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Late Devensian and Holocene Relative Sea-Level Change around Ireland

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Thesis submitted for the degree of Doctor of Philosophy.

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Summary

Late Devensian and Holocene Relative Sea-Level Change Around Ireland

Anthony Brooks Trinity College, Dublin October 2006

All records of Irish sea level since the Last Glacial Maximum (LGM) reflect the interplay of both glacio-isostatic and eustatic components with the isostatic signature being derived from the lithospheric response to local, regional and extra-regional ice cover. In addition to providing valuable information in their own right, relative sea level (RSL) data are powerful tools in the wider study of glaciation, climate change and Earth structure. Their intimate link with glacio isostatic adjustment (GIA) and eustatic sea level means that they contain within them information regarding former ice extent, thickness and melting history, as well as data pertaining to the physical structure of the Earth and its response to loading. This information can be distilled through the combination of RSL data and geophysical modelling.

This thesis can be defined by three main interrelated research objectives: Firstly, the compilation and examination of existing published RSL information and the tabulation of this information into a standardized sea-level database. Secondly, the collection of empirical data in the form of 'sea level index points' derived from sedimentary archives found along the west coast of Ireland. Finally, the development of a new model of the Irish Ice Sheet through the iterative procedure of glacial rebound modelling.

The sea-level database presented in this thesis is a first attempt at compiling a standardized sea-level database for Ireland, which is consistent with established protocols for the evaluation of sea-level information. The available data are of varied quality, yet limitations associated with this are mitigated by applying a rigorous appraisal of data errors and the consequent adoption of a quality tier system. Four quality tiers have been recognized in the database and together, these contain over 200 observations of sea level dating back to the late Devensian.

The entries contained within this newly developed database have been incorporated into geophysical modelling analyses and provide useful constraints on the evolution of the Irish Ice Sheet prior to and following the LGM. These analyses show that: (1) existing GIA ice model reconstructions for Ireland are inconsistent with some of the new glaciological and RSL data, which support the suggestion that LGM ice thickness was generally greater than previously envisaged for this region; (2) predictions of RSL are responsive to the specific rate and timing of deglaciation prior to the LGM with the Irish RSL data strongly favouring very rapid deglaciation after 21 000 cal. yrs BP; (3) predictions of postglacial RSL in this region are especially sensitive to shallow Earth viscosity structure. Empirical RSL observations from both northwest Ireland and the Irish Sea Basin provide an important constraint on the inferred viscosity of the Earth upper mantle in this region. This new finding has direct implications for the validity of existing GIA modelling-based reconstructions of the British Ice Sheet.

Detailed litho- and biostratigraphical analysis has been employed in conjunction with accelerator mass spectrometer radiocarbon dating to develop a suite of sea level index points from the inner Shannon estuary on the west coast of Ireland. These new data reveal two distinct phases of RSL change: The first of these occurs in the early Holocene (c. 9000 – 7000 cal. yrs BP) and is associated with a period of rapid sea-level rise of around 5-6 mm yr^{-1}. The second phase of RSL change takes place in the mid Holocene (c. 7000 – 3000 cal. yrs BP) and is characterized by a much slower rate of RSL rise of c. 1.0 mm yr^{-1}. Elsewhere along the west coast of Ireland (outer Shannon estuary, Connemara, Donegal) further radiocarbon dates from freshwater peat horizons provide maximum limits for the position of palaeo mean sea level back to 14 000 cal. yrs BP. These new RSL observations significantly improve the quality and quantity of sea level information available from the west coast of Ireland. The new RSL data are in agreement with the glacial rebound model predictions of RSL for the west coast of Ireland although contribute little to the debate surrounding possible higher than present late Devensian RSL along this stretch of coastline.

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Abbreviations used in the text

HAT	Highest astronomical tide
MHWST	Mean high water of spring tide
MHWNT	Mean high water of neap tide
MSL	Mean sea level
MLWNT	Mean low water of neap tide
MLWST	Mean low water of spring tide
LAT	Lowest astronomical tide
AMS	Accelerator mass spectrometer
AZA	Assemblage zone approach
BIIS	British and Irish ice sheet
BIM-1	Shennan *et al.* (2006a) British Isles ice model
DCA	Detrended correspondence analysis
GIA	Glacio-isostatic adjustment
GRM	Glacial rebound modelling
IGCP	International Geological Correlation Programme
IM	Indicative meaning
IOD	Irish ordnance datum
IR	Indicative range
LGM	Last glacial maximum
MWP	Melt water pulse
NRA	National Roads Authority
OD(B)	Ordnance datum Belfast
ODM	Ordnance datum Malin Head
OS	Ordnance Survey
Pmsl	Palaeo mean sea level
Psu	Practical salinity units
RSL	Relative sea level
RWL	Reference water level
SIS	Scandinavian ice sheet
SLIP	Sea level index point
SWLI	Standardized water level index
TFA	Transfer function approach
v_{LM}	Lower mantle viscosity
v_{UM}	Upper mantle viscosity
WA-PLS	weighted averaging partial least squares

CHAPTER ONE

Introduction

1.1 BACKGROUND

Changes in the level of the sea with respect to the land surface manifest from the interaction of a variety of interrelated environmental factors operating over varying spatial and temporal scales. Together, these environmental variables control movements of relative sea-level (RSL) which may be preserved in the geological record. As such, RSL data can be employed directly or otherwise to further our understanding of Quaternary environmental processes including glacial eustacy, isostacy, ice sheet dynamics, deglaciation histories, palaeoclimate and changing palaeogeographies (Long, 2003).

Evidence from both the geomorphological and sedimentary records bears testimony to the fact that Ireland has experienced repeated episodes of glaciation interspersed with periods of ice retreat in the recent geological past (e.g. Richter *et al.*, 2001). The build up and wastage of the Quaternary ice sheets has resulted in both locally and globally driven changes in RSL which have led to major shifts in shoreline position over the shelf seas surrounding the Irish coast. Variations in the extent and thickness of the Irish Ice Sheet during the Devensian have induced temporally and spatially different patterns of land surface deformation in response to loading and unloading episodes. These differences in 'glacio-isostatic adjustment' (GIA) superimposed upon a general 'eustatic' rise in sea level generated by global melt-water input, have resulted in uneven patterns of RSL change around Ireland. This is evidenced by 'raised beaches' in the north indicating RSL fall along with drowned valleys in the south reflecting RSL rise. The raised shorelines of the north expose vertical displacements of the Earth's crust well in excess of 100m and the legacy of glacio-isostatic rebound left by the last ice sheet is still being experienced today (Todd, 1981; Carter, 1982b; Devoy, 1999).

In the north of the British Isles, the magnitude of glacio-isostatic rebound subsequent to the LGM has been similar to the global meltwater influx. These components have dominated at different

times in the postglacial period and this has resulted in RSL histories that are highly non-monotonic in time (e.g. Shennan and Horton, 2002). Long time-series records of change have been captured in coastal sedimentary archives from Scotland (e.g. Shennan *et al.*, 1995b, 2000a) and now Ireland (Clark *et al.*, 2004) and detail these variations in RSL as far back as 19 000 BP. These records have been used on their own and in conjunction with geophysical models to constrain shallow Earth viscosity structure (e.g. Peltier *et al.*, 2002), local to regional ice sheet histories (e.g. Shennan *et al.*, 2002, 2006a) as well as the magnitude/timing of global melt water signals (e.g. Shennan, 1999).

An essential quality of a sea-level indicator, whether this is a geomorphological feature (e.g. raised beach), a lithostratigraphical unit (e.g. inter-tidal silt), or a biological assemblage (e.g. salt-marsh foraminiferal group), is that its vertical distribution can be related consistently to sea level (Van de Plassche, 1986; Edwards *et al.*, 2004a). In Ireland, much of the available sea level data are in the form of 'raised shoreline' evidence (a generic term used to describe geomorphological features such as marine notches, washing limits and raised beaches) which cannot be unambiguously related to sea level. Until recently, little emphasis had been placed on unravelling the evidence for postglacial sea-level changes contained within the thick sequences of unconsolidated sediments found in many Irish estuaries and tidal inlets (Penney, 1983). Evidence which is sourced from the lithostratigraphical record often only records wholly terrestrial environments (e.g. freshwater peat deposits) which also possess no quantifiable relationship with sea level. Most of these records come from studies looking into various aspects of Quaternary environmental change and the sampling design employed by the individual researcher was often not specifically defined to test particular hypotheses relating to trends in RSL. As a consequence, these data are commonly qualitative, or semi-quantitative (large error terms) in nature, and so simply constrain the former position of RSL, rather than reconstruct or define it. The lack of systematic research focussed on the reconstruction of RSL change means that data are unevenly distributed in both space and time and, in areas such as the west coast of Ireland, even generalised patterns of change are unknown. Published syntheses of data seeking to progress understanding beyond the local scale have been attempted (e.g. Carter, 1982a; Shaw and Carter, 1994; Carter *et al.*, 1989). However, most of these investigations struggle to produce more than a semi-quantitative first assessment due to fundamental limitations in the available data.

Attempts to predict postglacial RSL for locations around the coast of Ireland on the basis of geophysical models have also been made (e.g. Lambeck, 1996, Lambeck and Purcell, 2001). In brief, this modelling approach involves the combination of an ice model predicting the thickness, distribution and melting history of terrestrial ice, and an Earth model simulating the Earth's deformation under varying loads of ice and meltwater. The modelling process follows an iterative procedure in which the output from the model runs is compared with empirical field evidence. However, uncertainties over the quality of the RSL data lessen the extent to which this observational evidence can be used to constrain model parameters (Lambeck, 1996; Cooper *et al.*, 2002). Since the work of Lambeck (1996) and Lambeck and Purcell (2001), a significant body of new glaciological evidence has emerged from Ireland (e.g. O'Cofaigh and Evans, 2001; Rae *et al.*, 2004; McCabe *et al.*, 2005; Ballantyne *et al.*, 2006). Many of these recent findings contradict the ice model adopted by Lambeck (1996) and in doing so provide a strong caveat on the validity of the RSL predictions by this model. In addition, a number of methodological developments in both sea-level reconstruction and geophysical modelling have recently occurred (e.g. Shennan *et al.*, 2002, 2006a; Milne *et al.*, 2006). Consequently, there is now a pressing need to revisit the topic of glacial rebound modelling in Ireland in light of these empirical and methodological advances.

1.2 Scientific Context

Analysis undertaken in this thesis is structured towards meeting three complementary research aims:

1 To compile a database of standardised sea level information from Ireland.

Database development will closely follow established protocol for the analysis and interpretation of sea level data developed during International Geological Correlation Programme (IGCP) projects 61 and 200 and described in 'Sea-level research: A manual for the collection and evaluation of data' (Van de Plassche, 1986). The information contained within the database will serve as a useful resource for scientists working in the coastal zone and will also provide important constraints on glacial rebound model input parameters (see aim 3 below).

2 To collect new, reliable RSL data from sites around the Irish coastline.

The development of the sea-level database (Aim 1) will highlight those geographical areas where movements of postglacial RSL are poorly understood. The collection of new data will be focused upon these localities and will be achieved through a combined litho-, bio- and chronostratigraphical approach which will enable the development of sea level index points (SLIPs). These SLIPs may be combined with existing data (sourced from the sea-level database) to construct local chronologies of RSL change.

3 To develop a new geophysical model capable of predicting postglacial RSL change around Ireland.

The new sea-level database will be employed to test the Irish component of the British and Irish Ice Sheet (BIIS) proposed by Shennan *et al.* (2006a). Misfits between the empirical data and model RSL predictions will be highlighted and a new, improved Irish ice model will be advanced. The refined model will also be used to generate a series of palaeogeographic maps for the period 20 000 BP to present, which may be used to elucidate aspects of late Quaternary coastal evolution around the western margin of the British Isles.

1.3 Thesis Structure

Chapter 2 introduces the reader to some basic concepts and terms which are familiar to the contemporary sea-level scientist. The published evidence for postglacial sea-level movement around Ireland is then considered in the context of more recently established protocol for the analysis and interpretation of sea level data.

Chapter 3 outlines in detail the methods used in selecting field sites, and the methodology employed in collecting sea level data from these individual study sites. It also considers the relationship between both foraminifera and diatoms and sea level and considers their role in quantitative palaeoenvironmental reconstruction.

Chapter 4 looks at the rationale and methodology behind the development of a sea-level database for Ireland. The criteria used for the categorization and rejection of data are explained in full along with the strengths and limitations of the final database.

Chapter 5 presents the results of the glacial rebound modelling analyses. The validity of the Irish component of the BIIS model proposed by Shennan *et al.* (2006a) is assessed using the newly developed sea-level database laid out in Chapter 4. An improved ice-Earth model combination is advanced which is in better agreement with available sea level and glaciological evidence.

Chapter 6 describes the environments encountered at the individual study sites in terms of their litho-, bio- and chronostratigraphy. The newly presented data from this chapter is combined with existing data and used to develop local chronologies of RSL change.

Chapter 7 presents a discussion of the findings from the previous two chapters and considers the strengths and weaknesses of the newly acquired data. The chapter brings together the newly developed local chronologies of RSL change outlined in chapter 6 with geophysical model predictions of RSL derived from the glacial rebound model developed in chapter 5. The 'goodness of fit' between the two sets of data is discussed and the most important discrepancies are singled out.

Chapter 8 presents the conclusions of this thesis, and assesses to what extent the aims (outlined in section 1.2) have been met. Based upon the findings and discussion contained in chapters 5 - 7, recommendations for future research are made.

CHAPTER TWO

Late Devensian and Holocene RSL Change in Ireland: A Review of the Literature Base

2.1 INTRODUCTION

Investigations related to past sea-level change in Ireland have a relatively long history, dating back to the end of the nineteenth century with the works of Praeger (1893, 1896). Numerous investigations have followed with a particular surge in interest in the period between the early 1960's to mid 1980's. This chapter does not represent an inaugural attempt at reviewing sea-level research undertaken in Ireland. Several authors in the more recent past have attempted similar reviews or regional examinations, the most comprehensive of these being undertaken by Devoy (1983). However, over 20 years has elapsed since this work was compiled and a number of investigations have been undertaken since. Moreover, the early 1980's saw a virtual paradigmatic change in the sea-level community with IGCP project 61 leading to a complete revaluation of the way sea level data were analysed and interpreted (e.g. Kidson, 1982; Shennan *et al.*, 1983; Tooley, 1985; Van de Plassche, 1986). It is now desirable to look back over the Irish literature base in the light of these more recent developments.

2.2 UNDERSTANDING SEA LEVEL DATA

Observations of past sea-level changes around the coast of Ireland come from both the geomorphological record along with both organic and minerogenic deposits contained within sedimentary sequences. To be of use as a gauge of palaeo-sea-level (a 'sea-level indicator'), it is fundamental that the feature was formed *in situ* and can be systematically related to a former tide level. This latter point refers to the concept of the 'indicative meaning' (IM) and describes the vertical relationship between the local environment in which a sea-level indicator accumulated and a contemporaneous reference tide level (Preuss, 1979; Shennan, 1986; Van de Plassche, 1986; Horton *et al.*, 2000) (Figure 2.1). It is defined in terms of the modern vertical range occupied by the sea-level indicator (the indicative range) measured relative to a given tide level (the reference water level) such as mean high water spring tide (Edwards, 2006b). If the sample location, age and

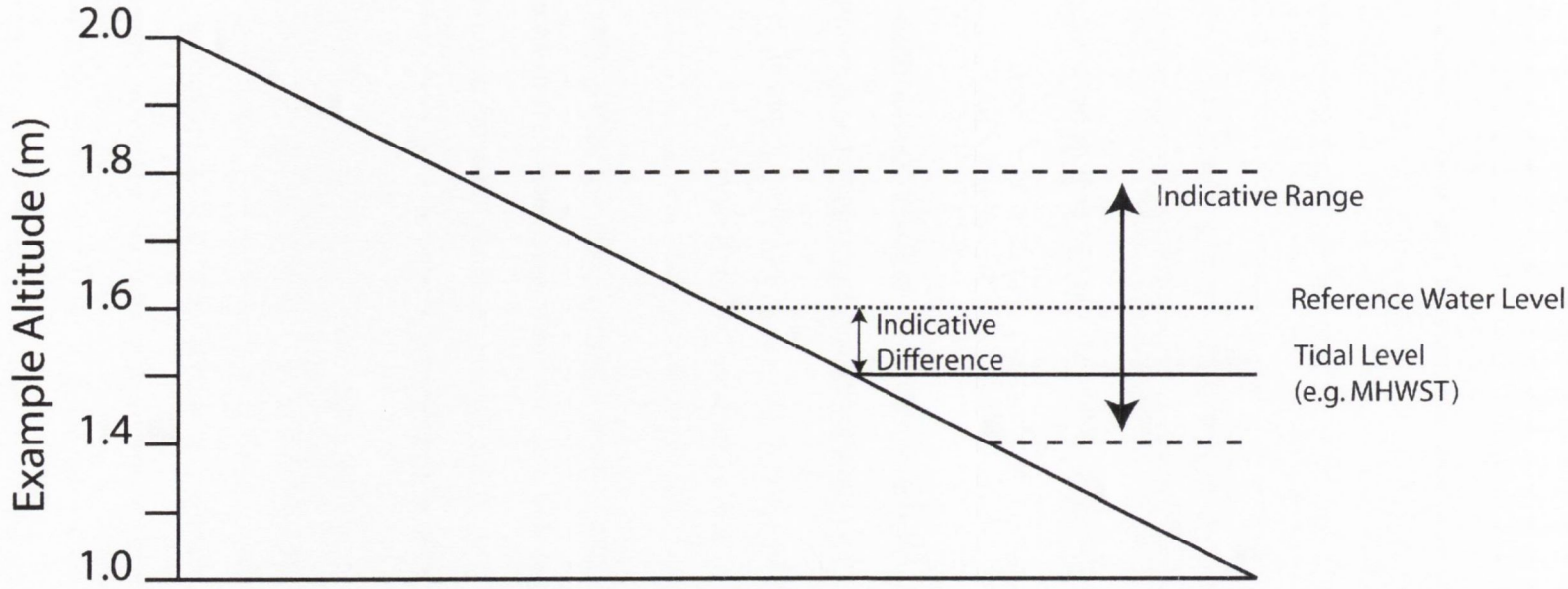


Figure 2.1 A diagram showing the components of an indicative meaning. The indicative range refers to the vertical interval over which a sea-level indicator is found. The indicative difference describes the position of the indicative range (called the reference water level) relative to a defined component of the tidal frame. Here, the reference water level is MHWST + 0.1 m, and the indicative range is ± 0.2 m. (Taken from Edwards, 2006b)

altitude are also known, the information may be combined to form a sea level index point (SLIP) which fixes the past altitude of sea level in time and space (Tooley, 1978; Preuss, 1979; Van de Plassche, 1986). Details of the sea level index point methodology are given in section 3:1.

Whilst a large volume of sea-level research has previously been undertaken in Ireland, the majority of these studies have employed controvertible indicators of sea level. These include raised beaches and other ambiguous sources such as freshwater peat deposits which possess a largely unquantifiable indicative meaning. As a consequence, many of these data are largely qualitative, accompanied by large vertical error components which only allow for the former position of sea level to be constrained, not defined (See Chapters 3 and 4 for further discussion).

It is important to note that in Ireland, a number of different levelling datums have been adopted in sea-level studies including Dublin, Belfast, Malin Head and mean sea level. Here, all heights have been converted to OD (Belfast) - a mean sea level datum. Corrections were made as follows: Ordnance Datum (Belfast) lies approximately 0.03m below Ordnance Datum (Malin Head) and 2.7m above Irish Ordnance Datum (IOD) (Admiralty Tide Tables, 1997). Dixon (1949) states that Irish Ordnance Datum is approximately 8 feet below Ordnance Datum (Newlyn) thus Ordnance Datum (Belfast) lies approximately 0.3m above Ordnance Datum (Newlyn).

All ages are expressed in calendar years before present (1950 AD) unless otherwise stated. Conversions between radiocarbon ages and calendar years have been carried out using CALIB 5.0.1; 95% confidence limits (Stuiver *et al.*, 2005). Further details on the errors associated with the radiocarbon dating procedure can be found in section 3.7.

2.3 Late Devensian Sea Levels

Around the Ulster and north Leinster coastline, an abundance of geomorphological evidence such as raised beach deposits, washing limits and marine notches found well above present sea level offered clear evidence to many early researchers of higher than present sea levels at some point in the past. In particular, the ubiquity of the features as well as the ease of access to them has led to their over representation in studies of Irish RSL (e.g. Synge, 1977a, 1985; Mitchell, 1977; Stephens and McCabe, 1977; Stephens and Synge, 1965). Much of the focus on them has been associated

with the development of raised shoreline correlation diagrams which trace spatial and temporal changes in shoreline height (e.g. Synge and Stephens; 1966; Synge, 1977a).

Consideration of the work by Stephens and Synge (1965) documenting a series of Lateglacial raised shorelines along the north Ulster coastline suggests that the earliest marine transgression following the LGM is recorded by marine washing limits, notches and shingle bars on the Inishowen Peninsula (Donegal). The highest of these raised shorelines are found at altitudes of 19-22m OD(B) at Malin Head and are found to decrease in altitude to the west, broadly signifying decreasing proximity from the centre of former ice cover. The authors go on to propose the absence of raised shorelines to the west of Bloody Foreland and along the north west coast of Donegal yet the findings of Gaulin (*cited in* Devoy, 1983) would suggest that such an assertion was born out of an absence of proof rather than proof of absence. Further work by Synge and Stephens (1966), Synge (1977a, b) and Mitchell (1977) documents other raised (Lateglacial) shorelines from along the north Ulster and east Leinster coastline with the highest of these shorelines found at a height of c. 25-30m OD(B) along the Co. Down coast.

Dating these raised shorelines has proved notoriously difficult. Based on the relationship of various strandlines to known ice limits, the oldest of these shorelines in Ulster is said to have formed around 22 000 BP (Synge, 1977a) with a series of successively younger (and hence less steeply tilted shorelines) postdating the oldest of these shorelines in the region. This age was assigned on the basis of correlation with radiocarbon dated basal organic sediments sampled on the Isle of Man. However these radiocarbon dates are now believed to have been contaminated with older carbon, either from a calcareous diamict substrate or from isotopic fractionation during subaqueous photosynthesis of *Drepanocladus revolvens*, from which the dates were derived, producing a hard water error (Dackombe and Thomas, 1985; Thomas *et al.*, 2004). Attempts to correlate the shorelines from the north of Ireland with those from southwest Scotland have been made (e.g. Synge and Stephens, 1966; Stephens, 1968). However, these have met with limited success contributing to the suggestion that tectonic activity took place in the North Channel during isostatic recovery (Stephens, 1970; Peacock, 1971; Stephens and McCabe, 1977).

However, despite the prevalence of raised shoreline diagrams in Irish RSL literature, during the past 25 years or so they have been overlooked in favour of sea level information derived from litho-

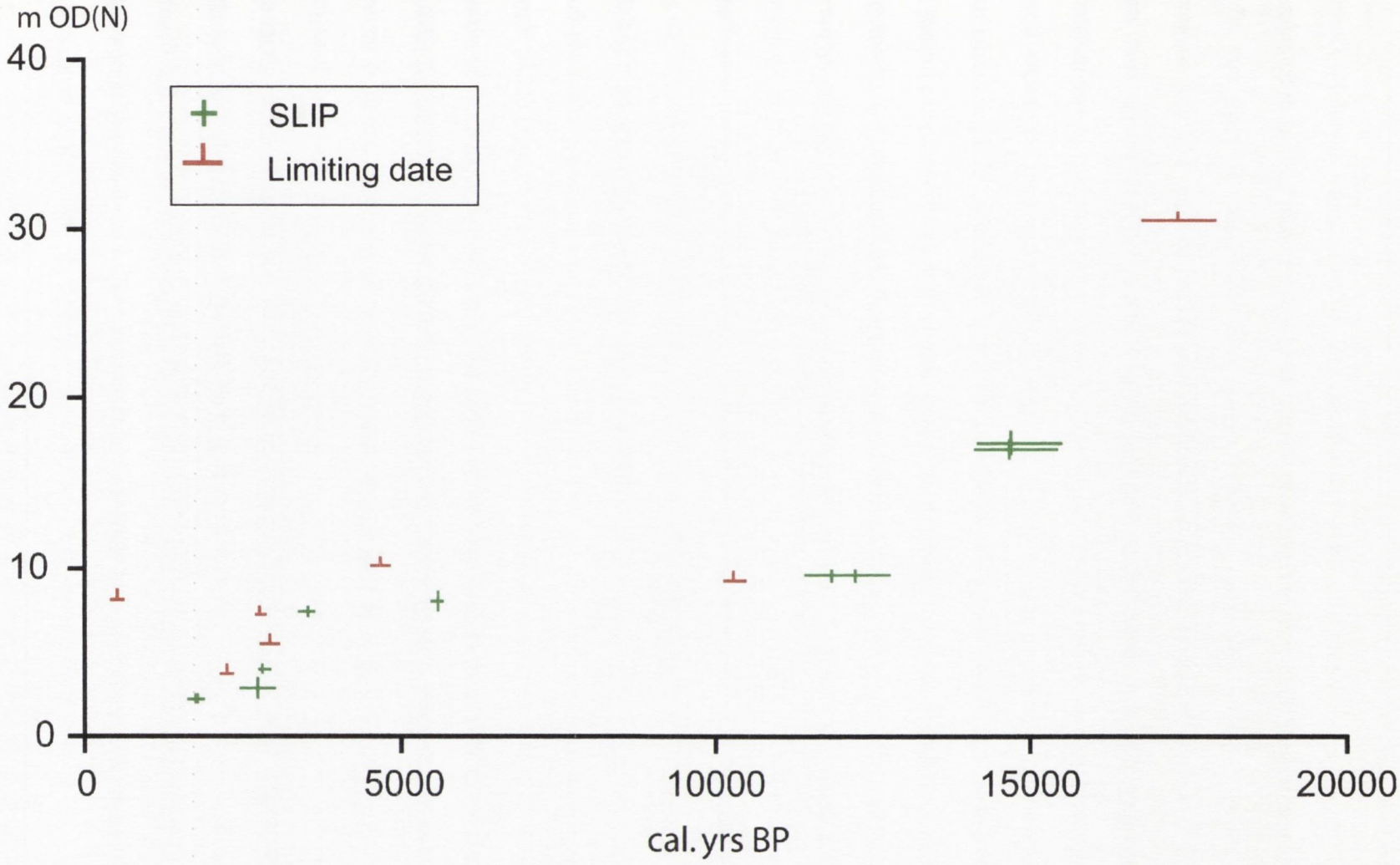


Figure 2.2 Relative sea level age-altitude graph from Knapdale, Scotland. (Shennan *et al.*, 2006b)

and biostratigraphical investigations. This move away from raised shoreline research owed much to the findings of IGCP project 61 (and the works that accompanied it) which emphasized the errors inherent in sea-level investigations. It became recognized that raised shoreline features only represent limited expressions of marine activity which, while related to sea level, do not necessarily reflect it in any consistent or coherent manner (Carter, 1982a). Furthermore, the raised shoreline diagrams were often comprised of a range of different sea-level indicators (such as beach deposits, washing limits and marine notches) and the reconstructed sea level is unlikely to represent a single sea surface altitude (Devoy, 1983). As well as the belated realization that these shoreline features possessed varying and indefinable indicative meanings, it became increasingly clear that the ages assigned to the features were speculative at best. Few absolute ages for raised shorelines exist and inferred ages have commonly been a product of circular reasoning (e.g. Addyman and Vernon, 1966).

