prehistory*, 52-67.

Shennan, I. (1983) Flandrian and Late Devensian sea-level changes and crustal movements in England and Wales. In *Shorelines and isostasy,* Academic Press London, pp. 255-283.

Shennan, I. (1986) Flandrian sea-level changes in the Fenland. II: Tendencies of sea-level movement, altitudinal changes, and local and regional factors. *Journal of Quaternary Science*, 1, 155-179.

Shennan, I. (2015) Chapter 2 Handbook of sea-level research: framing research questions. In *Handbook of Sea-Level Research,* (Eds, Shennan, I., Long, A.J. & Horton, B.P.) John Wiley & Sons Ltd, pp. 3-25.

Shennan, I., Lambeck, K., Horton, B., Innes, J., Lloyd, J., McArthur, J. and Rutherford, M. (2000) Holocene isostasy and relative sea-level changes on the east coast of England. In *Holocene Land-Ocean Interaction and Environmental Change around the North Sea,* (Eds, Shennan, I. & Andrews, J.E.) Geological Society, Special Publications, London, pp. 275-298.

Shennan, I., Bradley, S., Milne, G., Brooks, A., Bassett, S. and Hamilton, S. (2006) Relative sea-level changes, glacial isostatic modelling and ice-sheet reconstructions from the British Isles since the Last Glacial Maximum. *Journal of Quaternary Science*, 21, 585-599.

Shennan, I., Milne, G. & Bradley, S. (2009) Late Holocene relative land-and sea-level changes: providing information for stakeholders. *GSa today*, 19, 52-53.

Shennan, I., Milne, G.A. & Bradley, S. (2012) Late Holocene vertical land motion and relative sea-level changes: lessons from the British Isles. *Journal of Quaternary Science*, 27, 64-70.

Shennan, I., Long, A.J. & Horton, B.P. (2015) *Handbook of Sea-Level Research*. John Wiley & Sons,

Shennan, I., Bradley, S.L. & Edwards, R. (2018) Relative sea-level changes and crustal movements in Britain and Ireland since the Last Glacial Maximum. *Quaternary Science Reviews*, 188, 143-159.

Sheridan, A. (2013) Early Neolithic Habitation Structures in Britain and Ireland: a Matter of Circumstance and Context. In *Tracking the Neolithic House in Europe: Sedentism, Architecture and Practice,* (Eds, Hofmann, D. & Smyth, J.) Springer, New York, NY, pp. 283-300.

Sherrod, B.L., Rollins, H.B. & Kennedy, S.K. (1989) Subrecent intertidal diatoms from St Catherines Island, Georgia: Taphonomic implications. *Journal of Coastal Research*, 5, 665-677.

Shi, G.R. (1993) Multivariate data analysis in palaeoecology and palaeobiogeography—a review. *Palaeogeography, palaeoclimatology, palaeoecology*, 105, 199-234.

Shore, J.S., Bartley, D.D. & Harkness, D.D. (1995) Problems encountered with the 14C dating of peat. *Quaternary Science Reviews*, 14, 373-383.

Simmons, I.G. & Innes, J.B. (1987) Mid-holocene adaptations and later Mesolithic forest disturbance in Northern England. *Journal of Archaeological Science*, 14, 385-403.

Simpkins, K. (1974) The late-glacial deposits at Glanllynnau, Caernarvonshire. *New Phytologist*, 73, 605-618.

Smith, A.G. (1984) *Newferry and the Boreal-Atlantic transition*, New Phytologist, Wiley Online Library.

Smith, A.G. & Morgan, L.A. (1989) A succession to ombrotrophic bog in the Gwent Levels, and its demise: a Welsh parallel to the peats of the Somerset Levels. *New Phytologist*, 112, 145-167.