Whilst there have been a number of recent studies from Scotland attempting to link raised shorelines with dated lithostratigraphic horizons (e.g. Smith, 2005; Smith *et al.*, 2002, 2006), many workers have preferred to use isolation basin records to accurately fix the altitude of former sea level (e.g. Shennan *et al.*, 1995b; 1999; 2000a; 2006b). Radiocarbon dating of the terrestrial-marine transition contained within basin deposits has enabled SLIPs to be established and in Scotland, these have provided an important record of sea-level change dating back over 16 000 years. Recently, Shennan *et al.* (2006b) have established isolation basin records of sea-level change from Knapdale and Islay in southwest Scotland. These locations both lie in close proximity to the north Ulster coast and as such, are worth taking into consideration here. At Knapdale, SLIPs dating to 14 680 BP show that sea level stood at +17m above present at this time, falling to around +9.5m above present by 12 200 BP (Figure 2.2). No evidence is found for Late Devensian sea level higher than 30m above present in this locality although the registration of sea level above this height may have been inhibited by local ice cover.

Despite the problems associated with using raised shoreline features to fix a former position of sea level in time and space, the fact that sea level along the north and east Ulster coastline occupied a height someway above present at the time of the LGM and during much of the Lateglacial would appear certain. However, by the time of the Lateglacial – Holocene transition, sea level had fallen to a lowstand depth well below present level, as evidenced by a number of radiocarbon dates

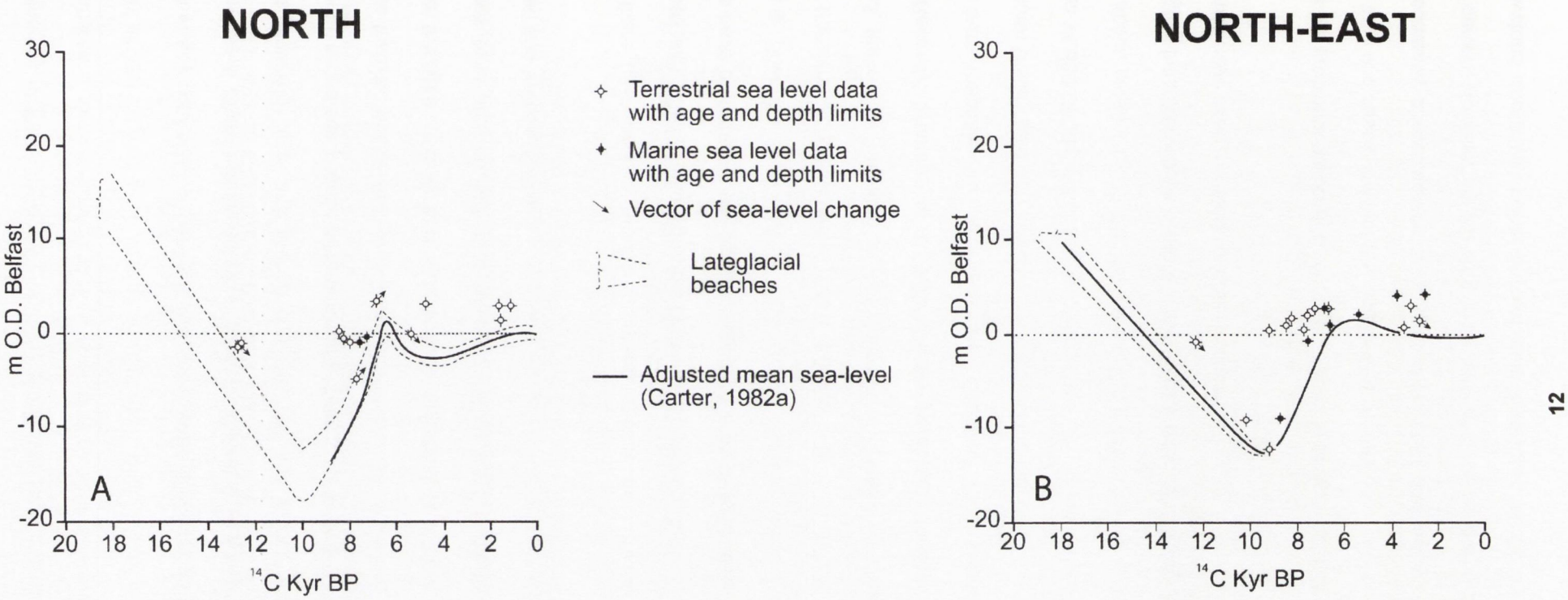


Figure 2.3 Relative sea level 'curve' for **A** the north and **B** north east of Ireland. (Taken from Carter, 1982a). NB: These 'sea level data points' are not synonymous with SLIPs. (See text for details)

taken from (now submerged) organic sediment horizons. This sea-level fall (induced by crustal rebound associated with the removal of ice) had taken place despite rapid rates of eustatic sea-level rise which at times exceeded 50 mm yr^{-1} during this time interval (Fairbanks, 1989; Bard *et al.*, 1990a). Radiocarbon dated peat deposits from Roddens Port (Morrison and Stephens, 1965) coupled with similar evidence from elsewhere along the north east Ulster coast (Prior and Holland, 1975; Prior *et al.*, 1981) shows that by about 14 000 BP, sea level had fallen below present level. These findings are broadly in agreement with the work of Carter (1993) whose work on an emergent gravel barrier on Rathlin Island (off the north coast of Ireland) suggests sea level had fallen below present by around 14 500 BP. Relative sea level continued to fall after this time, reaching a lowstand depth at the time of the Lateglacial-Holocene transition. It is suggested that with increasing distance from the centre of crustal loading, the lowstand depth increases in magnitude and occurs earlier in time until such a point that the sea level curves begin to resemble those from sites distal to the ice margin (Lambeck, 1995, 1996).

Up until recently, the best estimate for the altitude and age of this lowstand depth came from the efforts of Carter (1982a) who constructed two sea level 'curves' for the north and east Ulster coastline based on a suite of over 30 radiocarbon dates (Figure 2.3a,b). Along the north east Ulster coast, RSL is suggested to have fallen to about -15m OD(B) whilst on the east Ulster coast, a depth of -13m OD(B) is inferred. Both these lowstand minima were said to have occurred around 11 500 BP and this accords well with the glacial rebound model estimates of Lambeck (1996). (One must be cautious in attaching significance to this however, since the model predictions of Lambeck have been calibrated to fit the findings of Carter (1982a)). Despite these inferences, the data employed in the Carter (1982a) analyses are varied, inhomogeneous and unreliable (Lambeck, 1996) and contradicts the established guidelines for sea-level research laid out by IGCP project 61. Most of the dated sea-level 'indicators' originate from likely allochthonous material or are taken from deposits which have no clear relationship with sea level (see chapter 4 for further details). Furthermore, the compilation of sea level data from different locations onto a single age–altitude plot no longer remains acceptable practice in sea-level research not least because significantly different isostatic histories may exist over comparatively small geographical areas (Tooley, 1978). Ulster is a case in point as the rebound modelling studies of Lambeck (1996) suggest that since the start of the Holocene, sea-level histories may have differed by over 10m within the two curve regions defined by Carter (1982a).

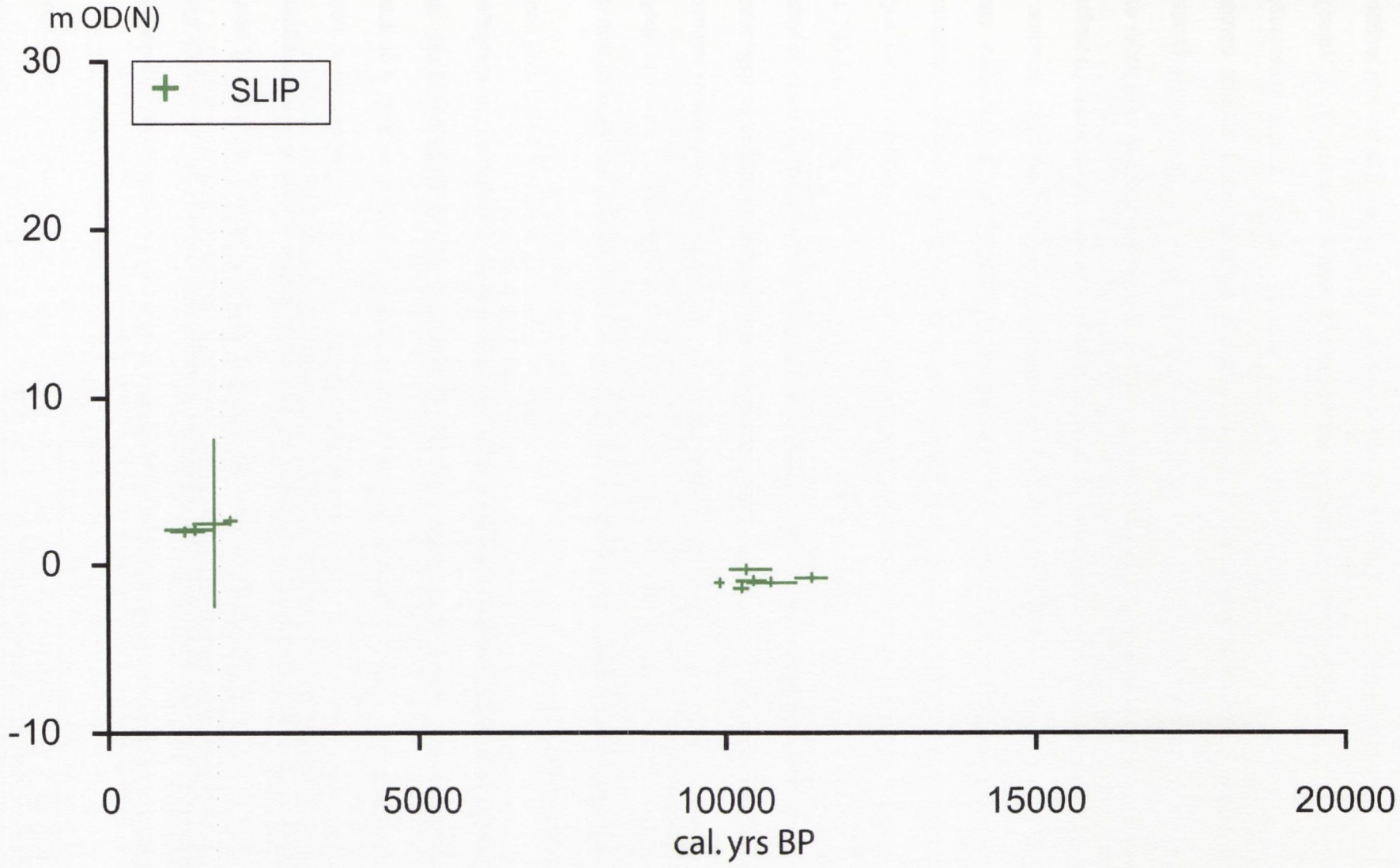


Figure 2.4 Relative sea level age-altitude graph from Islay, Scotland. (Shennan *et al.*, 2006b)

Cooper *et al.* (2002) had presented inner shelf seismic stratigraphic data to propose a lowstand depth of -30m OD(B) at Portballintrae on the north east Ulster coast although the more recent findings of Kelley *et al.* (2004) prove these earlier results to be invalid. Nevertheless, Kelley *et al.* (2006) have recently uncovered new evidence for this lowstand from submerged 'beach' deposits at -30m OD(B) in Belfast Lough (on the east Ulster coast). AMS radiocarbon dates obtained from an 'inter-tidal' molluscan assemblage return ages of around 13 400 BP. This appears to question both the age and depth for the lowstand inferred by Carter (1982a) by contradicting the (onshore) radiocarbon evidence from nearby Roddan's Port (Morrison and Stephens, 1965). However, whilst the lithostratigraphic units described by Kelley and co workers appear consistent with the characteristics of beach material, it doesn't necessarily follow that the units *are* beach material. Furthermore, it cannot be said with certainty that the dates obtained for this submerged shoreline were derived from *in situ* material. A short distance to the north of Ulster at Islay (southwest Scotland), a suite of SLIPs of Lateglacial-early Holocene age help to define the depth of the lowstand in this locality (Figure 2.4). Here, dates ranging in age from 9900 to 11 400 BP show that RSL was around 1m below present level at this time (Shennan *et al.*, 2006b).

The fragmentary evidence of Late Devensian sea level around the north and east of the country means that large gaps remain in our knowledge of sea-level movements in this region. Despite this, our understanding of sea-level change in the north of the country is still more complete than from elsewhere around Ireland's coastline where much of the Late Devensian sea-level record is thought to exist offshore (Devoy, 1999). Recently, bathymetric and seismic evidence considering shelf morphology around Ireland's south and west coasts has uncovered extensive evidence of submerged coastal (and terrestrial) landforms such as palaeovalleys and meltwater channels and are commonly found -80 to -100m below present sea level (Devoy, 1999). Many of these features are suggested to be of LGM or Lateglacial age although firm chronological control remains absent. In Cork the rock-cut channel of the former River Lee lies at -70 to -90m msl at Roche's Point, whilst in Dingle Bay seismic surveys show a complex anastomosing channel system cut through glacial debris and into the rock base at depths below -60m msl (Devoy, 1999). Further seismic data from off the coast of Waterford also suggests sea level at least as low as -55m msl at the LGM (Devoy, 2003).

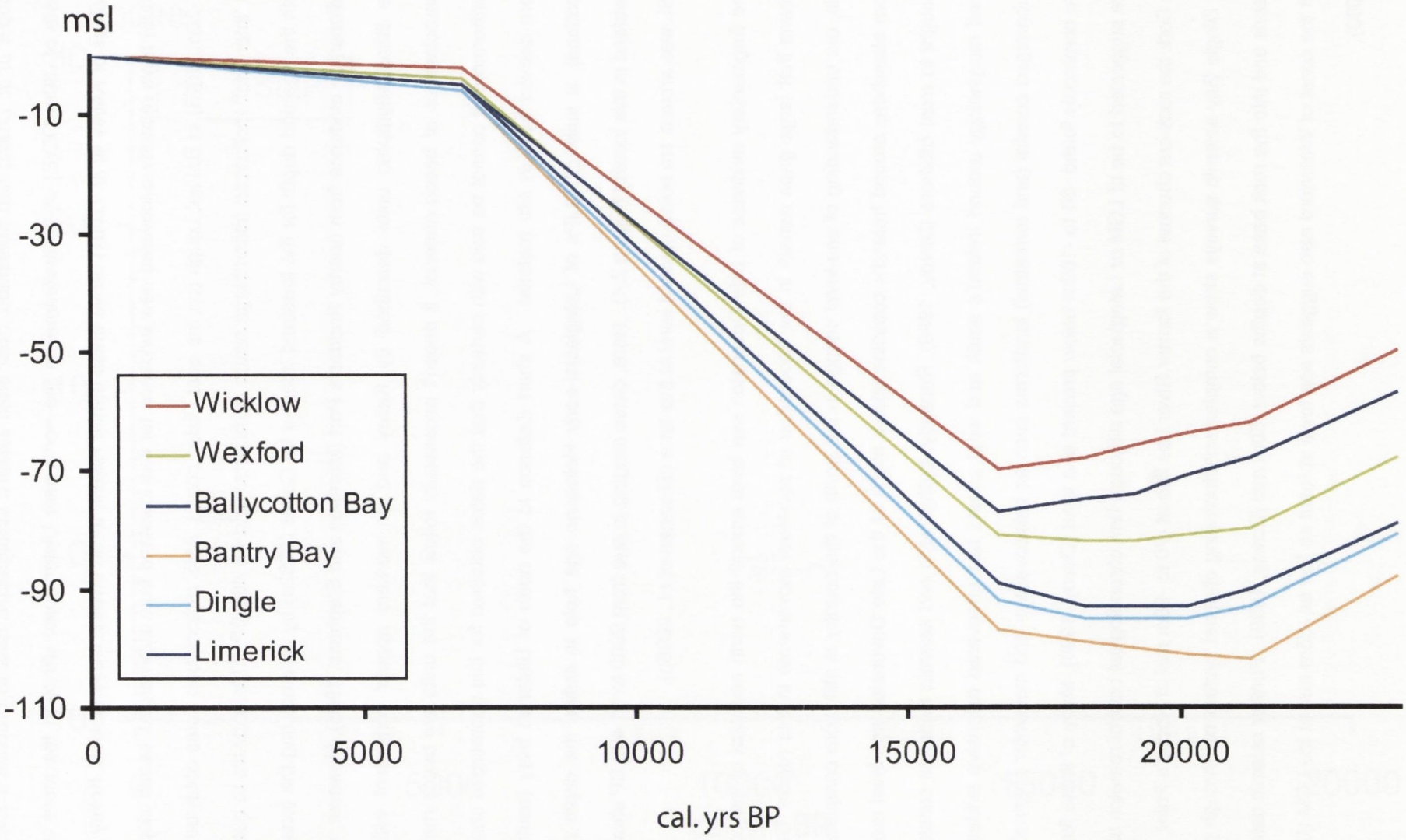


Figure 2.5 Glacial rebound model predictions by Lambeck (1996) for sea-level change around the south and southwest coast of Ireland

On the Irish Atlantic margin, deepwater cores (>1000m) from the Barra Fan and Porcupine Trough show the dominance of extensive terrestrial sediments and turbidite sedimentary sequences during the last glacial stage (Devoy, 1999). These data are said to evidence the seaward migration of the coastline at a time of low RSL, with the sub-aerial erosion of and extension of glacial and fluvioglacial systems across the shelf margin. Subsequent to the LGM, the replacement of land-derived sediments by pelagic and hemipelagic sediments in the stratigraphic record is said to record waning glacial activity and concomitant sea-level rise. Finally, in the Celtic Shelf region to the southeast of Ireland, Scourse and Furze (2001) bring together evidence from a number of short (<5m) sediment cores. Taken together, the data from this region provide a proximal to distal transect through the flow trajectory of the Late Devensian Irish Sea ice stream and support the general concept of sea level occupying an LGM depth of over -100m msl.

Despite this new seismic and stratigraphic data, the overall dearth of empirical evidence means that for the most part, estimates for the position of sea level during the Lateglacial rarely rise above the level of conjecture (e.g. Taylor *et al.*, 1986). An exception to this is the numerically based modelling efforts of Lambeck, (1996) which (arguably) deliver the most comprehensive picture of RSL change around Ireland, both in time and space. On the basis of this model, it is suggested that all locations around Ireland's south and west coast experienced falling sea level after the LGM as a consequence of glacio-isostatic rebound outpacing eustatic sea-level rise (Figure 2.5). A lowstand minimum was then reached at some point between 21 000 to 17 000 BP (depending on proximity to the centre of crustal loading). By 10 000 BP, sea levels are suggested to have risen to approximately -20 to -40m msl along the east coast and to levels nearer to -50m along the south and west coasts. In some areas of south/west Ireland, isolated observational evidence of Lateglacial sea levels are available to provide assessments of the validity of these predictions. In Bantry Bay, Stillman (1968) used both litho- and biostratigraphical evidence to suggest that sea level occupied a position around -60m msl at 13-14 000 BP. Elsewhere, Scourse and Austin (1994) use radiocarbon dated shell evidence from a Celtic Sea core to infer that sea level was about -70m shortly before 13 000 BP. Along Ireland's west coast, palynological evidence provided by Jessen (1949) shows that sea level can have been no higher than -5m at 14 000 BP. These observations all concur with the model predictions of Lambeck (1996) yet more recent debate

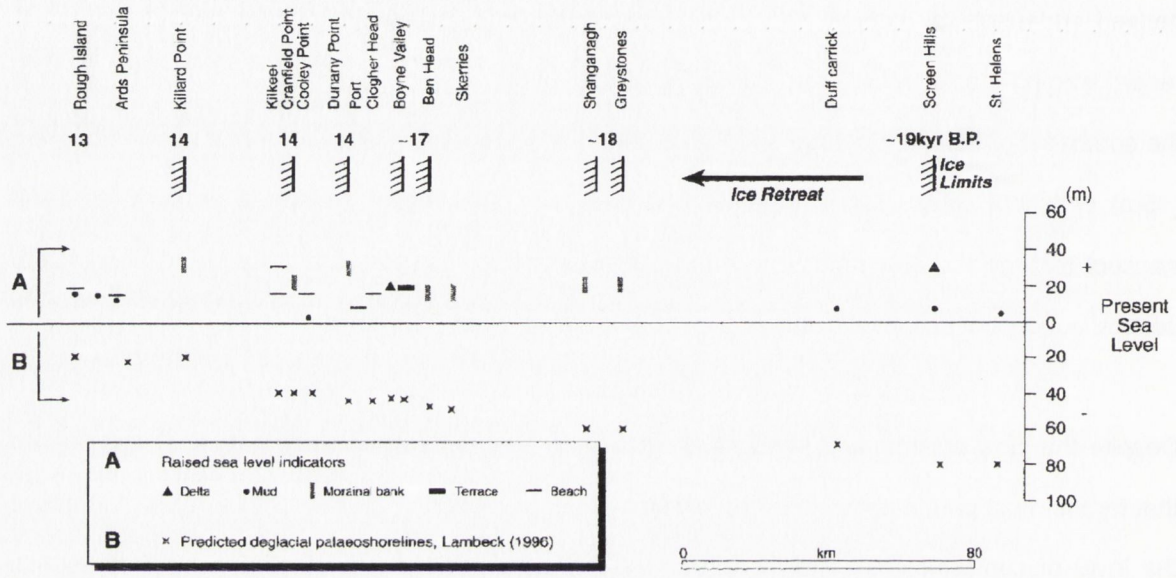


Figure 2.6 A comparison of water depths inferred from raised late-glacial sea-level indicators along the coast of eastern Ireland with Lambeck's (1996) predicted shoreline levels. (Taken from McCabe, 1997)

concerning the reliability of this model recommends a re-assessment of the Lambeck (1996) predictions for Ireland (e.g. Shennan *et al.*, 2006a).

The results of Lambeck (1996) are supportive of the conventional model of Late Devensian sea level in Ireland which has sea level (predominantly) above present in the north of the country and well below present towards the south. However a number of studies have come to light over the past two decades that have questioned this overall pattern, suggesting that higher than present sea levels were experienced around most of Ireland's coastline at this time. Through analysis by Eyles and McCabe (1989) and McCabe and O'Cofaigh (1996) of coastal sequences in the south/east; McCabe *et al.* (1986, 1993, 2005) in the north (Donegal) and west (Mayo), and through research by McCabe and Clark (1998) in the north east, evidence has been presented for Late Devensian shorelines commonly many tens of metres above heights given by geophysical model predictions (e.g. Lambeck, 1996; Shennan *et al.*, 2002) (Figure 2.6). This glaciomarine model for deglaciation suggests that the weight of ice at the LGM was sufficient to raise RSL's well above their present height, destabilising the ice margin and causing rapid deglaciation (McCarroll, 2001). An important component of these studies is the use of AMS radiocarbon dating on foraminifera furnished from glaciomarine sediments. These dates provide minimum altitudes for sea level and have, for the first time in Ireland, enabled distinct episodes of sea-level rise and fall to be identified and dated in the Lateglacial period (e.g. Clark *et al.*, 2004; McCabe *et al.*, 2005) (Figure 2.7). However, much of this new evidence derived from 'glaciomarine' mud sequences remains controvertible. In particular, many have questioned whether in fact the marine faunal evidence is *in situ*, instead suggesting that a number of the sedimentary units containing foraminifera have been deposited in a marine setting then subsequently transported and reworked by ice (Warren, 1991; Hoare, 1991).

Efforts to reconcile discrepancies between the numerical and glacio-sedimentary models have been limited; however the work of Lambeck and Purcell (2001) offers an objective contribution to the debate. Here, the authors set out to examine under what conditions glacial rebound models can predict high sea levels during deglaciation on both sides of the Irish Sea (e.g. Eyles and McCabe, 1989) whilst satisfying constraints placed by Holocene RSL evidence. Given that the Earth model employed in both the Lambeck (1995) and Lambeck (1996) papers is found to deliver a reasonable overall representation of the RSL change around the British Isles, no evidence is

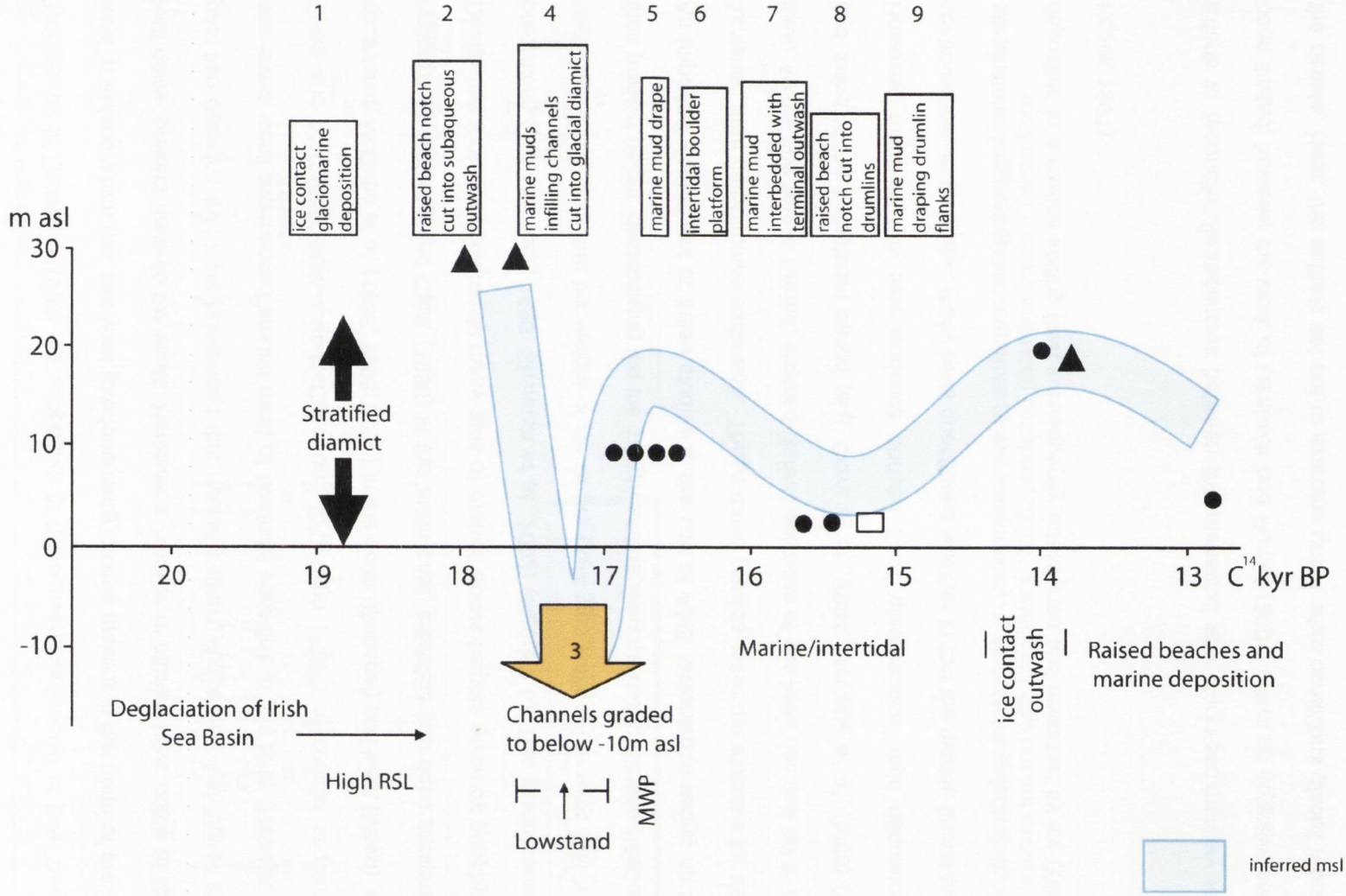


Figure 2.7 Summary chart of changes in (Lateglacial) RSL, County Down, eastern Ireland. (Adapted from McCabe *et al.*, 2005)

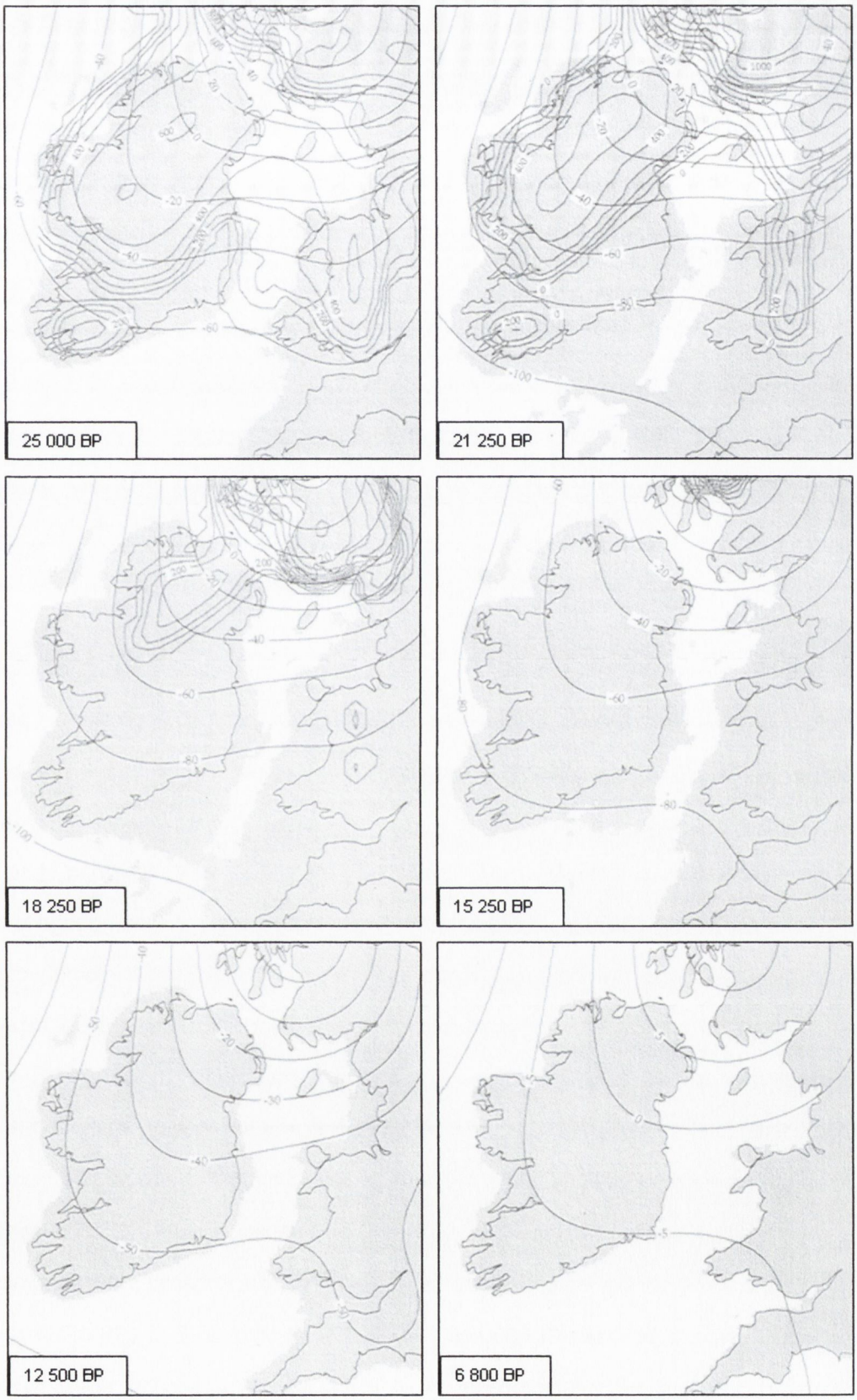


Figure 2.8 Predicted isobases of sea level relative to present (continuous black curves), palaeo-shorelines (limits of shaded area) and ice limits/heights. (Time expressed as calibrated yrs BP). (Taken from Lambeck, 1996)

found to change this component of the model. Instead, in order to generate high Late Devensian RSL's around the entire Irish Sea Basin, (substantial) changes are made to the Irish ice sheet. The geophysical model predictions delivered by Lambeck (1996) are based on an ice model that is no thicker than 700m over Ireland at the time of the LGM (Figure 2.8) yet in order to reconcile the findings of Eyles and McCabe (1989), a two to three fold increase in ice thickness over central and southern Ireland is required (Lambeck and Purcell, 2001). These modifications to the ice model may only be applied over central/southern Ireland or else model RSL predictions for the west coast of Britain will contradict available (Holocene) RSL evidence. Furthermore, these changes result in the ice surface height over Ireland becoming higher than over Scotland. The above leads the authors to the conclusion that "if such an ice model is extreme and incompatible with glaciological observations, then the alternative is to accept the interpretation of the Irish Sea Drift as terrestrial (and not glacio-marine) in origin" Lambeck and Purcell (2001), pp 497.

Consideration of how vertical adjustments in the level of the land with respect to the sea bring about change in palaeogeography is an important facet of studies considering patterns of past sea-level change. In this region, the possible emergence of a LGM/postglacial land bridge between Ireland and Britain has been of especial interest. For some time discussion has been prevalent in Irish Quaternary literature (e.g. Devoy, 1983, 1985, 1995; Plag *et al.*, 1996; Sleeman *et al.*, 1986) with the topic of plant colonization in particular instigating the suggestion of a land-bridge to Ireland (Mitchell, 1972). Landbridge development is particularly sensitive to the delicate interplay of glacio-isostacy, eustacy and coastal hydrodynamic factors. These components all operate together to produce conditions suitable for a landbridge over very short time-scales (10^1-10^2 years) meaning the former existence of a landbridge is far from an empirical certainty (Devoy, 1995; 2000). Whilst evidence for the presence of a landbridge in the geological record remains elusive, the most recent geophysical modelling results of Lambeck (1995; 1996) and Peltier *et al.* (2002) all suggest land bridges existed at some point in the Late Devensian. Reference to Lambeck (1996) suggests such a feature was operational between 21 000 and 18 000 BP and absent by 15 000 BP (Figure 2.8) and it appears likely that the most probable location for such a connection would have been between Cornwall and Waterford. The modelling results of Peltier *et al.* (2002) also suggest the presence of a Celtic Sea landbridge although landbridges are also suggested to have occurred across both the Irish Sea and North Channel. As with the findings of Lambeck (1996), these landbridges were also said to exist between 21 – 18 000 BP with their southerly limit migrating



Figure 2.9 Palaeogeographic reconstructions taken from the GRM work of Peltier *et al.* (2002). The colour coded regions that cover the surface of the ocean in the area surrounding the British Isles represent the migration of the coastline subsequent to the LGM

north through time (Figure 2.9). Differences between the palaeogeographic reconstructions of Lambeck (1996) and Peltier *et al.* (2002) can largely be explained in terms of the different adopted Earth models, which in turn impact upon GIA histories. In addition, changes in the thickness of extra regional ice cover has also been shown to have an impact on the development of landbridge(s) between Britain and Ireland: part of the British Isles lies within the peripheral bulge zone of Scandinavia and a reduction in ice volume over Scandinavia increases the former water depths hence reduces the geographical extent of any landbridge (Lambeck and Purcell, 2001). Both Mitchell (1976) and Devoy (1985) also consider the North Channel to be the most likely location for a Lateglacial landbridge although both these studies along with the later work of Preece *et al.* (1986) and Wingfield (1995) suggest a landbridge was still operational after 15 000 BP. Mitchell (1986) argues for its existence as late as 7500 BP on the basis of the continued arrival of freshwater taxa to Ireland until this date. However, such an assertion is extremely hard to reconcile this observational evidence suggesting that postglacial eustatic sea-level rise was virtually complete by 7 000 BP (e.g. Peltier, 2004).

2.4 Early-Mid Holocene RSL Changes

At the opening of the Holocene, eustatic sea level was still rising quickly at c. 7-8 mm yr^{-1} (e.g. Fairbanks, 1989, Bard *et al.*, 1990a, 1996; Cutler *et al.*, 2003). This background sea-level rise aligned with ever decreasing rates of glacio isostatic adjustment led to marine incursion around the coast of Ireland. In the northeast of Ulster, Carter (1982a) suggests that sea level began rising from its Lateglacial lowstand position shortly before 11 000 BP; an assertion made on the basis of two radiocarbon dates derived from submerged peat deposits around -10m OD(B) in Belfast Lough (Manning *et al.*, 1970; Pearson and Pilcher, 1975). However, the relationship between these peat deposits and sea level is unclear and they may not necessarily provide direct evidence of rising sea level at this time. Along the north Ulster coastline, the lack of data from the time of the lowstand makes bracketing the onset of sea-level rise problematic. Yet by 10 000 BP, Carter and Wilson (1990) suggest sea level had risen to around -7m OD(B) on the basis of radiocarbon and palynological evidence from a (submerged) peat unit in the Bann Estuary.

Elsewhere around the north east of Ireland, a number of radiocarbon dates provide both direct (and indirect) evidence of continual sea-level rise throughout the early Holocene. Probably the best evidence for this is derived from the litho- and biostratigraphic work of Singh and Smith (1973). This work from Lecale, Co. Down, considers a series of marginal marine basins ranging in altitude from -0.7m to 5.6m OD(B) which are used to infer fluctuations in sea level throughout the Holocene. Only the basin with the lowest threshold experienced marine inundation during the Holocene, with the first evidence for this occurring shortly after 7 500 BP. These findings are broadly in agreement with those of Harmon (2006) who uses a microfossil based transfer function along with AMS radiocarbon dating to develop sea level index points from just below present at 8791±188 and 7782±107 cal. BP. Along the east coast of Ireland, evidence for the position of sea level during the early-mid Holocene is almost entirely absent. The only exception to this is the work of both Mitchell (1976) and Harris (1979a) from Bray and Dublin Bay which together suggest sea level stood around -3.5m OD(B) at 7500 BP. This lack of accurate sea level data from the east coast of Ireland coupled with the paucity of long time-series records has inhibited detailed comparisons with sea level observations from the eastern side of the Irish Sea Basin.

After about 8 000 BP, the rate of eustatic sea-level rise declined towards, and eventually fell below the isostatic rate culminating in the development of Holocene highstands in the north east and east of Ireland. The magnitude and timing of the highstand event changes in both time and space as a result of variation in the rate of glacio-isostatic rebound. Recently, Simms (unpublished data) has suggested that along the Antrim coastline, Holocene sea level rose to around 10m above present, far higher than the +4m OD(B) maximum altitude inferred by Carter (1982a). This assertion is based on AMS radiocarbon dates associated with fossil limpet remains. However, aside from uncertainties pertaining to the local marine reservoir effect, it is unclear whether these fossil marine organisms can be related to former sea level in the manner they have been by the researcher. Putting to one side the work of Simms, nowhere else in the north of Ireland is there the quantity or quality of data to accurately define the height or timing of the highstand. Rarely even, is it possible to infer a maximum altitude for the highstand although in Dundalk (south east Ulster), it is unlikely to have exceeded +2m OD(B) (Penney, 1983). This value is commensurate with newly acquired sedimentary records from the Isle of Man which show that RSL rose from around -3.80 m below

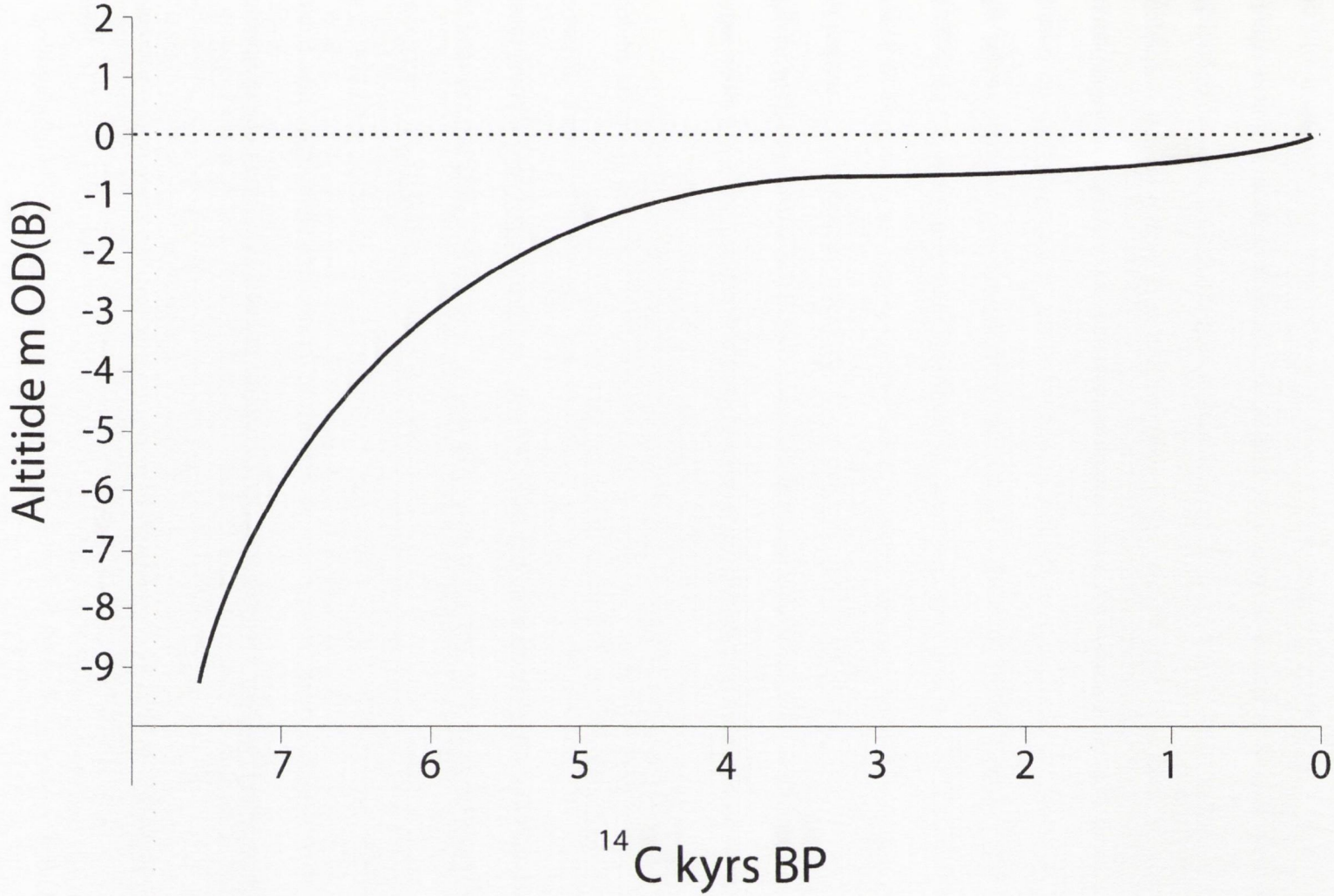


Figure 2.10 Relative sea level 'curve' for the northwest of Ireland (Donegal). (Taken from Shaw and Carter, 1994)

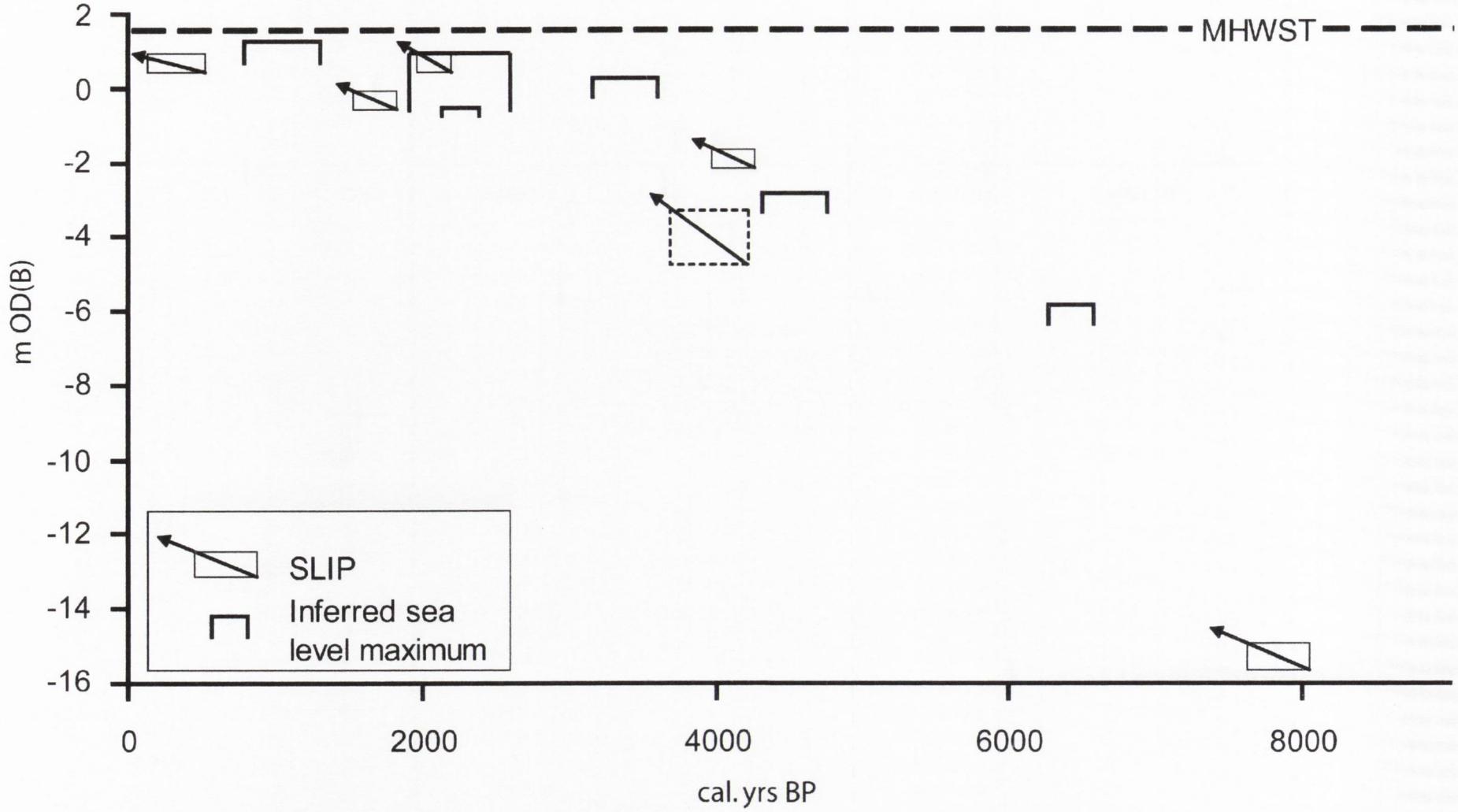


Figure 2.11 Relative sea level age-altitude graph for south and southwest Ireland. (Taken from Carter *et al.*, 1989)

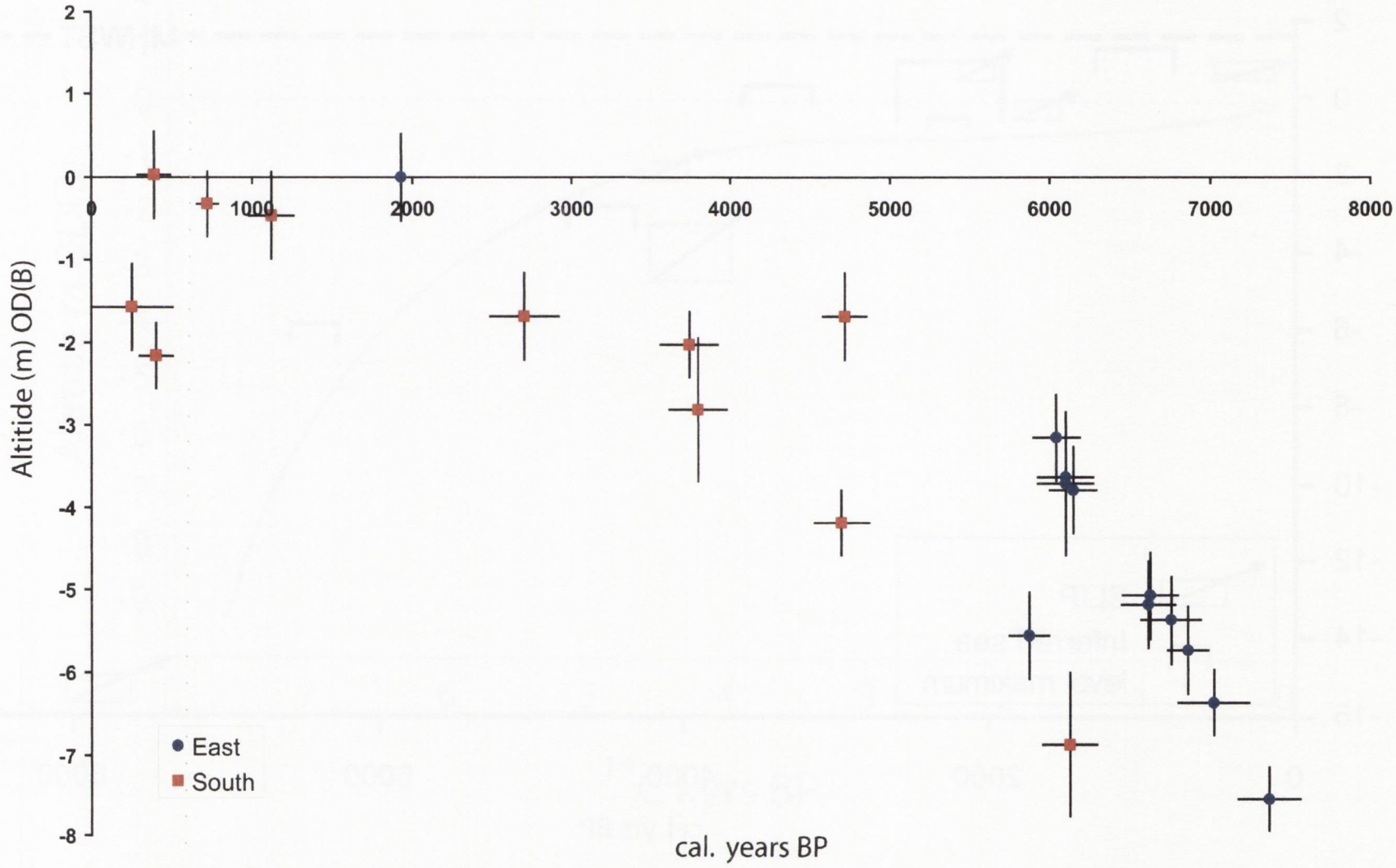


Figure 2.12 Relative sea level age-altitude graph for the south and south east of Ireland based on the SLIPs presented in Sinnott (1999)

present in the early Holocene up to a mid-Holocene maximum between 6600 and 4500 BP at c. 2.25 m above present (Roberts *et al.*, 2006).