Smith, D.E., Davies, M.H., Brooks, C.L., Mighall, T.M., Dawson, S., Rea, B.R., Jordan, J.T. and Holloway, L.K., (2010) Holocene relative sea levels and related prehistoric activity in the Forth lowland, Scotland, United Kingdom. Quaternary Science Reviews, 29(17-18), pp.2382-2410.

Smith, D.E., Barlow, N.L., Bradley, S.L., Firth, C.R., Jordan, J.T. and David, Long, (2019) Quaternary sea level change in Scotland. Earth and Environmental Science Transactions of the Royal Society of Edinburgh, 110(1-2), pp.219-256.

Smith, D.E., Tipping, R.M., Jordan, J.T. and Blackett, M. (2020) Holocene coastal change at Luce Bay, South West Scotland. *Journal of Quaternary Science*, 35, 743-759.

Smith, G. (2000) Prehistoric funerary and ritual monuments survey: Meirionnydd. *Archaeology in Wales*, 40, 60.

Smith, G. (2001) Prehistoric Funerary and Ritual Sites Survey: Meirionnydd. *Unpublished report: Gwynedd Archaeological Trust report*,

Smith, G. (2004) Aberdyfi to Dysynni Flood Alleviation Scheme - Penllyn Marshes, Tywyn.

Smith, G. (2005) Tywyn Coastal Protection Cheme Archaeological Assessment.

Smith, G., Davidson, A. & Kenney, J. (2002) North Wales Intertidal Peat Survey 2001-2002. *G1679*, 30-33.

Smith, J.B., Smith, L.B. & Society, M.H.A.R. (2001) *A history of Merioneth. v2. The Middle Ages*. Published on behalf of the Merioneth Historical and Record Society by the University of Wales Press, Cardiff.

Stein, J.K. (1984) Organic matter and carbonates in archaeological sites. *Journal of Field Archaeology*, 11, 239-246.

Stephenson, E.J., Mast, T.S. & Muller, R.A. (1979) Radiocarbon dating with a cyclotron. *Nuclear Instruments and Methods*, 158, 571-577.

Stockmarr, J., Stockmarr, J. & Stockmarr, S. (1971) Tablets with spores used in absolute pollen analysis.

Stoermer, E.F. & Smol, J.P. (1990) Applications and uses of diatoms: prologue. In *The Diatoms: Applications for the Environmental and Earth Sciences,* (Eds, Stoermer, E.F. & Smol, J.P.) Cambridge University Press, Cambridge, pp. 3-8.

Streif, H. (1979) Cyclic formation of coastal deposits and their indications of vertical sea-level changes. *Oceanis*, 5, 303-306.

Sturt, F., Dix, J.K., Grant, M.J., Steadman, S., Scaife, R., Edwards, R., Griffiths, S., Cameron, N., Thompson, C., Bray, S. and Jones, J. (2014) Life below the waves: palaeolandscapes preserved within the sub-tidal Bristol Channel. *Archaeology in the Severn Estuary*, 22, 41-66.

Suess, H.E. (1970) Bristle-cone pine calibration of the radiocarbon time-scale, 5200 BC To the present. In *Unknown,* pp. 303-311.

Sundberg, R. (1985) When is the inverse regression estimator MSE-superior to the standard regression estimator in multivariate controlled calibration situations? *Statistics & probability letters*, 3, 75-79.

Syvitski, J.P.M. (1991) *Principles, methods, and application of particle size analysis*. Cambridge University Press Cambridge,

Szkornik, K., Gehrels, W.R. & Kirby, J.R. (2006) Salt-marsh diatom distributions in Ho Bugt (western Denmark) and the development of a transfer function for reconstructing Holocene sea-level changes. *Marine Geology*, 235, 137-150.

Tallantire, P. (1974) Paleohistory of Grey Alder (alnus-Incana (l) Moench) and Black Alder (a-Glutinosa (l) Gaertn) Infennoscandia. *New Phytologist*, 73, 529-546.