To the west, Shaw (1985) has looked into patterns of Holocene sea-level change in Donegal. Sea level index point evidence testifies to a similar trend of rising RSL from an (unknown) Lateglacial lowstand depth to the mid Holocene. Shaw (1985) suggests that in northeast Donegal, ongoing glacio-isostatic rebound was of sufficient strength to induce a Holocene maximum sea level above present. This concurs with the findings of Wilson and McKenna (1996) who present sedimentological evidence for a Holocene highstand of around 2-3m OD(B) magnitude at 6500 BP in the River Bann Estuary, less than 30 kilometres to the east of Donegal. However, in southwest Donegal Shaw (1985) uncovers no evidence for higher than present sea level and this finding is strongly supported by the recent microfossil based transfer function work of both Gehrels (2004) and Harmon (2006) from west Donegal. This overall pattern appears to form a broad continuum with the work of Carter (1982a) with an east to west gradation in the magnitude (and to a lesser extent timing) of the Holocene maximum becoming evident across the north of Ireland. Later work by Shaw and Carter (1994) suggests that higher than present sea levels were only experienced to the east of Lough Foyle and not in Donegal (Figure 2.10). However, given the paucity of data and the nature of the sea-level indicators used by both Shaw (1985) and Shaw and Carter (1994), it is debatable whether it can be proved either way that Holocene sea levels above present existed in north east Donegal.

Although patterns of early-mid Holocene sea level in the north and east of Ireland have been studied in some detail, up until recently the south and particularly southwest of the country had attracted little attention from sea-level researchers. However, the work of Carter *et al.* (1989); Sinnott (1999) and Devoy *et al.* (in press) has necessitated a revision of this statement. These studies have all employed the sea level index point methodology to reconstruct RSL in this region and it can now be said that patterns of Holocene sea level are probably best understood in this part of the country.

Both Carter *et al.* (1989a) and Sinnott (1999) have shown that during the early Holocene, southern Ireland was characterised by rapid rates of sea-level rise (Figures 2.11 and 2.12); a sea level index point from Curlane Bank off Cork Harbour places sea level at around -15m OD(B) at 9204±75 BP

(Q2379) although by the mid Holocene, RSL stood at nearer to -3m OD(B). Sinnott (1999) also finds a similar pattern of change in the southeast, with litho- and biostratigraphic evidence demonstrating continuous sea-level rise from at least 7500 BP (Figure 2.12). Rates of RSL rise between 7500 to 6000 BP are in the region of 3-4mm yr^{-1} with a reduction in this rate occurring in both the south-eastern and southern regions after this time. Interestingly, the work of Sinnott (1999) provides tentative evidence for differential movements of Holocene sea level in southern Ireland. This study suggests that in comparison with sea level index points from the south/southwest, those to the southeast appear to reveal a consistent positive vertical displacement, especially before about 6000 BP. As Devoy 2000; pp 4 comments, these "southeast to west height differentials in RSL index points may provide evidence for the operation of distinct regional differences in Earth crustal behaviour. Either the south – southwest of Ireland has experienced crustal subsidence, at least during the early to mid Holocene, or the southeast has undergone low scale relative uplift". On the basis of the rebound modelling studies of Lambeck (1996), such vertical differences should be evident in the geological record. Most recently, Devoy *et al.* (in press) have presented a suite of Holocene sea level index points from Timoleague and adjacent estuary sites in Courtmacsherry Bay which date back to 6380±100 BP. These new data reveal a rise in RSL form -5.5m OD(B) at this time to present and are in agreement with the findings of Sinnott (1999). However, despite this later work from Devoy *et al.* (in press), the earlier assertion of Sinnott (1999) still holds: Many more SLIPs from Ireland's south coast covering the entire Holocene period are required before any firm conclusions regarding regional crustal movements can be drawn.

2.5 Late Holocene RSL Changes

By the mid/late Holocene, global eustatic sea-level rise had slowed greatly from the rates experienced in the early Holocene. Even though there may have been as much as 3m of eustatic sea-level rise since 7000 BP (Lambeck, 2002), rates of both eustatic and isostatic change were less than at any point since the LGM. Generally, the assertion of Devoy (1983) that patterns of mid to late Holocene sea-level movement are obscure in Ireland still holds yet more recent studies involving detailed litho- and biostratigraphical evidence (e.g. Carter *et al.*, 1989a; Sinnott, 1999; Devoy *et al.*, in press; Gehrels, 2004; Harmon 2006) means this statement can now be at least partly revised. In the north of Ireland, several studies have looked into patterns of late Holocene

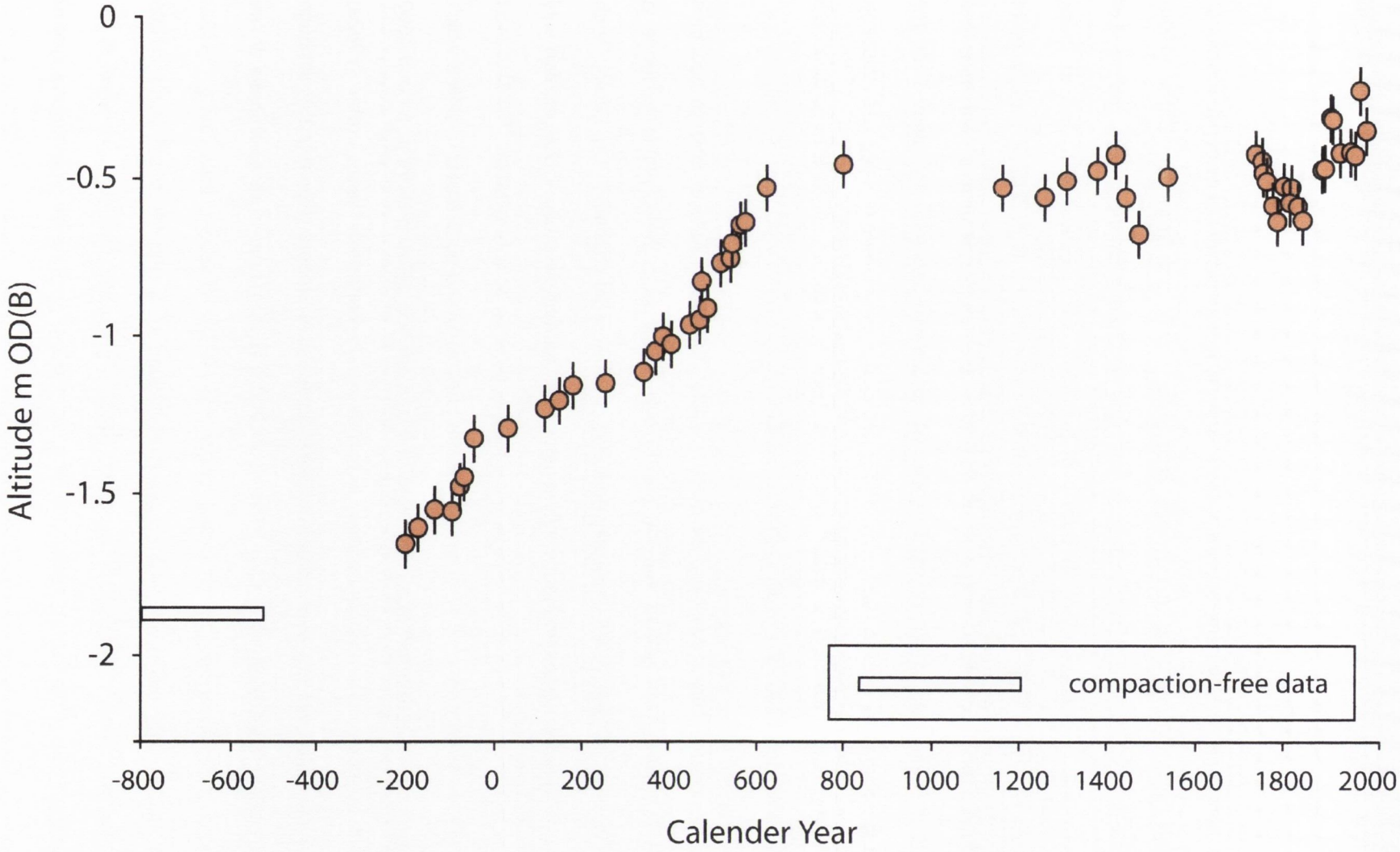


Figure 2.13 Relative sea level age-altitude graph from Bracky Bridge, (SW Donegal) for the late Holocene. (Taken from Gehrels, 2004)

sea-level change with Donegal attracting particular attention. Shaw (1985) developed a suite of SLIPs from 15 study sites around the Donegal coastline. In west Donegal, these data evidence sea-level rise throughout the late Holocene, rising from around -2m OD(B) at 4700 BP. In northwest Donegal, sea level is suggested to have been nearer to -1m OD(B) at this time whilst in north east Donegal, evidence is found for sea-level fall during the late Holocene. However, as previously mentioned, the later work of Shaw and Carter (1994) questions the evidence for higher than present Holocene sea level in Donegal. Both Shaw (1985) and Shaw and Carter (1994) accept that a number of more minor oscillations may have occurred within the generalized trend of sea level but the comparatively crude nature of the sea-level indicators employed in these investigations precluded their detection. This is not the case with the high precision analyses of Gehrels (2004) from southwest Donegal which detects RSL changes operating on sub-millennial and decimeter scales and identifies two episodes of accelerated sea-level rise since 2150 BP (Figure 2.13). The nearby work of Harmon (2006) supports the overall trend of rising late Holocene sea level in this locality although is unable to confirm these episodes of enhanced sea-level rise between 2150 BP and 1950 BP and 1550 BP to 1250 BP.

Elsewhere around the Ulster coastline, evidence for late Holocene sea-level change becomes more equivocal. Whilst the overall trend is one of falling sea level, the available evidence is insufficient to estimate rates of fall or identify episodes of accelerated change. In Co. Down, Singh and Smith (1973) have suggested marine influence at an altitude of over 2.5m OD(B) existed until as late as 3210 BP yet the evidence used to infer this may be seen as controvertible. No data exist to support the late Holocene trend envisaged by Carter (1982a) for the north Ulster coastline which sees sea level falling from a mid Holocene highstand to a position below present at around 4500 BP followed by rising sea level thereafter (Figure 2.3a). In northeast Ulster Carter (1982a) proposes a late Holocene sea-level history almost entirely above present (Figure 2.3b), although geomorphological evidence presented by Orford *et al.* (2003) suggests the regressive phase of sea level following the Holocene highstand lasted longer into the late Holocene than Carter (1982a) envisaged. Additional geomorphological evidence in support of a general falling trend in late Holocene sea level exists in the form of numerous swash aligned beaches from Magilligan at the mouth of Lough Foyle (Carter *et al.*, 1982a; Carter *et al.*, 1989a). Their development at the time of the Holocene highstand and seaward propagation since has been related to falling sea level although beach ridge development often reflects environmental parameters other than movement

of RSL. Present day tide gauge records from Belfast in north east Ulster suggest crustal response to loading is not yet complete (Woodworth and Jarvis, 1991; Devoy, 1999).

On the East coast, patterns of late Holocene sea level are also poorly recorded. Synge (1980a, 1980b) documents the existence of three isostatically driven shorelines dated to c. 7300, 4500 and 2500 BP respectively. These are tilted towards the south, reflecting an apparent decrease in isostatic uplift. However, much of the geomorphological evidence used in these studies falls within the vertical range of contemporary shoreline processes and thus provides only equivocal evidence of changing mid-late Holocene sea level (Carter, 1982a). Elsewhere along the east coast, Mitchell (1977); Mitchell and Stephens (1974) and Donner and Jungner (1980) all provide radiocarbon dates from both peat horizons and archaeological remnants from above present sea level, dating back to the mid Holocene. However, overall this evidence contributes little to our understanding of sea-level change during this time period.

In the southeast, the work of Carter *et al.* (1989a) and the more recent evidence of Sinnott (1999) suggests continued RSL rise throughout the Late Holocene and up until present day. Rates of RSL rise in this region are suggested to be less than 2mm yr^{-1} for the Late Holocene, reflecting the combination of minimal eustatic rise along with (marginal) crustal subsidence (Devoy, 1999). Along the south coast, the work of Carter *et al.* (1989); Sinnott (1999) and Devoy *et al.* (in press) evidences a similar pattern of late Holocene sea-level change. SLIPs from Cork Harbour and Dugarven Bay (east Cork), along with further data from Timeoleague and Rosscarbery Bay (west Cork) suggest sea level was a little below -2m OD(B) at 3000 BP. In the southwest, the pattern of regional coastal change during the late Holocene is also one of persistent submergence of wetlands under continuing RSL rise (Devoy, 2000). The most complete record of Late Holocene RSL change in this region is derived from the work of Carter *et al.* (1989a) and Delaney *et al.* (in press) from the Dingle Peninsula with SLIP data evidencing sea-level rise at a rate of around 0.45 mm yr^{-1} over the past 5000 years. Along Ireland's west coast, precise evidence for late Holocene sea-level movement is unavailable. This holds, despite a number of studies presenting radiocarbon dates from (freshwater) peat deposits which demonstrate coastal submergence during this period (e.g. Mitchell, 1976; Pearson, 1979; O'Connell, 1988).

2.6 SUMMARY

- The evidence for Holocene sea-level change in Ireland is fragmentary, but still offers a broadly coherent and explainable pattern (Carter *et al.*, 1989a).

- In the north east of Ireland, RSL histories appear highly non monotonic as a result of variations in the strength of the isostatic signal relative to the eustatic signal. Much of the Lateglacial was characterized by falling sea level in response to rapid isostatic rebound after ice retreat (e.g. Carter, 1982a; Lambeck, 1996). At the start of the Holocene, RSL rose from a lowstand position to a level above present as a result of isostatic recovery outpacing eustatic sea-level rise. Nowhere around Ireland is the magnitude or timing of this highstand well constrained.

- In the south of Ireland, little is known about the position of sea level at the time of the LGM or throughout much of the Lateglacial (e.g. Devoy, 2003). However, knowledge of sea-level change during the Holocene is far better understood as a result of SLIPs generated by a number of more recent studies employing recognized sea-level methodologies (e.g. Carter *et al.*, 1989a; Sinnott, 1999; Devoy *et al.*, in press; Delaney *et al.*, in press).

- Relative sea-level research in Ireland has been heavily biased to the north and south coasts, with little evidence presented from either the Irish sea coast in the east or Atlantic margin to the west.

- Around Ireland's coasts, rates of RSL change during the late Holocene have commonly been ≤ 1mm yr^{-1} on average. Variations in this figure occur though, primarily as a result of continued crustal response to deglaciation (Devoy, 1999). Broadly in the north east of Ireland, coastal areas continue to record land emergence today, with areas to the south and west of this zone showing relative land stability to net subsidence (Woodworth and Jarvis,1991; Devoy, 1999).

- The biggest unresolved debate in the Irish sea-level community concerns the nature of sea-level change coming out of the LGM and during the Lateglacial. Two contrasting models have been proposed: The first is based on numerically based rebound modelling and only predicts

higher than present sea level in north and northeast Ireland. The latter is based upon dated glacio-sedimentary sequences which may indicate fluctuating, higher than present sea level around most of Ireland's coastline (e.g. McCabe *et al.*, 2005).

- Several attempts have now been made to predict postglacial RSL for locations around the coast of Ireland, the most comprehensive of these being the geophysical modelling analyses of Lambeck (1996). The crude pattern for the north and south fits the basic model of global sea-level change predicted by Clark *et al.* (1978), which places Ireland on a hinge line between the highly glaciated north of Europe and the generally ice free area to the south. However, detailed agreement between the latest model predictions of Lambeck (1996) and the observational data varies from satisfactory to poor (Sinnott, 1999).

CHAPTER THREE

Methodology

3.1 THE NATURE OF SEA LEVEL DATA

Over thirty years of international research, coordinated under the auspices of a series of International Geological Correlation Programme (IGCP) projects, has produced a well-defined methodology for developing records of RSL change from sedimentary coasts (Edwards, 2005). These sediment archives which are commonly found in low energy environments such as estuaries, lagoons and tidal marshes contain a rich source of information on past changes in RSL (e.g. Godwin 1940, 1943, 1945; Godwin and Godwin 1940; Tooley 1978). Combined litho-, bio- and chronostratigraphic examination of the evidence contained with these sequences may be used to produce 'sea level index points' (SLIPs) (e.g. Preuss 1979; Shennan *et al.*, 1983; van de Plassche 1986) which can then be utilised to provide a temporal and spatial framework for RSL change and coastline position (Jennings *et al.*, 1995). Detailed consideration of SLIPs and their associated error terms are given in a series of publications (e.g. Kidson and Heyworth, 1979; Heyworth and Kidson, 1982; Shennan, 1982; 1986) and the tenets of this sea level index point methodology are outlined in Figure 3.1. For a sample to be established as a SLIP it must possess information regarding its location (latitude and longitude), its altitude (relative to a levelling datum), its age (commonly inferred from radiocarbon dating), and its vertical relationship to a contemporaneous tide level (termed the indicative meaning) (Figure 2.1). The latter is important to account for the differing vertical distributions of coastal sub-environments and associated sea-level indicators. The sea level index point methodology rests upon the principal that variation between terrestrial and marine sedimentation reflects change in the balance between land and ocean levels, moderated by local scale processes such as the rate of sedimentation (Van de Plassche, 1986).

Sea level index points are commonly derived from lithostratigraphic contacts between terrestrial and marine sediments. Alternations between these freshwater peats, organic-rich sediments and marine clastic sediments are the most common expression of changing RSL in the submerged coastal environments of the British Isles and such sequences are most regularly encountered

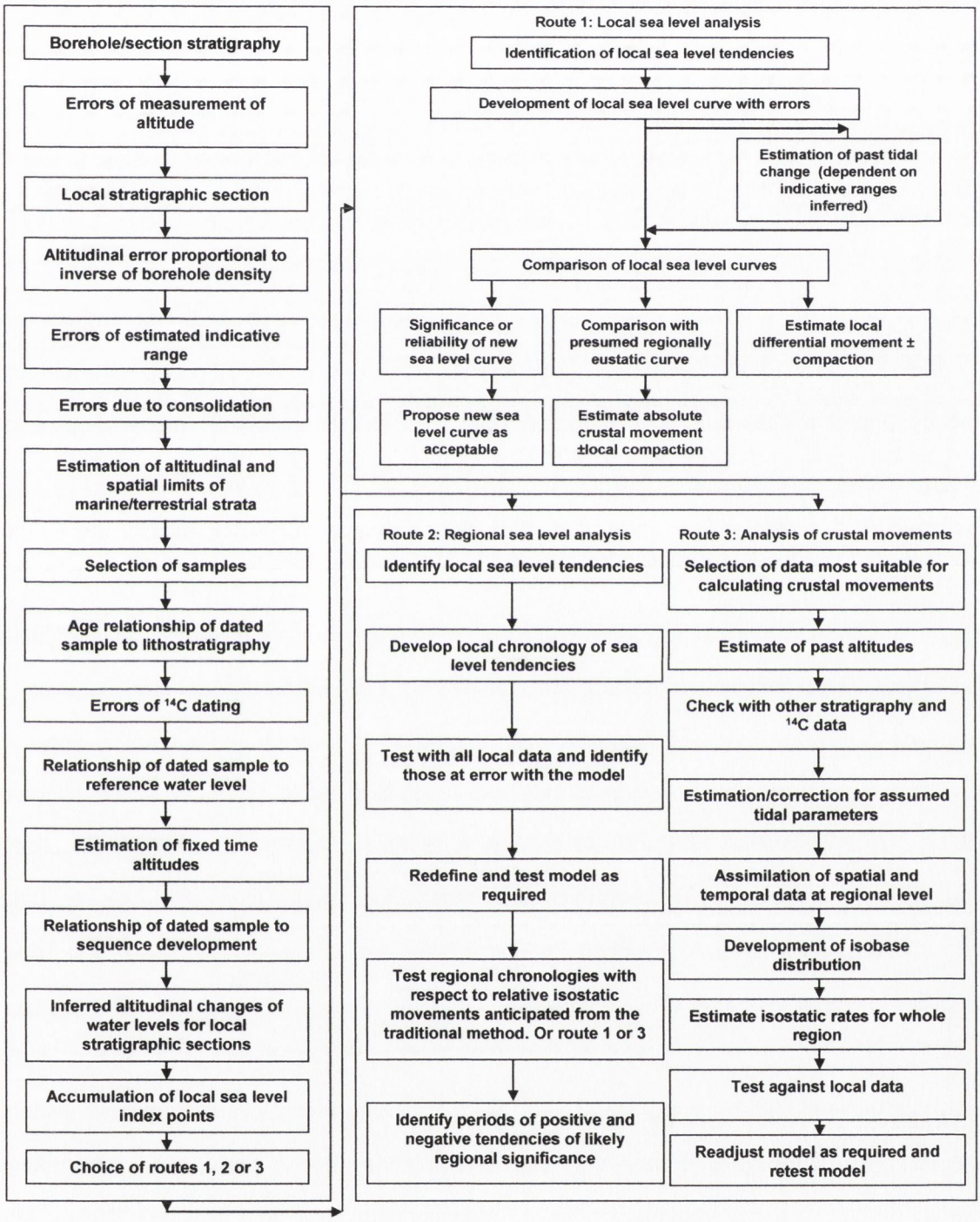


Figure 3.1 Expanded model of sea-level research methods with explicit allowances for errors. (Taken from Shennan, 1983)

beneath contemporary salt-marshes (e.g. Devoy, 1979; Heyworth and Kidson, 1982; Gehrels, 2004; Edwards, 2006a). Since the completion of IGCP projects 61 and 200, the recognition of sea-level changes based upon these lithostratigraphic transitions between organic and minerogenic sedimentary units has formed the basis of sea-level research in the UK. The focus on a restricted set of depositional environments came about because of the need to establish an indicative meaning (derived from the nature of the lithostratigraphy above and below a contact) and the requirement of sufficient organic material for radiocarbon dating (Edwards, 2006b). A change in lithostratigraphy from an organic-rich horizon to a clastic sedimentary unit is referred to as a transgressive contact whilst the reverse of this is termed a regressive contact. The former implies an increase in marine conditions whilst a regressive contact is indicative of a decrease in marine influence. Both terms are descriptive and are not synonymous with actual movements of sea level.

Shennan (1982) considered a range of modern flora-environment relationships in the Fenland coastal zone which permitted the quantification of the indicative meanings for lithological boundaries. The established approach assumes, (where no hiatus is present) that intercalated sedimentary facies formed in environments that existed adjacent to one another in space. By quantifying the range of elevations across which the transition between one sub-environment and another occurs, it is possible to assign an indicative range to a lithostratigraphic or biostratigraphic contact (van de Plassche, 1986). These findings from the Fenland have since been applied to similar coastal contexts from elsewhere and have underpinned many lithologically based reconstructions of sea level from around the British Isles (e.g. Shennan, 1986; Zong and Tooley, 1996; Sinnott, 1999; Shennan *et al.*, 2000d).

Lithologically-based sea-level reconstructions have traditionally employed microfossil evidence (such as pollen, diatoms and foraminifera) to confirm or refine interpretations based upon changes between organic and minerogenic modes of sedimentation (van de Plassche, 1986). More recently, a number of sea-level researchers have developed statistical based methods to reconstruct indicative meanings founded upon the relationship between the inter-tidal faunal assemblages themselves and elevation with respect to the tidal frame (e.g. Horton *et al.*, 1999; Zong and Horton, 1999; Gehrels *et al.*, 2001; Horton and Edwards, 2005; 2006; Horton *et al.*, 2006). These techniques have the potential to deliver higher precision reconstructions of water level than previous lithologically-based methods (e.g. Horton and Edwards, 2006) and represent a

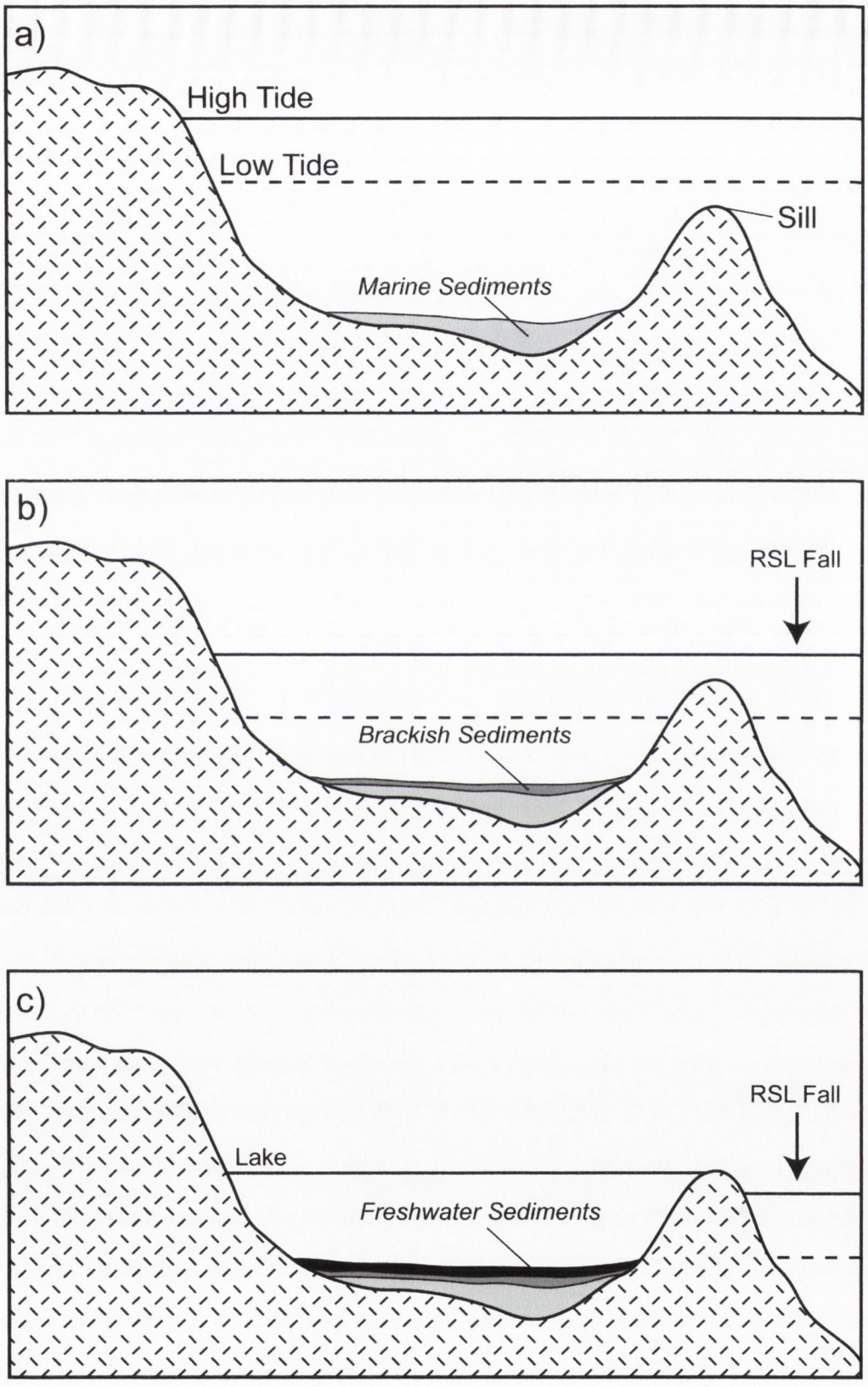


Figure 3.2 An illustration showing the isolation of a marginal marine basin by falling RSL. a) The basin is inundated at all stages of the tide and marine sediments accumulate. b) The basin is isolated from the sea at low tides and brackish conditions are encountered. c) High water is now below the height of the sill and the basin becomes a freshwater lake. (Taken from Edwards, 2006b)

useful tool in the search to elucidate the relationship between ocean levels and climate at the sub-millennial scale (Edwards *et al.* 2004b). Furthermore, they have expanded the range of environments from which sea level information may be extracted and may be used in conjunction with advanced AMS dating techniques to deliver SLIPs from clastic sedimentary horizons (Horton *et al.*, 1999; Edwards and Horton, 2000).

In situations where a dated sample does not possess an indicative meaning (e.g. a freshwater peat deposit or a sub-tidal foraminiferal assemblage), the sample cannot be used to precisely fix the former altitude of RSL. Instead, such samples may only provide limiting information for the maximum (or minimum) position of former sea level. In certain situations, samples collected from freshwater (or wholly marine) settings can provide important data that may be employed to test specific hypotheses related to sea level change and coastal evolution. This is because these environments must have either formed inland (or seaward) of the palaeo-coastline and above (or below) the height of palaeo sea level (Shennan *et al.*, 2002).

In the north of the British Isles, much of the postglacial sea-level story has been acted out above present level and this has resulted in the manifestation of markedly different coastal stratigraphies to those that have developed in submergent coastal settings. Whilst (raised) salt-marsh deposits have been used to deliver SLIPs from uplifted regions, a more common approach has been the use of isolation basin deposits to fix the former position of sea level (e.g. Shennan *et al.*, 1993, 1994, 1995b, 1996b, 1999, 2005, 2006b). An isolation basin is a naturally enclosed rock basin that was once below sea level but has subsequently been raised above it, and hence become isolated from the sea (Lloyd and Evans, 2002) (Figure 3.2). This isolation procedure is recorded in the litho- and biostratigraphy by a transition from marine to freshwater sedimentation. In order to establish a SLIP from an isolation basin sequence, a series of microfossil samples are taken from across the organic-minerogenic sedimentary transition found in the isolation basin and the altitude at which microfossils indicative of brackish conditions are completely replaced by freshwater species (or *vice versa*) is dated (Figure 3.3). However, instead of using the altitude of the sediment sample, SLIPs from isolation basins are constructed by using the altitude of the lowest part of the rock sill separating the basin from the open sea. This isolation basin approach to reconstructing former sea level is now well established and is said to be the most reliable procedure for constraining past sea level in glacially uplifted areas (Eronen *et al.*, 2001).

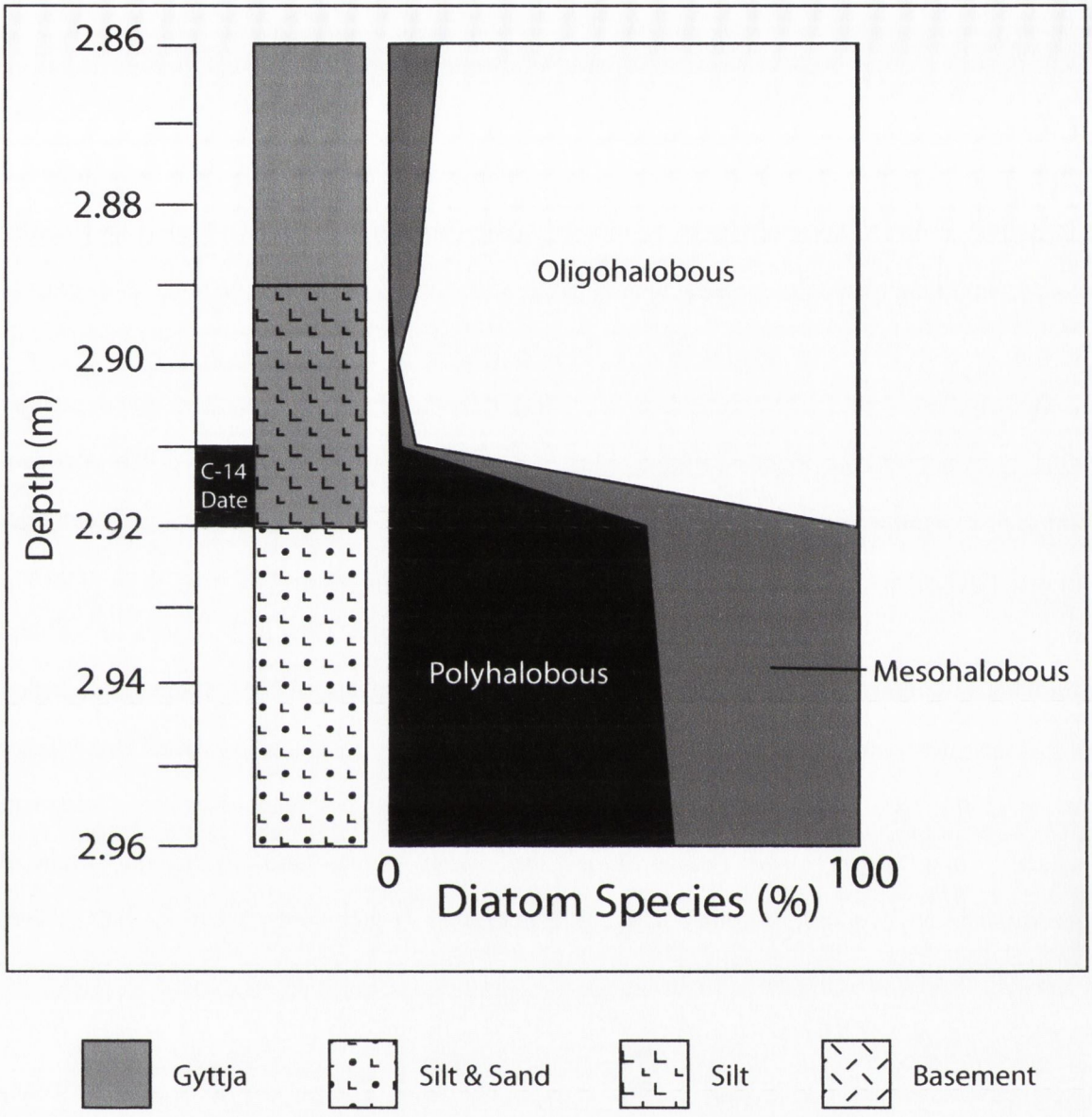


Figure 3.3 A simplified stratigraphic column from an isolation basin plotted alongside a summary of the diatom analysis which reveals the transistion from polyhalobous species (indicative of marine conditions) to oligohalobous species (associated with freshwater conditions). (Taken from Edwards, 2006b)

3.2 Low Energy Sedimentary Archives

3.2.1 Salt-marshes and Inter-tidal Flats

Salt-marshes are areas of land covered chiefly by halophytic vegetation which are regularly flooded by the sea. Together with wetlands and high inter-tidal flats, they make a major environmental contribution to the lowland coasts of Northwest Europe and occur chiefly on open and barrier coasts and in estuaries and embayments (Allen, 2000a). Ecologically, salt-marshes provide habitats for a wide range of transitional marine organisms and provide a refuge for many commercially important taxa (Montague *et al.*, 1981). Inter-tidal flats are comprised of fine grained clastic sediment and are found seaward of the salt-marsh system. These tracts of the inter-tidal zone are unvegetated and are only able to support algal mats. Both salt-marshes and inter-tidal flats are genetically related as the accumulation of sediment elevates inter-tidal surfaces, reducing the depth and period of tidal inundation. When they become sufficiently high relative to the tidal frame, these surfaces may support (salt-marsh) vegetation (Edwards, 2006b). Sediment accretion on the salt-marsh surface is further enhanced by the presence of vegetation which slows water movement and encourages sedimentation. The amount of accretion is both a function of sediment availability and tide or wave energy across the marsh surface and accordingly, rates of sedimentation vary greatly between different depositional environments (Pethick, 1981; Allen, 1990a,b,c; French and Spencer, 1993).

The deposits beneath the (active) salt-marshes of northwest Europe are dominated by tidally introduced mineral matter; these marshes may be described as mineralogenic, with the organic sediment component rarely higher than 10% dry weight (Allen, 2000a). Mineralogenic salt-marsh development has dominated during the late Holocene period (e.g. Devoy, 1979; Allen, 1991; Beets *et al.*, 1992; Shennan, 1994; Brew *et al.*, 2000; Long *et al.*, 2000; Baeteman *et al.*, 2002; Behre, 2004) although at various times during the mid and early Holocene however, conditions were such that salt-marsh peats formed in these coastal settings. These marshes, incorporating only modest amounts to no tidally introduced sediment, may be described as 'organogenic' and range(d) in altitude from HAT to below MHWNT (Allen, 2000a).

Curtis and Sheehy Skeffington (1998) have developed an inventory of the salt-marshes of Ireland, categorising their geomorphology and origin and placing them into one of five main groups;

estuary, bay, sand flats, lagoon and fringing marsh. Over 200 marshes have been documented and their botanical status has been outlined in Sheehy Skeffington and Curtis (2000). Because of restrictions in mineral sediment supply, coupled with the greater maritime influence on the Irish climate, Irish Atlantic seaboard marshes differ in character with other northwestern European sites (Burd, 1989; Allen, 2000a). As well as containing a higher percentage of organic matter, they lack a clear halophytic zonation with freshwater and euryhaline plants found across most of the marsh surface (Sheehy Skeffington and Wymer, 1991, Duffy and Devoy, 1999; Sheehy Skeffington and Curtis, 2000).

The lithological and biological characteristics of salt-marsh environments and inter-tidal flats are closely related to the tidal frame. As elevation within the tidal frame decreases (hydroperiod increases), environmental conditions alter, and plant and animal communities undergo change until almost marine conditions are reached at the lowest tidal levels (Edwards, 2006b). This vertical zonation in relation to tide level means that RSL information can be extracted from both the inter-tidal sediments themselves and the floral and faunal remains contained within them. Salt-marsh deposits have also been shown to form part of a seral succession of wetland vegetation (Figure 4.4). Pollen analysis has traditionally been used to evidence this succession which has then been used as a record of tendencies of RSL movement (e.g. Tooley, 1978; Devoy, 1979; Shennan, 1986).

Many authors have used salt-marsh 'peats' found at transgressive and regressive contacts in the stratigraphic record to develop sea level index points. However, the reliability of these SLIPs developed from salt-marsh peat has been questioned, perhaps most prominently by the numerical modelling analyses of Allen (1990c; 1995). As Allen, (1995) pp 42 notes:

> *"Because of the sensitivity of the facies to forcing factors, combined with lag effects, substantial intrinsic uncertainties far exceeding observational errors may be expected in sea-level curves constructed from radiocarbon-dated (salt-marsh) peats. Chronologically, the ages of conventional transgressive and regressive overlaps poorly constrain the timing of sea-level fluctuations. In the experimental marshes, the creation of organic-rich facies depends on forcing factors, and occurs at a variety of elevations relative to the tidal frame, generally well above MHWST."*

The analyses of Allen suggest that the indicative meaning of a lithostratigraphic salt-marsh unit may vary in both time and space and therefore cannot be seen as a precise and reliable indicator of sea level. Acknowledging the limitations associated with this lithologically-based approach to reconstructing the indicative meaning, more recent studies have focused on quantitative analysis of the faunal and floral sea-level indicators contained within salt-marsh sedimentary facies as a means of reconstructing the environment in which the sediment unit was deposited (Edwards, 2006b).

3.2.2 Isolation Basins

Isolation basins have been most extensively used in the reconstruction of RSL histories in regions formerly under the Laurentian and Scandinavian ice sheets, with work carried out in Greenland (e.g. Foged, 1977; Long *et al.*, 1999a, 2006), Canada (Retelle *et al.*, 1989) and Scandinavia (e.g. Faegri, 1940; Kjemperud, 1986; Eronen *et al.*, 2001). Recently however, the isolation basin approach has been employed in the north of the British Isles to develop SLIPs which may be used to elucidate patterns of glacio-isostatic rebound (e.g. Shennan *et al.* 1995b, 1996b, 2000a, 2006b).

Traditionally, much of the RSL research undertaken in areas affected by postglacial uplift has focused on evidence from the geomorphological record, mainly in the form of strandlines and marine notches which have been employed in the construction of shoreline-altitude graphs. In Ireland, much of this type of work was undertaken from the mid-1960's up until the late 1970's (e.g. Stephens and Synge, 1965; Synge and Stephens, 1966) and followed closely in the footsteps of the pioneering work of Sissons from the UK. However, as previously mentioned in chapter 2, one may be circumspect of such evidence for several reasons. Firstly, most of the raised shoreline features have no quantifiable indicative meaning and may form over a broad indicative range, spanning from MSL to well above HAT. Moreover, despite widespread morphological evidence of raised shorelines, there are few absolute age controls on the features. Instead, ages for the features have been inferred indirectly through their stratigraphic relationship to dated sedimentary horizons.

Studies of isolation basins represent part of a general advance in work on RSL change in the UK and have delivered significant improvement in terms of both precision and accuracy over data

sourced from the geomorphological record (Laidler, 2002). Consideration of the stratigraphic record within marginal marine basins around the Ulster coastline thus appears to represent the most logical pathway towards a detailed understanding of the postglacial sea-level record in this region. In the north of Ireland, rates of postglacial crustal rebound will be more comparable to those experienced in northern Britain than to areas once covered by the (large) Laurentian and Fennoscandian ice masses. Consequently, the research rationale and methodology outlined in this section will follow closely that employed by Shennan and co workers in their analyses of isolation basin deposits from Scotland. The methodology employed by the authors (*ibid*) has evolved specifically to facilitate the interpretation of this relatively long transition from marine to freshwater conditions or vice versa. Examples from Scottish isolation basin studies include Rumach VI, (Shennan *et al.*, 2000a); Loch nan Corr and Dubh Lochan, (Shennan *et al.*, 2000a; Lloyd 2000); Ardtoe, (Shennan *et al.*, 1996c).

Detailed comparison of isolation basins on a local to regional scale reveals marked variation in basin sediment lithostratigraphy. These are caused by differences in (*inter alia*) sediment flux to the basin, preservation potential and significantly, the speed of isolation. In regions formerly covered by major ice sheets (e.g. Scandinavia or North America), the pace of crustal rebound commonly outweighs the rate of eustatic sea-level rise throughout the entire postglacial period. This means that almost invariably all basin deposits reveal a strong negative sea-level tendency and a rapid isolation signature. However, in regions formerly covered by smaller ice caps (such as the BIIS), oscillations in the strength of the eustatic signal relative to the rate of isostatic rebound has led to highly non-monotonic sea-level histories since the LGM. The corollary of this is that many isolation basins have a history of being both isolated and inundated thus may have more than one isolation contact in their stratigraphic record. The Loch nan Eala basin at Arisaig, NW Scotland is an exemplar of such a basin (Shennan *et al.*, 1994, 2000a).

Traditional isolation basin studies adopted a hypothesis testing methodology whereby radiocarbon dates were taken from above and below the isolation contact and accompanied by both pollen and macrofossil analysis (Laidler, 2002). Since the work of Kjemperud (1981b, 1986) however, their has been a much greater emphasis upon rigorous analysis of both the litho- and biostratigraphy contained within the basins such that it is now common to employ a multi-proxy approach so as to reduce uncertainties over the exact timing of isolation/inundation. In the most recent isolation basin

studies of Shennan and co workers, pollen, diatoms (and to lesser extent foraminifera) have been extensively used in conjunction with detailed lithostratigraphic analysis to interpret the isolation signal. Pollen has a long history of use in studies of RSL and has commonly been use to infer tendencies of sea-level movement as well as validating age determinations obtained via radiometric dating. However, arguably the most commonly used and valuable microfossil assemblage employed in isolation basin studies is diatoms. Like foraminifera, species composition is controlled by key environmental parameters but significantly, they are found in sub, inter and supra tidal environments. Given that salinity is the dominant control on certain key species, they provide an excellent indication of whether marine, brackish or terrestrial conditions prevailed. Foraminiferal assemblages have also been used in the interpretation of basin deposits since salinity is known to exert a strong control on species composition (e.g. De Rijk, 1995a). The presence and absence of tests can thus be used to infer a marine influence on the basin although since elevation of the basin is largely standardized by the sill (Laidler, 2002), assemblage zones cannot be used to directly reconstruct RSL in the same way as they have been in employed in salt-marsh environments (e.g. Scott and Medioli, 1978).

Despite the potential for isolation basin studies to deliver high resolution time-series records of RSL change, the observations are beset with a number of generally well-understood difficulties (Lambeck *et al.*, 1998). Unlike RSL investigations sourcing SLIPs from intercalated marine-terrestrial sequences, isolation basins commonly only deliver a single 'snap shot' in time; if detailed local chronologies of RSL change are to be obtained, a series of basins in close geographical proximity with varying sill heights are required. Basins containing such sequences may or may not be present, depending on a number of site scale factors including coastal topography, coastal aspect and the nature of anthropogenic influence subsequent to deposition.

To constrain the maximum altitude of marine influence at a given locality, the presence/absence of marine sediments between basins with differing sill heights has often been considered (e.g. Shennan *et al.*, 1994, 2005, 2006b). However, this approach may not provide unequivocal evidence of the marine limit since factors such as basin infilling by peat or the presence of local ice (during the deglacial phase) may prevent marine inundation of the basin (e.g. Shennan *et al.*, 1995b).

Isolation basins may only deliver high resolution observations of RSL if the precise altitude of the basin sill is determined. Commonly, this is ascertained through systematic coring of the entrance to the basin yet even with detailed mapping of the sub surface bedrock profile, it may be difficult to distinguish between bedrock and boulder deposits. Where basins with rock thresholds are employed, erosion of the sill need not be considered as a further source of error. However, the same cannot be said of basins with a sill composed of boulder clay, beach deposits or other non-lithified strata. The threshold itself may have been eroded subsequent to isolation/inundation or even lowered by land drainage, such that its contemporary height is not a true representation of sea-level change (Lambeck *et al.*, 1998).

3.3 The application of Micropalaeontology to RSL Research

For a long time, coastal stratigraphers have espoused the virtue of using evidence from both the lithostratigraphic *and* biostratigraphic record to detect past changes in the coastal environment (e.g. Godwin *et al.*, 1935; Godwin and Clifford, 1938; Godwin, 1940). Traditionally, pollen analysis has been the most commonly employed indicator of change and has been used both as a chronological tool and as a device to reveal changes in the regional and local pollen floras which could then be related to water level changes (Tooley, 1986). More recently, however, workers have focused on the systematic use at an increasing resolution of a wide range of biological proxies sensitive to the frequency of tidal flooding and water chemistry (Allen, 2003). These include diatoms (e.g., Devoy, 1979; Palmer and Abbott, 1986; Juggins, 1992; Nelson and Kashima, 1993; Hemphill-Haley, 1995; Zong, 1997; Zong and Horton, 1999; Gehrels *et al.*, 2001; Sawai *et al.*, 2004; Shennan and Hamilton, 2006; Horton *et al.*, 2006), foraminifera (e.g. Scott and Medioli, 1978, 1980a, 1986; Haslett *et al.*, 2001; Horton, 1999; Edwards and Horton, 2000; Horton *et al.*, 2000; Gehrels *et al.*, 2001, 2002; Boski *et al.*, 2002; Horton *et al.*, 2004; Edwards, 2006a; Southall *et al.*, 2006), ostracodes (e.g., Boomer, 1998; Andrews *et al.*, 2000) and testate amoebae (Charman *et al.*, 1998; Charman, 2001; Gehrels *et al.*, 2001; Roe *et al.*, 2002). Gehrels *et al.* (2001) considered the benefits of adopting a quantitative multiproxy approach to sea-level reconstruction, incorporating foraminifera, testate amoebae and diatom evidence. However, whilst this route led to an improvement in both the accuracy and precision of reconstructed water levels, they considered the gain to be disproportionate to the amount of time invested in the analyses.

Traditionally, SLIPs have been sourced from both transgressive and regressive contacts in the lithostratigraphic record as these contacts are assumed to evidence the time at which a change in marine influence occurred. However, this change in litho-facies represents the *result of* a variation in marine influence and may lag the timing of the *actual* event itself (Heyworth and Kidson, 1982; Shennan, 1987,1989; Allen, 1990a,1995). Conversely, the response of both floral and faunal communities to environmental change is essentially instantaneous and therefore SLIPs derived from biostratigraphical contacts may provide a more accurate record of sea level.