Tappin (1994) The Geology of Cardigan Bay and the Bristol Channel (no. 8). *HM Stationery Office*,

Taylor, J.A. (1973) Chronometers and chronicles. A study of palaeo-environments in west central Wales. *Progress in Geography*, 2, 250-334.

Telford, R.J. & Birks, H.J.B. (2005) The secret assumption of transfer functions: problems with spatial autocorrelation in evaluating model performance. *Quaternary Science Reviews*, 24, 2173-2179.

Telford, R.J. & Birks, H.J.B. (2009) Evaluation of transfer functions in spatially structured environments. *Quaternary Science Reviews*, 28, 1309-1316.

Ten Hove, H.A. (1968) The Ulmus fall at the transition atlanticum-subboreal in pollen diagrams. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 5, 359-369.

Ter Braak, C.J.F. (1995) Non-linear methods for multivariate statistical calibration and their use in palaeoecology: a comparison of inverse (k-nearest neighbours, partial least squares and weighted averaging partial least squares) and classical approaches. *Chemometrics and Intelligent Laboratory Systems*, 28, 165-180.

Ter Braak, C.J.F. & Prentice, I.C. (1988) *A Theory of Gradient Analysis. vol. 18 of Advances in Ecological Research*. Academic Press, Ter Braak, C.J.F. & Juggins, S. (1993) Weighted averaging partial least squares regression (WA-PLS): an improved method for reconstructing environmental variables from species assemblages. 485-502.

Thomas, G.S.P. (1985) The Late Devensian glaciation along the border of northeast Wales. *Geological Journal*, 20, 319-340.

Thomas, G.S.P. (1989) The Late Devensian glaciation along the western margin of the Cheshire-Shroshire Iowland. *Journal of Quaternary Science*, 4, 167-181.

Thomas, G.S.P., Chester, D.K. & Crimes, P. (1998) The Late Devensian glaciation of the eastern Lleyn Peninsula, North Wales: evidence for terrestrial depositional environments. *Journal of Quaternary Science*, 13, 255-270.

Thomas, G.S.P. & Chiverrell, R.C. (2007) Structural and depositional evidence for repeated ice-marginal oscillation along the eastern margin of the Late Devensian Irish Sea Ice Stream. *Quaternary Science Reviews*, 26, 2375-2405.

Tipping, R. (1993) A detailed early postglacial (Flandrian) pollen diagram from Cwm Idwal, North Wales. *New Phytologist*,

Tipping, R. (2002) Climatic Variability and 'Marginal' Settlement in Upland British Landscapes: A Re-Evaluation. *Landscapes*, 3, 10-29.

Tooley, M.J. (1974) Sea-level changes during the last 9000 years in North-West England. *Geographical. Journal*, 140, 18-42.

Tooley, M.J. (1978) *Sea-level changes: north-west England during the Flandrian Stage*. Oxford University Press,

Tooley, M.J. (1982) Sea-level changes in northern England. *Proceedings of the Geologists' Association*, 93, 43-51.

Törnqvist, T.E., De Jong, A.F., Oosterbaan, W.A. and Van Der Borg, K. (1992) Accurate dating of organic deposits by AMS 14 C measurement of macrofossils. *Radiocarbon*, 34, 566-577.

Törnqvist, T.E., Wallace, D.J., Storms, J.E., Wallinga, J., Van Dam, R.L., Blaauw, M., Derksen, M.S., Klerks, C.J., Meijneken, C. and Snijders, E.M. (2008) Mississippi Delta subsidence primarily caused by compaction of Holocene strata. *Nature Geoscience*, 1, 173-176.

Törnqvist, T.E., Rosenheim, B.E., Hu, P. and Fernandez, A.B. (2015) Radiocarbon dating and calibration. *Handbook of Sea-Level Research, edited by: Shennan, I., Long, AJ, and Horton, BP*, 349-360.

Tovey, K.N. & Paul, M.A. (2002) Modelling self-weight consolidation in Holocene sediments. *Bulletin of Engineering Geology and the Environment*, 61, 21-33.