In the analyses that follow, both foraminifera and diatoms have been chosen as preferred palaeoenvironmental indicators.

3.3.1 The role of Foraminifera Analysis in Sea-Level Research

Foraminifera are uni-cellular (protozoan) organisms found in marine and brackish habitats (Scott and Medioli, 1986). Their tests may either be constructed using cemented detritus (agglutinating or arenaceous forms), or secreted using calcium carbonate (calcareous forms) (Haslett *et al.*, 2001). In open marine settings, calcareous foraminifera dominate, either as planktonic species (living in suspension in the water column) or as benthic species (living in or on the sea floor, or attached to the substrate i.e. epiphytic). In contrast, agglutinated foraminifera become abundant in marginal marine settings (such as salt-marshes) where brackish salinity regimes are encountered.

3.3.1(i) The intertidal distribution of foraminiferal assemblages

The distribution of agglutinated foraminiferal species in salt-marsh environments has been of most interest to sea-level researchers. The analyses of Scott and Medioli (1978, 1980a) demonstrated that the salt-marsh foraminifera of North America are vertically zoned in a similar way to floral assemblages, but that these zonations are more tightly constrained (Horton, 1999). The authors (*idem*) suggested that under certain circumstances, foraminifera could be used to reconstruct sea level with a precision of ±5 cm, making them one of the most accurate indicators of past sea level contained within the sedimentary record. Salt-marsh foraminiferal zonation is therefore a significantly more accurate indicator of sea level than undifferentiated marsh deposits because the well-defined zones subdividing the marsh increase the vertical resolution of the deposits (Horton and Murray, 2006). Since the studies of Scott and Medioli, a number of authors have considered fossilised foraminiferal assemblages as a means of achieving high resolution records of RSL

change, particularly around the North Atlantic margin (e.g. Gehrels, 2000; Haslett *et al.*, 2001; Horton *et al.*, 2000, 2004; Edwards and Horton, 2000, 2006; Edwards *et al.*, 2004a, b; Harmon, 2006).

The vertical zonation concept (Scott and Medioli, 1978; 1980a) implies that all of the most important environmental parameters determining foraminiferal distribution (e.g. salinity, pH, elevation, nutrient availability and substrate) are influenced by depth and frequency of tidal submergence which is correlated with the elevation of the marsh surface (De Rijk, 1995a). However, De Rijk (1995a,b), and De Rijk and Troelstra (1997) questioned this relationship by proposing that salinity instead of elevation was the dominant control on the distribution of foraminifera within the inter-tidal zone. Their results from the Great Marshes, Massachusetts suggest that the simple pattern of vertical zonation may only be encountered in marshes where the surface gradient is low and decreases uniformly towards the sea. In marsh systems where the surface topography is irregular, the zonation pattern is interrupted and foraminiferal distribution becomes strongly related to local spatial and temporal changes in a number of environmental variables (notably salinity), and not altitude. Instead of direct indicators of sea level, foraminifera can only be used to infer former salinity conditions which reflect the balance between seepage, precipitation and flooding. This balance can only be interpreted in terms of *both* climate and sea level.

Taking into consideration these competing hypotheses, Horton (1999); Horton *et al.* (1999) considered the relationship between foraminifera and altitude at ten inter-tidal studies areas from around the UK. The vertical zonation of the ten study areas into two zones – (a high- and middle marsh zone dominated by agglutinated species; and a low-marsh and tidal flat zone dominated by calcareous species) - supports the conclusions of Scott and Medioli (1978, 1986) and others (Scott and Leckie, 1990; Jennings *et al.*, 1995) that the distribution of foraminifera in inter-tidal environments is usually a direct function of altitude with the duration and frequency of inter-tidal exposure as the most important factors. As such, the foraminiferal zonation of the UK does provide a tool for interpreting former sea levels from fossil environments.

3.3.1(ii) Post-depositional alteration of foraminiferal assemblages

An inherent assumption in reconstructing past sea-level change using foraminifera is that modern assemblages collected at the surface of the marsh are accurate analogues of microfossil assemblages in subsurface sediments (Patterson *et al.*, 1999). Often, this may not be the case because of post-depositional changes which are the result of diagenetic processes such as selective preservation, transportation of tests, bioturbation, infaunal habitats, and predation (Murray, 1991; De Rijk and Troelstra, 1999). In general, the most significant of these postmortem processes are transport (which can lead to loss or gain of tests) and destruction (especially dissolution of the calcareous tests but also disintegration of fragile agglutinated taxa) (Murray and Alve, 1999).

a) Selective preservation

Selective preservation of foraminifera is caused by differences in test structure and composition (De Rijk, 1995b; De Rijk and Troelstra, 1999). In temperate salt-marshes, calcareous taxa are commonly most prone to post depositional alteration which occurs due to the dissolution of the test by organic matter or mineral oxidation (Green *et al.*, 1993). The general non-preservation of calcareous foraminifera in many marsh settings as a result of dissolution has been well documented (Parker and Athearn, 1959; Phleger, 1960, 1965, 1970; Akers, 1971; Scott, 1976; Scott and Medioli, 1980a; Jonasson and Patterson, 1992, Goldstein and Watkins, 1999; Edwards and Horton, 2000; Horton *et al.*, 2005; Culver and Horton, 2005; Horton and Murray, 2006) and this can significantly alter the composition of fossil samples leading to an overrepresentation of agglutinated forms. The significance of calcareous test dissolution to salt-marsh palaeo-environmental studies will depend upon whether calcareous foraminifera are diagnostic species for any of the vegetated salt-marsh environments.

A number of studies have reported good preservation of agglutinated foraminifera in salt-marsh deposits (e.g. Ellison and Nichols, 1976; Scott and Medioli, 1978, 1982; Thomas and Varekamp, 1991; De Rijk, 1995b), suggesting that they are more resistant to post-mortem change than calcareous taxa (e.g. Phleger, 1970). However, some species with delicate tests are prone to dissaggregation (e.g. Murray, 1991; De Rijk, 1995b; De Rijk and Troelstra, 1999). Goldstein and Watkins (1998, 1999) have indicated that certain agglutinated species were less likely to be

preserved than others as a result of bacterial degradation of the cements holding agglutinated particles together. Similarly, von Daniels (1970) noted that the instability of the cement in certain agglutinated foraminifera has led to the loss of tests in the fossil record. Agglutinates may also be at risk from compaction and drying, often leading to the collapse of chamber walls (De Rijk, 1995b).

b) Transportation

Foraminiferal assemblages may be changed by both transportation into (mixing) and out of (loss) an area (De Rijk, 1995b) and it is rarely possible to separate the allochthonous and autochthonous components of the assemblage. Mechanisms for the transportation of tests include bed load, suspended load, ice or floating plants, although transport by water is considered to be the most important (Murray, 1991). The transport itself reveals selective bias and will depend largely upon foraminifera size and specific gravity (Murray, 1991). Where water energy on the marsh surface is very low, it is rare for tests greater than 63μm to be re-suspended (De Rijk and Troelstra, 1999). However, when water energy is increased (e.g. by high tides or storm events) the introduction of allochthonous foraminifera may occur. It is often the case that in samples where post-mortem dissolution has been negligible, species diversity of a dead foraminiferal assemblage is greater than that of the living assemblage from the same sample (Murray, 2003): This is commonly a result of the transportation of shelf taxa by high tidal energy and their subsequent incorporation into the death assemblage (e.g. Murray, 1983, 1987, 2003; Horton and Murray, 2006).

c) Bioturbation

Bioturbation refers to the stirring or mixing of particles as a consequence of burrowing or boring by sediment dwelling organisms. Commonly, the majority of bioturbation occurs in the top few centimetres of the sediment column (Aller and Cochrane, 1976) although it may take place at a greater depth (e.g. Gallagher and Plumley, 1979). The roots of salt-marsh plants may cause bioturbation although the influence of vegetation is likely to be subordinate to larger organisms such as fiddler crabs (De Rijk, 1995b; De Rijk and Troelstra, 1999).

d) Infaunal foraminifera

Infaunal Foraminifera are benthic species that live within the sediment column. Sea-level reconstructions based on modern surface distributions of foraminifera implicitly assume that

infaunal populations do not constitute a significant proportion of the total assemblages recovered (Edwards *et al.*, 2004a). However, the composition of assemblages preserved in subsurface marsh sediment may sometimes differ from that of surface assemblages and the occurrence of these infaunal dwellers may go a long way to explaining this pattern (Goldstein and Harben, 1993). The depth to which infaunal activity occurs has attracted some debate and is strongly conditioned by the nature of the marsh substrate. Whilst the findings of Horton (1997); Horton *et al.* (1999); Saffert and Thomas (1998); De Rijk and Troelstra (1999); Alve and Murray (2001) suggest that the majority of foraminifera may be found living in the top 1-2cm of the marsh, a number of other studies have shown that foraminifera may be found living to a depth of at least 10cm in the sediment column (e.g. Ozarko *et al.*, 1997; Patterson *et al.*, 1999; Goldstein and Watkins, 1999). Furthermore, in a study from Georgia, (USA) Goldstein *et al.* (1995) found that many of the most common (and palaeoenvironmental significant) salt-marsh foraminifera occupy infaunal microhabitats, and some live as deep as 30 cm below the marsh surface.

e) Predation

Foraminiferal predators include other foraminifera, nematodes, gastropods, crustaceans, fish, and fiddler crabs (Lipps and Valentine, 1970; Murray, 1991; Green *et al.*, 1993; De Rijk, 1995b). Several authors have speculated about the role played by predation including Goldstein and Watkins (1999) who noted half tests belonging to agglutinated foraminifera and attributed this to predation, (possibly by fiddler crabs). When present in high densities, foraminifera may provide a large food reserve (De Rijk, 1995b) although by and large, predation should not cause significant alteration to the extant foraminiferal assemblage (Murray, 1991).

3.3.2 The role of Diatom Analysis in Sea-Level Research

Diatoms are photosynthetic unicellular organisms possessing a siliceous exoskeleton (Morris, 1992). They are found throughout freshwater, brackish and marine environments and have been widely used in palaeoenvironmental studies looking at changes in (*inter alia*) salinity, pH, nutrient availability and temperature (Juggins, 1992). A number of researchers have used diatoms in studies attempting to reconstruct patterns of sea-level movement (e.g. Haggart, 1986; Long, 1992; Zong, 1992; Plater and Shennan, 1992; Robinson, 1993; Shennan *et al.*, 1995a; Zong and Tooley, 1996; Sawai *et al.*, 2004). However, as with foraminifera, differences of opinion exist regarding the

extent to which altitude within the tidal frame influences diatom distribution. For example, Shennan *et al.* (1995a, 1996a), Zong (1997) and Zong and Horton (1998, 1999) all present evidence for strong vertical zonation of diatom assemblages. This concept implies that all environmental variables that determine diatom distribution are related to tidal inundation, which in turn, is correlated with the altitude of the salt-marsh surface (Zong and Horton, 1999). Others have suggested that diatom assemblage composition is influenced by a number of other environmental factors (such as pH, salinity and nutrient availability) which are not always directly related to altitude. This necessarily provides a caveat to their value as indicators of sea level.

In this thesis, diatom data have only been used to validate sea level index points rather than as a tool for reconstructing palaeo sea level. Their use is confined to isolation basins (where the sill height defines the indicative meaning) and lithostratigraphic contacts in salt-marsh deposits where foraminifera are absent. In such instances, reconstructed water levels have been derived from lithostratigraphical evidence alone. This follows the established protocol for the reconstruction of coastal history based on diatom biostratigraphical techniques (e.g. Zong and Tooley, 1996). Such studies rely strongly on the simple classifications of taxa into freshwater, brackish or marine forms, providing qualitative estimates of ecological conditions.

The comprehensive ecological classifications of coastal diatoms, particularly for salinity and life form (e.g. Denys, 1991–2; Vos and de Wolf, 1993) have been widely used for diatom-based sea-level studies (Zong and Horton, 1999). The system used to classify diatoms according to their salinity tolerance is called the halobian system of classification. It was first devised by Kolbe (1927) although a number of modifications have now been made to it (e.g. Hudsted, 1953, 1957) and Hemphill-Haley (1993) (Table 3.1). Polyhalobous species flourish in fully marine conditions, (with a salinity exceeding 30 practical salinity units (psu)) whilst mesohalobous diatoms thrive in salt concentrations of between 0.2 psu and 30 psu. Oligohalobous diatoms generally occur in salt concentrations less than 0.2 psu. Vos and de Wolf (1993) and Juggins (1992) subdivided the latter category into oligohalobous-halophilus, which have an optimum in weakly brackish waters, and oligohalobous-indifferent, which show a preference for fresh water, but are tolerant of slightly brackish conditions (Horton *et al.*, 2006). Halophobous diatoms are intolerant of saline conditions and are found exclusively in fresh water.

Classification	Salinity range (‰)	Description
Polyhalobous	>30	Marine
Mesohalobous	0.2 to 30	Brackish
Oligohalobous-halophile	<0.2	Freshwater – stimulated at low salinity
Oligohalobous-indifferent	<0.2	Freshwater – tolerates low salinity
Halophobous	0	Salt-intolerant

Table 3.1 The halobian classification system (Hemphill-Haley, 1993)

As Hillier (2006) notes, a basic interpretation of this classification system should see a change in the salinity classes of the diatom assemblages with increasing/decreasing altitude in the inter-tidal zone: Polyhalobous species occur in sub tidal areas and on the tidal flat along with mesohalobous diatom species. As marine influence decreases oligohalobous-halophilous and oligohalobous-indifferent species will increase at the expense of polyhalobous and mesohalobous species. Finally halophobous species will occur above HAT in the freshwater environments.

3.4 Site Selection

3.4.1 Salt-marshes

For the purpose of studying trends in Holocene RSL, Edwards (1998) notes that salt-marsh sites should (ideally) contain the following attributes:

1: Deep sequences of intercalated organic and minerogenic sediments, preferably including thin basal peats overlying a bedrock surface at a range of altitudes.

2: Sediments that contain microfossil assemblages in sufficient density to enable the statistical analysis of samples.

3: Be located in a low energy environment, with low tidal range and relatively open communication to the sea

4: Have an absence of disruptive anthropogenic activities such as embanking and reclamation.

With respect to point 1, sequences containing intercalations of marine and terrestrial sediments offer the potential to deliver a suite of SLIPs from a geographically restricted area and if basal peats are present, the index points are likely to be virtually compaction free (e.g. Shennan and Horton, 2002). If these basal peats are found at varying altitudes, they are likely to be diachronous and can potentially then be used to 'calibrate' index points which are likely to have experienced autocompaction.

Unfortunately, the final criterion (point 4) is seldom met at most low lying coastal localities. It is common for there to have been a degree of anthropogenic manipulation of the natural environment stretching back to the Mesolithic and the ability to inflict widespread change has only increased over time (Healy and Hickey, 2002). Such human-environment interaction can introduce 'artificial' changes in RSL, either through the destruction or creation of coastal sequences and may also significantly alter local tidal characteristics. This necessarily complicates interpretation of the indicative meaning.

A thorough review of existing literature (chapter 2) (coupled with the development of the sea-level database for Ireland) revealed where past studies related to RSL had been undertaken. Where little or no published RSL research had been undertaken (e.g. County Limerick, Claire, Galway), potential sites were sourced (primarily) through reference to 'The Salt Marshes Of Ireland: An Inventory and Account Of Their Geographical Variation' (Curtis and Sheehy-Skeffington, 1998). This directory of Irish salt-marshes was used in conjunction with both 1:50000 OS maps and aerial photographs, available through the National Coastline Survey. The total number of salt-marshes documented in the Irish salt-marsh inventory exceeds 250 thus sampling was biased in favour of sites with substrate that was likely to be penetrable with basic hand auger equipment. Commercial borehole records existed for potential study sites, and these were obtained through local councils.

3.4.2 Isolation Basins

Many of the criteria used in the selection of suitable salt-marsh sites and potential isolation basin sites are similar. As with salt-marsh-based RSL research, it is desirable to look for basins in micro to meso tidal localities, sheltered from prevailing wave attack and situated in areas with minimal anthropogenic interference. However, unlike salt-marshes which are more or less ubiquitous

around Ireland's coasts, by their very nature isolation basins are exclusively located in areas that have experienced higher than present sea levels in the recent geological past. Reference to the literature base reveals that in Ireland, isolation basins are only going to have the potential to form in north and east Ulster as well as along the north Leinster coastline (e.g. Carter, 1982a; Penney, 1983; Shaw, 1985; Shaw and Carter, 1994).

Ordnance survey maps (and to a lesser extent aerial photographs) were used to source topographic hollows and small lakes in coastal localities. Only basins that were located immediately adjacent to the coastline with a clearly definable 'pathway' between basin and coast were chosen. (This is important to facilitate the location of the basin sill). Reference to OS maps specifically highlighted the Fanad Peninsula in northeast Donegal as the most suitable study area.

3.5 Field-Based Techniques

3.5.1 Site Coring

Coring was undertaken along a baseline transect commonly orientated perpendicular to the coastline. This primary transect was augmented by additional coring either side of it to ascertain the extent of lateral variability in lithology at each study site. Core frequency varied within and between sites and was primarily guided by the spatial complexity of the stratigraphy. Generally, initial core spacing did not exceed 30m.

In order to quickly achieve an overview of the lithostratigraphy at each site, basic hand auger coring equipment was used to retrieve sediment. Although reference to Long *et al.* (1999b) showed that hand gauge coring is not without its problems (e.g. sediment contamination and compaction), it was adopted because the apparatus offered a flexible, cheap and relatively quick way of obtaining stratigraphic data to depths of c.10m (Tooley, 1981). The stratigraphic data was drawn up in the field to provide a cross section of the coastal sedimentary sequence. From this, a location offering a deep profile with the potential to deliver SLIPs could be located. Material for palaeoenvironmental analysis was removed using a Russian-type coring device and checked in the field for contamination before being placed in plastic tubing and sealed with tin foil. On return to the laboratory, all cores were stored below 4^0 C to retard biological and chemical processes. In the case of the cores extracted from the Inner Shannon estuary, a 100mm bore MOSTAP sampler was

employed. This system employs a hollow stem sampler with an inner mesh liner to recover continuous soft sediment samples. Cores were later stored in a cool environment, out of direct sunlight.

3.5.2 Stratigraphic Description

> *'Any scheme for the stratigraphic description of coastal sediments must be broad enough to cover the range of organic and inorganic deposits likely to be encountered, flexible enough to allow quantification of the separate components of what may be a heterogeneous deposit, and yet simple enough not to hinder extensive data collection'* (Long *et al.*, 1999b; pp 268).

The approach devised by Troels-Smith (1955) fulfils the above criteria and this system (or modifications of it e.g. Aaby and Berglund 1986; Birks and Birks 1980) has been commonly used by coastal stratigraphers e.g. Tooley (1978); Shennan *et al.* (1995a); Tooley and Smith (2005).

The Troels-Smith (1955) scheme is based on the semi-objective classification of unconsolidated sediments. It is purely descriptive and independent of any knowledge of depositional processes (Long *et al.*, 1999b). Stratigraphic units are described according to their physical components (e.g. clay, sand etc), physical properties (e.g. colour, elasticity etc) and humicity with each group assessed on a 5 point scale. Whilst there are caveats to the scheme (e.g. problems associated with the classification of humified deposits and differentiating between the size of mineral sediments), the overall ease with which the system can be employed coupled with its widespread use in Holocene RSL studies along temperate coastlines meant this scheme was preferred over others. (A full list of sediment types and definitions of physical properties can be found in Appendix III).

3.5.3 Sampling for Contemporary Foraminiferal Assemblages

Modern samples were collected along transects encompassing a range of altitudes and vegetation zones between HAT and MHWNT in order to provide analogous conditions to those environments encountered in the fossil cores (Horton *et al.*, 2000). Each transect was oriented across the salt-marsh (and upper reaches of the inter-tidal flat), broadly perpendicular to the shoreline. Disturbed areas (such as drainage ditches and ponds) were avoided since boreholes containing these

deposits are not considered reliable indicators of RSL change (e.g. Edwards *et al.*, 2004a). Unfortunately, it is the case that the majority of contemporary salt-marshes have experienced some form of anthropogenic interference such as embankment. This will affect the hydrological regime of the marsh (Van der Molen, 1997) and thus may affect the vertical zonation of foraminiferal assemblages across the marsh. Where possible, the sampling transect was located away from obvious 'artificial' features and orientated across tracts with well developed floral zonation. Such areas are suggested to better depict the natural, (pre-engineered) state of the marsh. Indeed, the monospecific *Jadammina macrescens* assemblage (or zone 1a - Scott and Medioli, 1978) found in the high marsh is probably the most accurate and precise indicator of sea level in temperate regions, and this seems to be best developed in areas where 'natural' transitions to upland environments are encountered (Edwards *et al.*, 2004a). Samples were not taken at fixed horizontal intervals but instead at marked changes in topography, such that the vertical interval of samples was usual about 10cm. Each sampling station was then levelled to OD(B). Each sample was collected from the top 0-1cm of the marsh surface and measured $10cm^3$ in volume in accordance with similar studies (e.g. Scott and Medioli, 1980b; Scott and Leckie, 1990; De Rijk, 1995a; Horton *et al.*, 1999).

3.5.4 Levelling

Levelling to Ordnance Datum benchmarks was carried out at most of the sites described in Chapter 6. These benchmarks were located through reference to the Ordnance Survey of Ireland's six inch to 1 mile map series (1922 edition). Where benchmarks were not available, the site in question was levelled to the position of high water and altitude relative to Ordnance Datum then calculated via reference to Admiralty Tide Tables. At all sites from the Inner Shannon, a differential Leica GPS 500 was used to calculate heights and geographical locations of the cores. This equipment produces an altitudinal accuracy of within ±10 mm.

3.6 Laboratory Based Techniques

3.6.1 Preparation and Analysis of Samples - Contemporary Foraminifera

Sample preparation followed that of Scott and Medioli (1980b) and Horton (1999): Samples were washed through both a 500μm and 63μm sieve and the fraction between 500μm and 63μm was retained. (Plant debris larger than 500μm can inhibit counting whilst agglutinated foraminifera from

temperate coast marshes are rarely less than 63μm in size). Within 48 hours of collection from the field, this sample was stored in buffered ethanol along with Rose Bengal and refrigerated. Rose Bengal identifies organisms living at the time of collection (Walton, 1952) and this has commonly been used in previous studies to differentiate between living and dead foraminifera (e.g. Scott and Medioli, 1980b; Murray, 1991; Horton *et al.*, 1999; Patterson *et al.*, 1999). The assumption was made that tests containing protoplasm were living at the time of collection although in order to ensure that bacteria residing in the test were not responsible for the red staining, only tests with the last few chambers stained red were considered living (Murray, 1991). After staining, the samples were washed in tap water, stored in 50% ethanol solution and refrigerated. Since processing methods can introduce an important source of error in quantitative micropaleontological studies (Hart, 1983; Hodgkinson, 1991), it was important that a reliable and efficient method, capable of splitting samples into representative parts was adopted. De Rijk (1995a) showed that wet splitting was the most time efficient preparation method and the results were comparable in accuracy to more time intensive procedures such as decantation. Samples were thus suspended in approximately 600ml of water and sub-divided into eight aliquots using a wet splitter (Scott and Hermelin, 1993). The prepared samples were then analyzed wet under a binocular microscope at approximately 40x magnification – (water immersion helped in the identification of marsh taxon because organic matter contained in the samples often adheres to the tests if they are dry, thus impeding accurate identification). Foraminifera were later identified through reference to the taxonomic guide produced by Horton and Edwards (2006).

In contemporary foraminiferal assemblages, samples commonly possess a greater number of dead than living specimens (e.g. Horton, 1999; Edwards *et al.*, 2004a etc). However, there has been much debate concerning which of these populations (dead, living or both) most accurately represents the sub surface, 'fossil' assemblage. Some researchers have advocated using the total foraminiferal assemblage (e.g. Scott and Medioli, 1980b; Jennings *et al.*, 1995; De Rijk, 1995b; Laidler, 2002; Tobin *et al.*, 2005) although others have cautioned against such an approach (e.g. Murray, 1991; Murray and Alve, 1999). However, the recent studies of Horton (1997, 1999); Horton and Edwards (2005, 2006) have suggested that death (thantocoenosis) assemblages are the most accurate representation of that found in subsurface, fossil distributions. Unlike the life assemblage, the long-term study of foraminiferal death assemblages shows that the assemblage is in

equilibrium with the depositional environment in which it is found, and is less susceptible to seasonal fluctuations. On the strength of these latter findings, in this study only the foraminiferal death assemblage was considered.

3.6.2 Preparation and Analysis of Samples - Fossil Foraminifera

Samples measuring $5cm^3$ were taken across each lithostratigraphic contact in the core. The samples were then soaked in tap water (along with dispersant depending on the nature of the sediment) and then passed through 500μm and 63μm sieves. Thereafter, sample preparation was the same as for the contemporary samples, with the exception of adding Rose Bengal.

3.6.3 Preparation and Analysis of Samples - Diatoms

Sample preparation followed the guidelines set out by Palmer and Abbott (1986): Approximately 0.01 grams of dried sediment sample (or 0.1 gram of wet sediment) was placed in a test tube and moistened with a few drops of H_2O_2 to help dissaggregation of the sediment matrix. 5mls of 30% H_2O_2 was then added to the sample and the test tubes were then placed in a water bath at 80^0 c for 2-4 hours to remove the organic component of the sediment sample. The test tubes were subsequently removed from the bath and 1-2 drops of 50% HCl were added to each sample to remove any remaining H_2O_2 and carbonates. The test tubes were then topped with distilled water and centrifuged at 1200rpm for four minutes. After centrifuging, the resultant supernatant liquid was decanted off and the diatoms resuspended in more distilled water. This procedure was repeated a further four times. During the final wash, 1-2 drops of weak NH_3 solution was added to each sample to help keep any clay in suspension and prevent diatom clumping during slide preparation.

The cleaned diatom suspension was then diluted to a suitable concentration and a 1ml pipette was used to place 0.5ml of (well mixed) diatom suspension on a cover slip. The cover slip was then left for 24hous to dry. After the cover slip had dried, it was mounted on a drop of naphrax on a glass slide and heated by hotplate to 130^0 c. After 15 minutes, the slide was removed and left to cool.

As part of a larger project, the diatom analysis presented in this thesis was undertaken by Caroline Hillier with diatom species classified according to the halobian classification system (de Wolf, 1982; Vos and de Wolf, 1993).

Where possible, diatom counts were expressed as a percentage of total diatom valves (%TDV).

3.6.4 Sample Counts

As Horton and Murray (2005) observe, some controversy exists concerning the statistically correct number of counts required for quantitative examinations, particularly the effect of variations in the number of species between samples and the significance of varying fractional abundances on the reliability of results (e.g. Patterson and Fishbein, 1989; Buzas, 1990; Bennett, 1994; Hayek and Buzas, 1997; Fatela and Taborda, 2002). In order to undertake statistically significant analysis of foraminifera/diatom samples, it is generally held that a target count of at least 200 specimens per sample is required (Lipps, 1993) although ideally, this figure should be between 300 and 400 specimens (Murray, 1991). However, Patterson and Fishbein (1989) state that this number is highly dependent upon assemblage diversity and the relative abundance of the most important minor species. They go on to note that low counts only affect quantitative assessments if important indicator species are present in small abundances. In certain circumstances, sample counts as low as 40 have been incorporated in statistical analysis (e.g. Horton, 1999). Low foraminifera counts are often associated with samples from the most landward end of the marsh at the upper limit of marine influence. Such samples are often characterized by very low species diversity and therefore can be reliably represented by low counts (Edwards *et al.*, 2004a). In this study, a target count of 300 (identifiable) specimens was set for each sample.

3.7 Chronological Control for Coastal Sequences

Placing coastal sequences (or at least part of a sequence) within a firm chronological framework is an essential requirement for the development of reliable SLIPs. In the British Isles and elsewhere around northwest Europe, nearly all studies of RSL rely on radiocarbon dating to determining the age of various (organic) sea-level indicators. However, a number of more recent studies have investigated the viability of alternative methodologies such as luminescence dating to assess the age of clastic sedimentary horizons in coastal archives (e.g. Plater and Poolton, 1992; Boomer and Horton, 2006).

3.7.1 Radiocarbon Dating: Sampling, Preparation and Calibration

Here, an age framework for the encountered stratigraphy was delivered using accelerator mass spectrometry (AMS) radiocarbon dating of organic sediment horizons/remains. The merits of using AMS radiocarbon dating instead of the conventional radiometric analysis have been discussed elsewhere (e.g. Walker *et al.*, 2001) although principally, the AMS method was chosen because it allows dates to be obtained from far smaller sediment samples.

All samples presented in Chapter 6 were submitted to the Beta radiocarbon dating laboratory. In the case of peat/organic samples, approximately 5g of sediment was extracted from the core and plant rootlets removed under microscope. At Beta, the sample was sieved and the (organic) sample fraction >180 microns was used for dating. All peat samples underwent an acid/alkali/acid pre-treatment procedure.

Due to the allocthonous nature of many shell deposits in coastal sequences, they cannot be relied upon to deliver a dependable age for a sediment horizon. (After initial deposition it is common for shells to be remobilized and therefore they are often likely to be older than the sediment matrix in which they are found). Some researchers have employed dates from shells in conjunction with dated calcareous foraminifera to provide 'paired' dates for a sedimentary unit (e.g. Horton *et al.*, 2000). If only pristine, non-abraded foraminifera are selected from a sediment sample, the likelihood of dating older, allochthonous material is reduced. In turn, the foraminiferal-based date may help identify a potentially erroneous shell date derived from the same sediment sample. This approach has been adopted here. All shell samples dated in this study have been exposed to an acid etch pre treatment at the Beta laboratory. However, because only a small quantity of carbon is available from foraminiferal samples, no laboratory pre-treatment method was possible.

In Chapter 6, radiocarbon ages have been expressed in calendar years BP to facilitate comparison with glacial rebound model predictions of RSL (Chapter 5). Radiocarbon ages were calibrated to calendar years BP using CALIB 5.0.1 (Stuiver *et al.*, 2005). The following options were chosen: A laboratory multiplier effect of 1; 95% confidence limits and their dataset IntCal04. For marine carbonate samples (such as shells and foraminifera) the dataset Marine04 was employed. All

dates have been expressed with an error term of ± two standard deviations, giving a statistical probability of 95% that the true value lies within the given range.

3.7.2 Age Uncertainties

The reliable interpretation of the age data rests upon the assumption that the dated sea-level indicator reflects the timing of the sea-level change of interest. The main challenges to this assumption are briefly outlined here. An in depth discussion concerning the age of sea-level index points and dating material for sea-level reconstructions can be found in Mook and van de Plassche (1986), Aitkin (1990) and Edwards (2004).

3.7.2(i) *Delayed response of the lithostratigraphy to RSL change.*

As already mentioned, a number of researchers have questioned the validity of RSL age – altitude 'curves' which are based upon the interpretation of radiocarbon dated peat facies (e.g. Allen, 1990c, 1995; French, 1993). This is because (theoretical) evidence exists of a lag between a change in RSL and its registration in the lithostratigraphic record. Accordingly, many sea-level researchers now choose to date biostratigraphic transisitions instead since the response of both floral and faunal communities to environmental change can usually be regarded as instantaneous.

3.7.2(ii) *Contamination of the sediment sample*

Contamination of a radiocarbon sample involves the incorporation of either older or younger carbon into the sample. Downward penetration of rootlets and humic acid can contribute to an artificially young age for a sediment sample. Similarly, younger carbon may also be introduced if fungal growth is allowed to develop on the sample surface after collection from the field (Wohlfarth *et al.*, 1998). A sample may yield an erroneously old age if contamination with older carbon occurs. This is most problematic in lake contexts where older organic carbon can be washed in from the catchment slopes or from local bedrock and result in an ageing effect for the sediments contained in the lake (Walker, 2005). Shennan *et al.* (2006b) draws attention to this problem when using isolation basin SLIPs to reconstruct changes in RSL.

3.7.2(iii) *Calibration of the radiocarbon timescale*

Because temporal fluctuations in atmospheric ^{14}C activity have taken place, significant departures exist between the radiocarbon and sidereal timescales. Accordingly, it is necessary to calibrate radiocarbon ages and (as already mentioned), this is carried out here using CALIB 5.0.1 (Stuiver *et al.*, 2005). However, the calibration operation is relatively imprecise as a consequence of age errors associated with the dendrochronological records and uranium-thorium dated coral series that make up the calibration dataset (Walker, 2005). In addition, the variations in atmospheric ^{14}C production (coupled with the statistical nature of radioactive decay) mean that the calibration of an individual ^{14}C age estimate will yield a range of calendar ages. This age range may span up to 400 years where the timeframe of interest coincides with a period when the ^{14}C calibration curve exhibits a 'plateau' (Edwards, 2004).

3.7.2 (iv) *The marine reservoir effect*

^{14}C is transferred from the atmosphere to the ocean across the ocean surface but because the mixing rate of surface and deep waters is very slow, surface water often has an apparent age (Lowe and Walker, 1997). In the surface waters of the British Isles, the marine reservoir effect is approximately 400 years (although a degree of spatial variability exists e.g. Harkness, 1983 – Table 4.1) and this correction factor must be applied to dates derived from marine carbonate deposits. However, because perturbations have occurred in the rate of thermohaline circulation (and hence mixing of surface and deep water bodies), the atmosphere-sea surface ^{14}C difference is unlikely to have remained constant in time (e.g. Bard *et al.*, 1994). The magnitude of error introduced by this temporal variation is hard to quantitatively assess although is likely to be greatest for older (pre Holocene) samples.

3.8 Reconstructing the Indicative Meaning of a Sea Level Index Point

3.8.1 Salt-marsh Foraminifera and the Indicative Meaning

The understanding of former sea levels based on the identification and interpretation of salt-marsh foraminiferal assemblages requires that their indicative meaning is known. In order to establish this relationship between foraminifera and water level, detailed surface studies considering the modern distribution of foraminifera across the salt-marsh surface are required. Once surface samples of

foraminifera have been collected and identified, they may be used in one of two principal ways to define the indicative meaning of a fossil sample. The first requires contemporary marsh samples from a site to be classified into more or less homogenous groups which enable the vertical zonation of the marsh surface to be established. The indicative meaning is defined by the midpoint of each zone and the altitudinal range of the locally determined faunal zones restricts the indicative ranges (e.g. Scott and Medioli, 1978,1986; Horton, 1999). The second approach involves the use a transfer function to reconstruct the indicative meaning and is based upon a modern training set of marsh foraminiferal samples (e.g. Gehrels *et al.*, 2001; Edwards *et al.*, 2004b; Horton and Edwards, 2006). The application of transfer functions to quantify the indicative meaning represents a relatively recent development in sea-level studies (Long, 2003) and provides a means of delivering objective, quantitative and reproducible estimates of water level associated with explicitly stated error terms (Horton *et al.*, 1999).

Transfer functions were pioneered by Imbrie and Kipp (1971) and have become commonly used tools for quantitative palaeoenvironmental reconstruction (e.g. Jones and Juggins, 1995; Gasse *et al.*, 1997; Noon *et al.*, 2001; Rosén *et al.*, 2003). These statistical palaeoenvironmental reconstructions use multivariate techniques to determine patterns and trends in the natural system, with the goal of reducing the dimensionality of the data, so that they can be more easily interpreted (Kovach, 1995). The overall aim is to identify the controlling environmental variables, and then to express these as a function of the biological data. This function is termed a transfer function or 'biotic index' (ter Braak, 1987b; ter Braak and Prentice, 1988; Birks, 1995). If reconstructions made by the transfer function are to be considered reliable, however, there are a number of requirements that must be satisfied (Birks, 1994):

i. A biological system which produces abundant identifiable fossils and that is responsive and sensitive to the environmental variables of interest at the spatial and temporal scales of the study.
ii. A large, high-quality, training set of modern surface-samples and associated environmental data should be available.
iii. The fossil data-set should be of consistent taxonomy, and of comparable quality to the modern training set.
iv. Good independent chronological control is required for fossil data-sets in order to permit correlation and comparison.

v. Robust statistical methods are required that can model the complex, often unimodal relationships between modern taxa and their environment.
vi. Reliable and realistic statistical estimation of standard errors of prediction for the modern training set, and for each reconstructed value, is required.
vii. Critical ecological and statistical evaluations of all reconstructions are needed, as any statistical regression and calibration procedure will produce a result.

There also exist a series of assumptions that must be kept in mind throughout (Imbrie and Webb, 1981; Birks *et al.*, 1990):

i. The biological taxa in the modern training set are systematically related to the modern environment they inhabit.
ii. The environmental variable(s) to be reconstructed (X) is, or is linearly related to an ecologically important determinant in the ecological system of interest
iii. The taxa in the training set are the same biological entities as in the fossil data, and their response(s) to the environmental variables have not changed significantly over time.
iv. Mathematical methods in regression and calibration adequately model the biological responses to the environmental variables of interest and yield calibration functions with sufficient predictive powers to allow useful, accurate and unbiased reconstructions
v. Environmental variables other than the one of interest have negligible influence

3.8.1(i) Assemblage zone approach

In the original analyses of Scott and Medioli (1980a), contemporary salt-marsh foraminifera were visually (or qualitatively) grouped to define vertical zonations at a number of sites. More recent work has sought to group assemblages on statistical grounds which may then be related to distinct depositional environments (e.g., Patterson, 1990; Jennings and Nelson, 1992; De Rijk, 1995a) (Edwards *et al.*, 2004a). The process by which the vertical zonation of surface foraminifera is established has been set out in Horton *et al.* (1999) and is repeated here:

> *The vertical zonation of the study area is determined using unconstrained incremental sum-of squares cluster analysis and detrended correspondence analysis (DCA). Unconstrained cluster analysis based on unweighted Euclidean distance (no transformation or standardisation of the data) is used to classify contemporary samples into more or less homogeneous clusters (Prentice, 1986; Van Tongeren, 1987). DCA is applied to represent samples as points in a multi-dimensional space. Similar samples are located together and dissimilar samples apart. Birks (1986, 1992) stressed the complementarity of cluster analysis and DCA. Cluster analysis is effective in classifying the samples according to their foraminiferal assemblage but DCA*

gives further information about the pattern of variation within and between groups. Therefore, DCA provides a means of evaluating the reliability of cluster analysis (Prentice, 1986). The altitude of each station within the reliable clusters are analysed to determine a vertical zonation of the inter-tidal environment.

Cluster analysis and DCA are based on computations made with the Psimpoll release 4.25 (Bennett, 2005). Only samples with counts greater than 40 individuals and species that reach 5% dead foraminifera are included (following Gehrels, 1994; Shennan et al., 1996b).

3.8.1(ii) Transfer function approach

A number of quantitative palaeoenvironmental reconstructions have been based upon transfer functions which are constructed using locally derived modern training sets to model the biota-environment relationship. These site-specific studies that derive relationships reflecting the particular set of physiographic or climatic conditions found locally are capable of delivering the most precise fauna-environment associations (Horton and Edwards, 2006). However (as previously mentioned), the use of microfossils in palaeoenvironmental reconstruction is founded upon the premise that the modern relationship between particular indicator organisms and the environmental variable of interest has remained constant through time (Edwards *et al.*, 2004b). In northwest Europe, it is the case that modern salt-marshes (along with the biota contained within them) may not always provide representative analogues for the range of palaeoenvironments that are found in stratigraphical sequences. This is because factors such as climate and (particularly) human activity may combine to deliver a different set of contemporary marsh conditions than encountered in the past. For this reason, combined training sets from various sites are commonly used to obtain more realistic analogues for fossil microfossil assemblages, rather than locally compiled datasets (Gehrels *et al.*, 2001). Transfer functions constructed using modern analogues from a wider range of sites experiencing differing environmental controls (e.g. vegetation, salinity or tidal range) commonly possess greater predictive power since they are capable of quantifying species–environment relationships under a larger range of conditions (Gehrels, 2000).

Recently, Horton and Edwards (2006) have developed a transfer function (SWLI-2006) for tide level based upon a regional training set of contemporary salt-marsh foraminifera from the British Isles. This training set represents an expanded version of the training set successfully employed by

Study area	Tidal range	Salinity	No.	Site description
Alnmouth Marsh, Northumberland	macro (4.2m)	10-30	20	Between coastal dunes and Aln estuary
Cowpen Marsh, Tees estuary	macro (4.6m)	15-25	31	Within a small tidal creek on Tees estuary
Welwick Marsh, Humber estuary	macro (6.4m)	5-35	20	North shore, mid reach of Humber estuary
Thornham Marsh, Norfolk	macro (6.0m)	15-25	24	West end of barrier system on Norfolk coast
Brancaster Marsh, Norfolk	macro (6.0m)	15-25	23	Behind the barrier system on Norfolk coast
Bury Farm, Southampton Water	macro (4.0m)	5-15	16	Upper reaches of Southampton Water
Keyhaven Marsh, Solent	meso (2.1m)	10-30	13	Behind coastal spit in the Solent
Newton Bay, Poole Harbour	micro (1.6m)	20-30	6	Small embayment in lee of Furzey Isl.
Arne Peninsula, Poole Harbour	micro (1.2m)	15-25	12	Small embayment in lee of Round & Long Isl.
Roudsea Marsh, Morecombe Bay	macro (8.4m)	5-25	14	East shore, adjacent to Roudsea Wood
Nith Estuary, Solway Firth	macro (8.4m)	5-25	13	East shore, mid reach of the Nith estuary
Tramaig Bay, Jura	micro (1.4m)	>30	8	Small embayment, little freshwater input
Kentra Bay, Argyll	macro (4.2m)	>30	7	Sizeable embayment, little freshwater input
Rusheen Bay 1, Galway Bay	macro (4.5m)	10-30	9	Western side of tidal creek on Rusheen Bay
Rusheen Bay 2, Galway Bay	macro (4.5m)	10-30	20	East side of tidal creek on Rusheen Bay

Table 3.2 Characteristics of the study areas included in the Horton and Edwards (2006) SWLI-2006 regional transfer function

Study area	Tidal flat	Low marsh	High/middle marsh
Alnmouth Marsh, Northumberland	*Elphidium williamsoni* *Heynesina germanica*	*Miliammina fusca*	*Jadammina macrescens* *Trochammina inflata*
Cowpen Marsh, Tees estuary	*Elphidium williamsoni* *Heynesina germanica* *Quinqueloculina spp.*	*Miliammina fusca*	*Jadammina macrescens* *Trochammina inflata*
Thornham Marsh, Norfolk	*Ammonia beccarii* *Elphidium williamsoni* *Heynesina germanica*	Absent	*Jadammina macrescens* *Trochammina inflata*
Brancaster Marsh, Norfolk	*Heynesina germanica*		*Jadammina macrescens* *Miliammina fusca*
Bury Farm, Southampton Water	*Ammonia beccarii* *Heynesina germanica*	Absent	*Jadammina macrescens* *Miliammina fusca* *Haplophragmoides spp.*
Keyhaven Marsh, Solent	*Ammonia beccarii* *Elphidium incertum*	*Miliammina fusca*	*Jadammina macrescens* *Trochammina inflata*
Arne Peninsula, Poole Harbour	*Ammonia beccarii* *Elphidium williamsoni*	Absent	*Haplophragmoides spp.* *Jadammina macrescens* *Heynesina germanica*
Roudsea Marsh, Morecombe Bay	*Heynesina germanica*	*Miliammina fusca* *Quinqueloculina spp.*	*Jadammina macrescens*
Nith Estuary, Solway Firth	*Heynesina germanica*	Absent	*Jadammina macrescens* *Miliammina fusca*
Tramaig Bay, Jura	*Elphidium magellanicum* *Elphidium williamsoni*	Absent	*Jadammina macrescens* *Trochammina inflata* *Miliammina fusca*
Kentra Bay, Argyll	*Elphidium williamsoni*	Absent	*Jadammina macrescens*

Table 3.3 The dominant foraminiferal taxa of the study areas incorporated in the Horton and Edwards (2006) SWLI-2006 transfer function

Horton *et al.* (1999, 2000). The training set is comprised of 236 samples from 15 sites (Figure 3.4, Table 3.2, Table 3.3) and this combination of modern samples from a wide range of sites with differing physical, biological and hydrographic characteristics maximizes the range of palaeoenvironments that can be reliably interpreted by the transfer function data (Horton and Edwards 2005).

A full description outlining the development procedure of the SWLI-2006 transfer function is given in Horton and Edwards (2006). Only the key points of this method are summarized here. Sample elevation with respect to mean tide level is standardised to account for variations in tidal range between sites, and is presented as a 'Standardised Water Level Index' (SWLI), which is defined in Equation 3.1. (This standardization procedure has been verified by independent analyses from elsewhere (Gehrels *et al.*, 2001)).

$$X_{ab} = [(A_{ab} - MLWST_b)/(MHWST_b - MLWST_b)] \times 100 \quad \textbf{(Eq. 3.1)}$$

where X_{ab} is the SWLI of station or tide level a at site b; A_{ab} is the altitude (m ODB) of station or tide level a at site b; $MLWST_b$ is the mean low water spring tide level (m ODB) at site b; and $MHWST_b$ is the mean high water spring tide at site b (m ODB).

Detrended canonical correspondence analysis of the regional training set indicates a unimodal distribution with respect to SWLI. This form of response has been effectively modelled by weighted averaging partial least squares regression via the program CALIBATE (Juggins and ter Braak, 1997). 'WA-PLS is a modification of the unimodal method 'weighted averaging' (WA), which considers the variance along a single environmental gradient such as elevation (or the SWLI). Unmeasured environmental variables (e.g. salinity) will also influence foraminiferal distributions. This may distort fauna–elevation relationships and cause estimated elevations to differ from their observed values. WA-PLS improves predictions by using any structure present in these differences (called residuals) and, in effect, considers the combined influence of additional environmental variables such as salinity (Ter Braak and Juggins, 1993)' (Edwards *et al.*, 2004b; pp 5).



Figure 3.3 Location map showing the study marshes contributing the surface foraminiferal data used to produce the Horton and Edwards, (2006) SWLI-2006 regional transfer function for tide level. (Taken from Horton and Edwards, 2006)

Some of the samples contained within the modern training set possessed a poor relationship to SWLI. (Such situations arise where the foraminiferal assemblage is strongly influenced by another environmental variable, such as salinity; when the assemblage is not completely autochthonous; or if the assemblage has undergone any post-depositional modification, such as the dissolution of calcareous tests (Horton, 1997)). These samples decrease the predictive ability of the transfer function and have been screened out (Gasse *et al.*, 1995; Jones and Juggins, 1995).

The performance of the SWLI-2006 regional transfer function may be assessed in terms of the root mean square error of prediction (8.6) and the squared correlation (0.71). These provide, respectively, a measure of prediction errors and the strength of the relationship between observed versus predicted values.

3.8.2 Isolation Basins and the Indicative Meaning

The altitude of an index point derived from an isolation basin is reconstructed in a different manner to the altitude of an index point established from a salt-marsh sequence. Instead of using the altitude of the sediment sample itself, SLIPs derived from isolation basins use the altitude of the lowest part of the (rock) sill separating the basin from the sea (Edwards, 2006b). Significantly, this means that all SLIPs sourced from isolation basins are unaffected by compaction since the altitude of the dated sample within the basin sequence is not used in developing the SLIP. Although isolation basins are widely used in sea-level reconstructions, their sedimentary records can sometimes be difficult to interpret (Lloyd and Evans, 2002). This is particularly the case in the isolation basin sequences from regions experiencing relatively slow rates of uplift where the more gradual isolation signature contained in the basin may obscure interpretation of the indicative meaning. The Scottish isolation basin studies of Shennan and co-workers have commonly defined the indicative meaning of the basin by considering the diatomological isolation contact which is said to be near HAT. However, it may not be possible to ascribe a fixed indicative meaning to all isolation basins: The diatomological contact is controlled by salinity and whilst the height of the sill might be a strong influence on basin salinity, terrestrial freshwater input coupled with other factors such as basin size will all serve to modulate the impact of marine influence on basin stratigraphy (e.g. Shennan *et al.*, 1995b; Laidler, 2002).

3.9 RECONSTRUCTING THE ALTITUDE OF MSL

(In this thesis, vertical changes in RSL are expressed in terms of their *altitude* relative to a geodetic Datum (Ordnance Datum Belfast). In contrast, the term *elevation* is used when describing height relative to the local, contemporaneous tidal frame).

Sea level index points are routinely plotted as points on age-altitude diagrams with associated vertical and temporal error terms. These points fix the former altitude of RSL in time, but do not provide any information on the nature of sea-level change between points. The resolution at which past RSL changes can be reconstructed is therefore a function of the number and distribution of data points and the relative magnitudes of their associated error terms. Since RSL records are influenced by vertical land movements, it is important that sea level data are only combined from geographically restricted areas in order to avoid the influence of differential crustal movements (Tooley, 1978).

Error Source	Error range (m)
Id of boundary	±0.01
Measurement of depth in hand corer	±0.01
Angle of bore hole	±0.04
Sampling density *1 bore hole/2m²* *	±0.06
Sampling density *1 bore hole/5400m²*	±0.14
95% Level	±0.30
**A further point to consider is that the accuracy of the lowest altitude of a stratigraphic boundary is dependent on the sampling density and the local surface roughness Makarovic, (1973). Shennan (1986) notes that to avoid altitudinal errors of the magnitude of ±30cm, a borehole density equal to a 30m grid may be required*	

Table 3.4 Quantitative estimates of various altitudinal error sources (Shennan, 1982; 1986)

Having quantified the indicative meaning of the dated sample, local chronologies of RSL change were developed through reference to Equation 3.2.

$$X_i = ALT_i - (RWL_i - MSL) \quad \textbf{(Eq. 3.2)}$$

Where X_i is the reconstructed altitude of mean sea level for sample i, expressed relative to OD(B); ALT_i is the altitude of sample i (m ODB); RWL_i is the reference water level (m) of sample i and **MSL** is the value of (contemporary) mean sea level (m) at the site where sample i is located.

Altitudinal errors for each sample were calculated (where possible/applicable) using Equation 3.3 (Shennan, 1982; 1986), with estimates for some of these error terms contained in Table 3.4:

$$\sqrt{(a^2 + b^2 + c^2 + d^2 + e^2)} \quad \textbf{(Eq. 3.3)}$$

(a = Indicative range, b = error from boundary ID, c = depth measurement error, d = levelling error, e = sampling density error)

Additional vertical error components exist (e.g. sediment compaction and changes in tidal range over time). Attempts have been made to quantify these errors (e.g. Paul and Barras, 1998; Edwards, 2006a). However, for the most part, they remain hard to account for and the implications of this will be discussed further in sections 7.1.2(iii) and (iv).