Tröels-Smith, J. (1955) Karakterisering af lose jordarter. Characterization of unconsolidated sediments.

Turney, C.S., Jones, R.T., Thomas, Z.A., Palmer, J.G. and Brown, D. (2016) Extreme wet conditions coincident with Bronze Age abandonment of upland areas in Britain. *Anthropocene*, 13, 69-79.

Turney, C.S., Coope, G.R., Harkness, D.D., Lowe, J.J. and Walker, M.J. (2000) Implications for the dating of Wisconsinan (Weichselian) Late-Glacial events of systematic radiocarbon age differences between terrestrial plant macrofossils from a site in SW Ireland. *Quaternary Research*, 53, 114-121.

Uehara, K., Scourse, J.D., Horsburgh, K.J., Lambeck, K. and Purcell, A.P. (2006) Tidal evolution of the northwest European shelf seas from the Last Glacial Maximum to the present. *Journal of Geophysical Research: Oceans*, 111, C09025.

UK Hydrological Office (2020) *Admiralty Tide Tables, NP201B*.

Van Dam, H., Mertens, A. & Sinkeldam, J. (1994) A coded checklist and ecological indicator values of freshwater diatoms from the Netherlands. *Netherland Journal of Aquatic Ecology*, 28, 117-133.

van de Plassche, O. (1982) Sea-level change and water-level movements in the Netherlands during the Holocene. *Ph. D. dissertation, Vrije Universiteit Amsterdam*,

Vermeulen, S., Lepoint, G. & Gobert, S. (2012) Processing samples of benthic marine diatoms from Mediterranean oligotrophic areas. *Journal of applied phycology*, 24, 1253-1260.

Vos, P.C. & De Wolf, H. (1988) Geologie en Diatomeeën. *Grondboor & Hamer*, 42, 57-68.

Vos, P.C. & de Wolf, H. (1993) Diatoms as a tool for reconstructing sedimentary environments in coastal wetlands; methodological aspects. *Hydrobiologia*, 269-7, 285-296.

Vos, P.C. & De Wolf, H. (1994) Palaeoenvironmental research on diatoms in early and middle Holocene deposits in central north Holland (The Netherlands). *Aquatic Ecology*, 28, 97-115.

Wachnicka, A., Collins, L.S. & Gaiser, E.E. (2013) Response of diatom assemblages to 130 years of environmental change in Florida Bay (USA). *Journal of paleolimnology*, 49, 83-101.

Waller, M.P. (1993) Flandrian vegetational history of southeastern England. Pollen data from Pannel Bridge, East Sussex. *New Phytologist*, 124, 345-369.

Waller, M.P., Long, A.J., Long, D. and Innes, J.B. (1999) Patterns and processes in the development of coastal mire vegetation: Multi-site investigations from Walland Marsh, Southeast England. *Quaternary Science Reviews*, 18, 1419-1444.

Waller, M.P. & Long, A.J. (2003) Holocene coastal evolution and sea-level change on the southern coast of England: a review. *Journal of Quaternary Science*, 18, 351-359.

Walter E. Dean, J. (1974) Determination of Carbonate and Organic Matter in Calcareous Sediments and Sedimentary Rocks by Loss on Ignition: Comparison With Other Methods. *Journal of Sedimentary Research*, 44,

Wanless. (1982) *Sea-level research: a manual for the collection and evaluation of data: A manual for the collection and evaluation of data*. Springer Netherlands,

Watcham, E.P., Shennan, I. & Barlow, N.L.M. (2013) Scale considerations in using diatoms as indicators of sea-level change: lessons from Alaska. *Journal of Quaternary Science*, 28, 165-179.

Watkins, R. (1991) *Postglacial Vegetational Dynamics in Lowland North Wales*. Bangor University (United Kingdom).

Watson, E. (1962) The Glacial Morphology of the Tal-y-llyn Valley, Merionethshire. *Transactions and Papers (Institute of British Geographers)*, 15-31.

Wendland, W.M. & Bryson, R.A. (1974) Dating climatic episodes of the Holocene. *Quaternary Research*, 4, 9-24.

Wentworth, C.K. (1922) A Scale of Grade and Class Terms for Clastic Sediments. *The Journal of Geology*, 30, 377-392.

White, J.C. & Beamish, D. (2014) A lithological assessment of the resistivity data acquired during the airborne geophysical survey of Anglesey, North Wales. *Proceedings of the Geologists' Association*, 125, 170-181.

Whitehouse, P. & Bradley, S. (2013) Sea level studies: eustatic sea-level changes, glacial-interglacial cycles. In *Encyclopedia of Quaternary science,* (Ed, Elias, S.)

Wilks, P. (1977) *Holocene sea-level change in Cardigan Bay*, PhD thesis, University of Wales.

Wilks, P.J. (1979) Mid-Holocene sea-level and sedimentation interactions in the Dovey estuary area, Wales. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 26, 17-36.

Williams, D.M. & Round, F.E. (1986) Revision of the genus Synedra Ehrenb. *Diatom research*, 1, 313-339.

Williams, H. (2003) Modeling shallow autocompaction in coastal marshes using cesium-137 fallout: Preliminary results from the Trinity River Estuary, Texas. *Journal of Coastal Research*, 180-188.

Wills, L.J. (1924) The Development of the Severn Valley in the Neighbourhood of Iron-Bridge and Bridgnorth. *Quarterly Journal of the Geological Society*, 80, 274-308.

Wilson, G.P. & Lamb, A.L. (2012) An assessment of the utility of regional diatom-based tidal-level transfer functions. *Journal of Quaternary Science*, 27, 360-370.

Woodbridge, J., Fyfe, R.M., Roberts, N., Downey, S., Edinborough, K. and Shennan, S. (2014) The impact of the Neolithic agricultural transition in Britain: a comparison of pollen-based land-cover and archaeological 14C date-inferred population change. *Journal of Archaeological Science*, 51, 216-224.

Woodroffe, S.A. (2009a) Recognising subtidal foraminifera: implications for quantitative sea-level reconstructions using foraminifera-based transfer function. *Journal of Quaternary Science*, 24, 215-223.

Woodroffe, S.A. (2009b) Testing models of mid to late Holocene sea-level change, North Queensland, Australia. *Quaternary Science Review*, 28, 2474-2478.

Woodroffe, S.A. & Long, A.J. (2009) Salt marshes as archives of recent relative sea-level hange in West Greenland. *Quaternary science reviews*, 28, 1750-1761.

Woodroffe, S.A. & Long, A.J. (2010) Reconstructing recent relative sea-level changes in West Greenland: Local diatom-based transfer functions are superior to regional models. *Quaternary International*, 221, 91-103.

Worsley, P. (1967) *Some aspects of the Quaternary evolution of the Cheshire Plain*. Manchester.

Worsley, P. (1970) The Cheshire–Shropshire lowlands. In *The Glaciations of Wales and adjoining regions,* Longmans, London, pp. 83-106.

Worsley, P. (2005) Glacial geology of the Condover area, south Shropshire. *Merican Geologist*, 16, 107-114.

Wright, A.J., Edwards, R.J. & van de Plassche, O. (2011) Reassessing transfer-function performance in sea-level reconstruction based on benthic salt-marsh foraminifera from the Atlantic coast of NE North America. *Marine Micropaleontology*, 81, 43-62.

Zong, Y. & Horton, B.P. (1999) Diatom-based tidal-level transfer functions as an aid in reconstructing Quaternary history of sea-level movements in the UK. *Journal of Quaternary Science*, 14, 153-167.

Zussman, J. (1967) *Physical methods in determinative mineralogy*. Academic Press, London.