3.10 SUMMARY

- In this thesis, a sea level index point methodology is adopted to fix the former position of RSL in terms of both age and altitude. This approach requires information on the location, altitude (relative to a levelling datum), age and indicative meaning of a chosen sea-level indicator.

- Along submergent coastlines, the search for information on past sea level is focused upon low energy depositional environments. In particular, salt-marshes (and the sedimentary sequences found beneath them) are chosen since they are the most likely archive of both litho- and/or biostratigraphic evidence of past sea level change.

- Because foraminiferal assemblages display a strong relationship with elevation in the tidal frame, they have been chosen as the preferred sea-level indicator in this study. Two methods have been advanced as a means by which to reconstruct the indicative meaning of fossil (foraminiferal) samples: The 'assemblage zone' approach requires that contemporary marsh samples from a single site are classified into more or less homogenous groups which enable the vertical zonation of the marsh surface to be established. The 'transfer function' approach is based upon a modern training set of marsh

foraminiferal samples and utilizes multivariate statistical analyses to reconstruct the indicative meaning.

- In the absence of foraminifera from the stratigraphic record, diatom analysis has been recommended as a tool to validate sea level index points. However, in such situations a quantitative estimate for the indicative meaning is derived from analysis of the lithostratigraphy.

- Along glacio-isotatically uplifted coastlines, marginal marine basins (and the sediments contained within them) are highlighted as the most likely repositories of past sea level information. Diatom analysis can be used to elucidate the depositional environment of the lithostratigraphic units contained within the basin whilst the height of the basin sill is used to reconstruct the indicative meaning.

- Firm chronological control of the sedimentary sequences contained in both saltmarshes and marginal marine basins is a necessary requirement for the development of reliable sea level index points. Here, AMS radiocarbon dating is chosen to provide age data for the coastal sequences. Although potential sources of error associated with this dating procedure may compromise both precision and accuracy, AMS radiocarbon dating has been adopted instead of other (higher precision) methods because of its ability to deliver age control for the entirety of the postglacial period.

CHAPTER FOUR

THE DEVELOPMENT OF A SEA-LEVEL DATABASE FOR IRELAND

4.1 INTRODUCTION

The importance of understanding RSL changes around Ireland is reflected by a number of attempts to collate published sea level data. Many of these catalogue data from geographically restricted areas, reflecting the particular focus of the research in question (e.g. Carter, 1982a; Penney, 1983; Sinnott, 1999). A notable exception to this is the paper by Carter *et al.* (1989) which presents a series of age-altitude plots from around the Irish coast. This remains the most comprehensive published record of RSL change in Ireland although is now in need of updating to reflect the results produced by studies during the last 17 years. This need has previously been highlighted by researchers seeking to further our understanding of British and Irish Ice Sheet history and associated patterns of glacio-isostatic adjustment (e.g. Peltier *et al.*, 2002; Shennan *et al.*, 2002; 2006a). Accordingly, the development of an Irish sea-level database is a necessary prerequisite for the meaningful validation of glacial rebound model input parameters adopted in the modelling analyses undertaken in chapter 5.

The sea-level database for Britain (Shennan, 1989; Shennan and Horton, 2002) currently holds approximately 1250 sea level index points (from in excess of 2000 records), with over 80 fields of information entered for each data point. Collectively, the information in this database delivers a detailed picture of both spatial and temporal variation in RSL across the UK from a timescale spanning the Holocene and back into the Late Devensian. The majority of the index points included in the database have been collected from low energy (salt-marsh) environments although a relatively small but important group of data are derived from isolation basins, taken primarily from the northwest/ west coast of Scotland (e.g. Shennan *et al.*, 1995a). Work from a plethora of researchers has been included within this catalogue, although the database has been particularly augmented by the work of Shennan (1986, 1992) from the (north) east coast of England; Waller (1994) from the east coast of England; Shennan *et al.* (1995a, 1996, 1999) from the west coast of

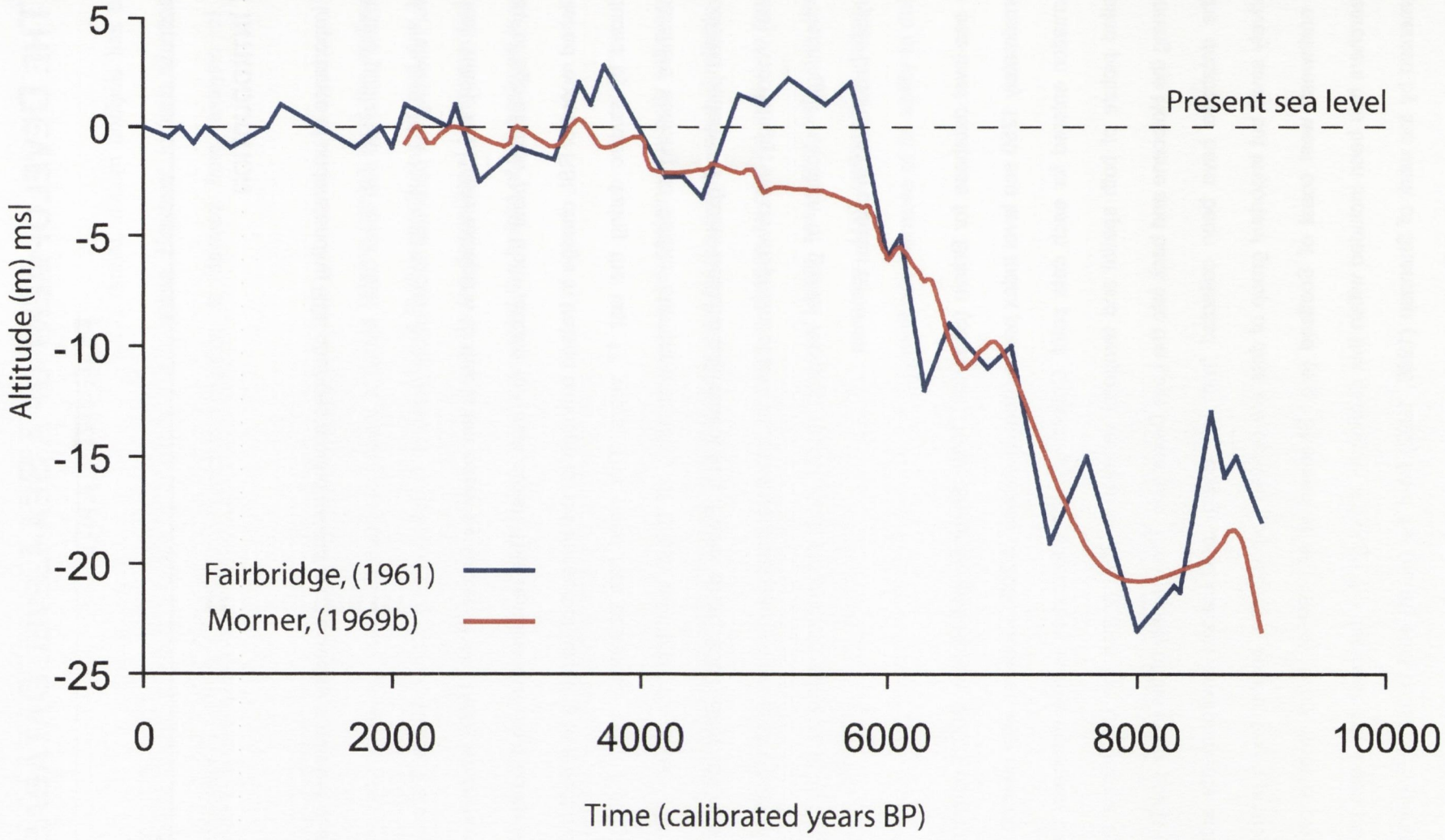


Figure 4.1 Age-altitude graph showing the 'oscillating' Holocene eustatic sea-level trends envisaged by Mörner (1969) and Fairbridge (1961)

Scotland; Tooley (1978) from the north west coast of England and also Long (1992) from the south east coast of England.

The reasons for the inauguration of a standardized sea level database for the UK are closely rooted in the history and development of thought that shaped and guided sea-level research from about 1960 up to the mid 1980's. As Pirazzoli (1996) noted, this period was essentially characterized by three somewhat divergent schools of thought regarding the nature of eustatic sea-level change during the Holocene. These opposing ideas came about as a result of several factors, most notably an incomplete understanding of crustal response to loading coupled with a tendency to underestimate the true extent of the temporal and altitudinal error sources which are inherent to sea-level studies.

In the first school, researchers were seen to adhere closely the work of Fairbridge (1961) who advocated the idea of oscillating eustatic sea level throughout the Holocene (Figure 4.1). Whilst it was accepted that some of these oscillations could be explained in terms of local scale parameters, the overriding tendency was to explain the majority of the intercalations of marine and terrestrial sediments in the geological record in terms of actual variations in eustatic sea level. Such ideas can be clearly seen in the work of Mörner (1969b, 1976b) who concluded that sea-level fluctuations with a range of up to about 2.5 m had occurred in the Holocene. Others such as Godwin *et al.* (1958), Coleman and Smith (1964) argued that a more likely scenario was that of a steady sea level after 5000 BP whilst there existed a third line of thought suggesting sea-level rise took the form of a smooth exponential decay curve throughout the Holocene (Jelgersma, 1961, 1979; Shephard, 1963; Kidson and Heyworth, 1973, 1979; Larsonneur, 1977; Heyworth and Kidson, 1982). In this last scenario, alternations of marine and terrestrial sediments are interpreted in terms of local geomorphic changes and manifest as a result of differential rates of sea-level rise, sedimentation and isostatic rebound. At this time in Ireland, research was still heavily biased towards the geomorphological record of sea-level change. The nature of this work meant that there existed only a tentative engagement by the Irish sea-level community with the debate on Holocene eustatic sea-level change.

The divergent nature of these three ideas augmented by the disparate range of research methodologies employed when collecting sea level information led to a virtual impasse within the

sea-level community. Yet by the mid 1980's, the intensity of the debate had subsided as a result of, *inter alia*, a greater realization of the errors involved in sea-level research, the confirmation of geoidal relief of the Earth and a greater appreciation of the large spatial variation in patterns of glacio isostatic adjustment. The beginnings of this change in outlook can be traced back to the late 1970's and early 1980's, when two significant developments took place: Firstly, the work of Mörner (1976b), Hopley (1978) and Everard (1980) marked the inaugural attempts to compile, on the same age/altitude graph, curves of sea level from different areas. It was only after this that the sea-level community began to appreciate the true extent of variation in sea level that occurred on a regional scale. Of equal significance was the pioneering work of Shennan (1979, 1983, 1986) and Tooley (1978, 1982) with regards the development of a rigorous, standardized methodology for sea-level research carried out along cool temperate, humid, mid latitude coasts (See section 3:1 for further details). This new methodology placed significant emphasis on the continual assessment of errors and their inclusion as error 'boxes' on age/altitude graphs depicting changes in RSL over time (see van Straaten, 1954; Kidson and Heyworth, 1979; Streif, 1979; Shennan, 1982; Tooley, 1982). Thus from the above it becomes clear why the implementation of a sea-level database for the UK was both essential and intrinsically linked to the new pathway being adopted by the sea-level community: Only through setting up a database whose format closely mirrored the accepted methodology for the collection and analysis of sea level data did it become possible to objectively consider both the quality and limitations of the published data in terms of both age and altitudinal errors.

4.2 A Sea-Level Database for Ireland

The database presented in this thesis represents the first attempt at assembling and documenting all published records of Irish sea level in one, standardized database. On viewing the database, it becomes apparent that a corpus of both quantitative and semi quantitative data regarding RSL change is available from a range of locations around the Irish coast, although these data are disparate in nature and of varying quality. Whilst it can be said with some certainty that all published data derived specifically from attempts at reconstructing RSL histories are contained within this database, it may be the case that a number of radiocarbon dated samples from Irish coastal environments are absent. Such records are likely to be associated with

Figure 4.2a

Index Point Tier	Region	Site	Irish Grid Reference	Site Location (Latitude)	Site Location (Longitude)	Labcode	C-14 Age BP	% C-13	Calibrated Age BP (2 σ)	Midpoint	±	Altitude OD(Belfast)	Depth (cm)	Material	Sample position with sediment unit	Microfossil Evidence	Reference Water Level	Indicative Range	Tidal port 1	HAT	MHWST	MHWNT	MTL/MSL	MLWNT	MLWST	LAT	Indicative Meaning	RSL (Relative to msl)	Error Term (m)	Reference
1	4	Ballycotton Bay	W984 644	51.832	-8.024	Q-2384	6120±65		6798-7170	6984	186	-2.5	450	Monocot Peat	Base	Pollen, diatoms	MHWST	80	Ballycotton		1.63	0.6	-0	-1.37	-2.27		2.1	-4.62	0.87	Carter *et al.*, (1989)
2	8	Fergus Estuary, Shannon	136287 170334	52.781	-8.945	GrN-20139	1495±35		1306-1512	1409	103	-1.2		Fishtrap			MSL	50	Mellon Point		2.8	1.4	0.1	-1.2	-2.7		0.075	-1.275	1	O'Sullivan (2001)

Figure 4.2b

Limiting Date Tier	Region	Site	Irish Grid Reference	Site Location (Latitude)	Site Location (Longitude)	Labcode	C-14 Age BP	% C-13	Calibrated Age BP (2 σ)	Midpoint	±	Altitude OD(Belfast)	Material	Microfossil Evidence	Extent of Marine Influence	Tidal port 1	HAT	MHWST	MHWNT	MTL/MSL	MLWNT	MLWST	LAT	Indicative Meaning	Max/Min Height of msl	Error Term (m)	Reference
1	1	Wood Quay, (Dublin City)	O150 342	53.35	-6.27	(Pollen)				7500	500	-2.91	Marine mud/sand	Pollen, foraminifera, ostracoda	MHWST	Dublin North Wall		1.6	0.9	-0.1	-1	-1.8		1.69	-4.6	1	Harris, (1979a)
1	19	Castle Arcade, Belfast	J339 742	54.598	-5.929	St-3066	9130±120		9917-10648	10283	366	-12	Peat		MHWST	Belfast		1.5	0.99	-0	-0.91	-1.6		1.5	-13.5	1	Manning et al., (1970)
2	9	Meelick Rocks, Shannon		52.67	-8.71	GrN-21929	6240 ± 25		7027-7253	7140	113	-2.76	Tree trunk		MHWST	Mellon Point		2.8	1.4	0.07	-1.2	-2.7		2.73	-5.49		O'Sullivan (2001)
2	20	Killard Point, Co. Down		54.31	-5.52	AA 22821	14355±105	-2.70	16187-17047	16617	430	4.36	Foraminifera		HAT	Killard Point	1.96	1.4	0.66	-0.6	-1.94	-2.6		2.6	1.76		McCabe and Clark, (1998)

Figure 4.2 Example database entries taken from **a:** The primary and secondary SLIP tiers of the database; **b:** The Limiting date (type I and II) tiers of the database. (Red infill denotes maximum height of sea level; blue infill denotes minimum height of sea level)

archaeological/historical investigations and the vast body of literature related to these topics precludes an exhaustive search through them.

The database displays sea level data with information spanning over 20 fields, covering fundamental variables such as location, altitude and age, coupled with important supporting data such as tidal information, type of dated material and the indicative meaning of the sample (see Section 4.2.2 for details). These data have been extracted from published literature and a full set of source references is given in Appendix II. The database takes the form of an EXCEL™ spreadsheet and is included as a CD-ROM (Appendix I). It may also be downloaded directly from a website maintained within the School of Natural Sciences, Trinity College Dublin (http://www.naturalscience.tcd.ie/sl_database.php). An example database entry is shown in Figure 4.2a, b.

4.2.1 Data Classes

The contributing sea level data from around the Irish coast are disparate in nature, being derived as they are from material produced by a range of authors employing differing methodologies and often not specifically targeting sea-level reconstruction. Accordingly, the database presented here discriminates four tiers of sea level data on the basis of the range of fields for which information is available, and the associated precision of these data. These four classes represent the different quality of existing sea level data and reflect the extent to which the samples meet the stipulated requirements of a sea level index point. The various tiers are defined as follows:

4.2.1(i) Primary index points

These represent the highest quality sea level data currently available from Ireland, and possess information on location, age, altitude and indicative meaning. Importantly, the relationship between the dated sea-level indicator and the environment in which it formed is quantified and error estimates for age and altitude are included. There are approximately 50 Primary Index Points, and these are largely derived from relatively recent studies in the south of Ireland (Sinnott, 1999; Devoy *et al.*, in press).

4.2.1(ii) Secondary index points

These offer lower quality information on the former position of sea level than Primary Index Points. They are derived from dated sea-level indicators, but one or more of the core variables is unquantified or associated with significant uncertainty. This may include: an absence of accurate levelling data; poor chronological control; limited accompanying microfossil analysis (ambiguous environment); an unclear relationship between the dated material and sea level; association with an erosive contact or stratigraphic unconformity. In addition, sea level data sourced from archaeological evidence are included in this category since their indicative meanings are often poorly constrained (Long and Roberts, 1997).

As a consequence of the uncertainties associated with these data, only tentative inferences on the position of former sea level can be made on the basis of Secondary Index Points alone.

4.2.1(iii) Limiting dates (type I)

These are derived from samples of known age and associated with a known environment that has no quantifiable relationship with sea level. As a consequence, it is only possible to infer whether sea level was above or below a certain altitude at a given time. These limiting dates are derived from material such as freshwater peat and *in-situ* tree stumps.

When utilizing stratigraphic data to reconstruct RSL, it is important to be able to demonstrate that no hiatuses are present. Supplementary evidence (most commonly in the form of microfossil analysis) is required for this. In such situations where detailed lithostratigraphic information is unavailable and a (continuous) transition between marine and terrestrial environments has not been established through accompanying microfossil analysis (e.g. database entry st-3066, Figure 4.2b), the data entry has instead been employed as a limiting date. In some cases, this may represent a loss of information. However, it is preferable to sacrifice precision at the expense of accuracy than *vice versa*.

4.2.1(iv) Limiting dates (type II)

These differ from Type I dates in that they are derived from material whose source environment is unclear or contested. Once again, these data only indicate whether sea level occupied a position

above or below a certain altitude. In addition however, the (possible) allochthonous nature of the dated material advises against attaching great significance to the age of the deposit, a point emphasized by numerous sea-level researchers (e.g. Kidson, 1982). Instead, the date is seen as a 'maximum age' for the deposit since the dated material may have been deposited then reworked several times before finally occupying its current position in the stratigraphic column.

An example of Type II data are radiocarbon dates derived from wood that is not demonstrably *in-situ* (e.g. recumbent trunks in estuarine silt). In addition, a number of AMS dates derived from foraminifera contained with 'glaciomarine muds' are also included since, whilst potentially important sources of sea level data, their origins are the source of much controversy (see McCarroll, 2001 and section 2:3).

There does exist a large body of literature documenting 'raised shorelines' along Ireland's north and east coast (e.g. Synge, 1977a, 1985; Mitchell, 1977; Stephens and McCabe, 1977; Stephens and Synge, 1965). However, such features almost invariably lack firm chronological control and this coupled with their ambiguous relationship to sea level means none of this information is included here.

4.2.2 Database Fields

The database fields are closely related to the core attributes of a sea level index point outlined above. Given the variable quality of the different data classes, not all entries possess data for the entire set of fields.

4.2.2(i) Geographical location

This information is provided in the form of site name, grid reference and latitude / longitude in decimal degrees (to three decimal places). None of the studies documented in the database gave data points in decimal degrees thus a conversion between Irish grid references and latitude/longitude was necessary. This was achieved through the conversion program TRANSDAT (1994) World Wide co-ordinate transformation program. Importantly, all data points contained within the 'primary' index point tier are accurate to within 1km, a standard requirement for SLIPs in the British database. However, not all studies gave precise details of their location and in such

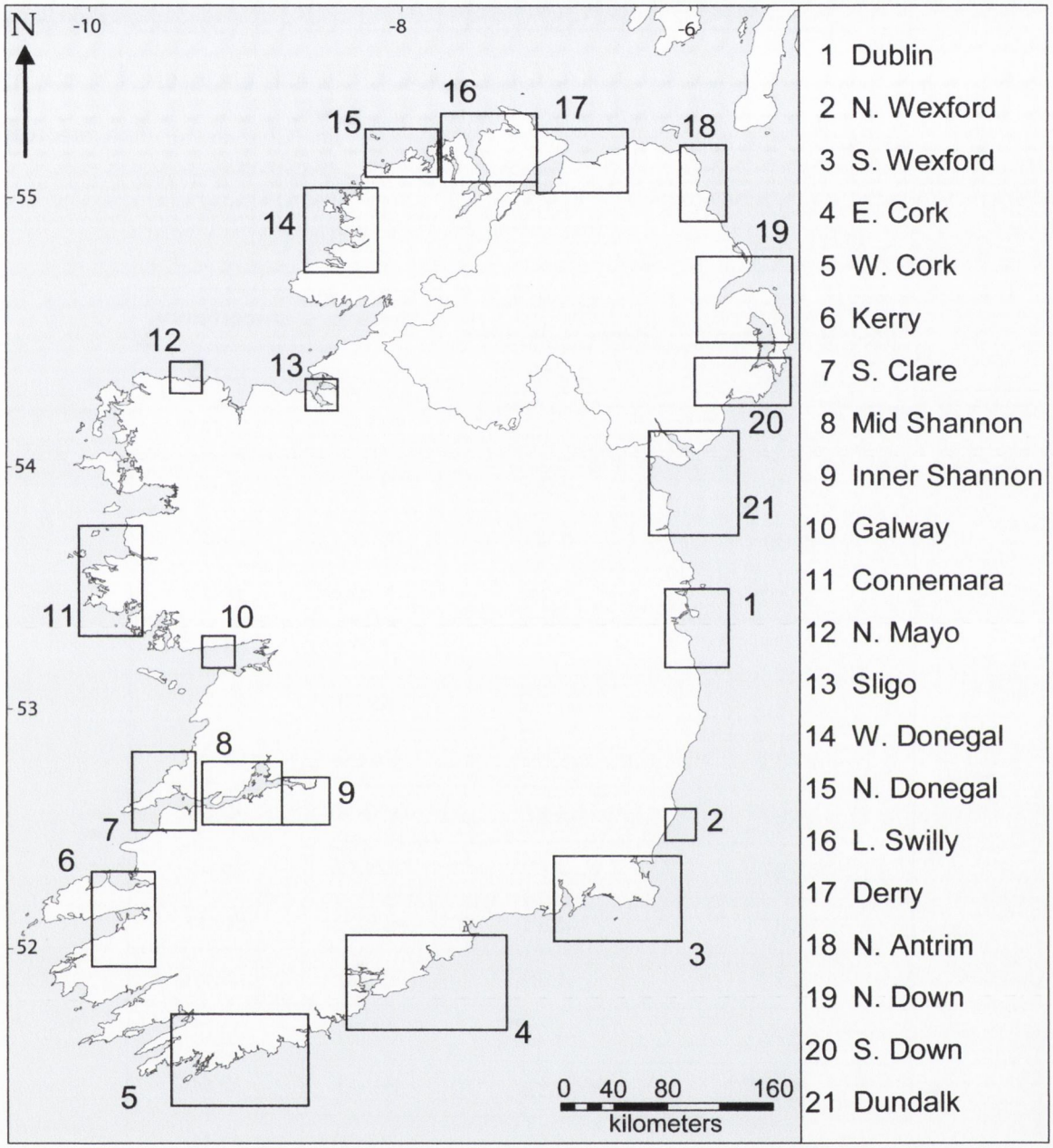


Figure 4.3 The geographical location of the 21 sea-level database regions around Ireland

cases, it is not possible to guarantee their documented accuracy to within 1km. However, the occurrence of such data points are confined to the 'secondary index point' and 'limiting date' tiers and the lower resolution nature of this data means a slight reduction in the precision of its given geographical location is arguably of only limited significance. For the purposes of displaying information and to facilitate their use in glacial rebound analysis (Chapter 5), data are also assigned to a broad geographical region (Figure 4.3).

Latitude (decimal degrees)	Longitude (decimal degrees)	Delta R	Uncertainty
55.03	-4.30	14.00	13.00
55.97	-3.15	24.00	57.00
52.00	-6.00	21.00	54.00
53.00	-6.00	-17.00	45.00
54.17	-5.00	-21.00	47.00
54.17	-5.00	30.00	63.00
55.83	-6.00	-65.00	46.00
55.83	-6.00	-74.00	52.00
55.87	-4.93	-110.00	42.00
55.87	-4.93	-124.00	54.00
55.87	-4.93	-27.00	80.00
55.73	-4.88	75.00	38.00
55.73	-4.88	89.00	33.00
55.97	-2.92	66.00	32.00
55.97	-2.92	-27.00	41.00
53.25	-4.50	-4.00	49.00
55.00	-5.00	48.00	50.00

Table 4.1 Regional marine reservoir corrections. (Taken from Harkness, 1983)

4.2.2(ii) Age

Most entries in the database have age information provided by radiocarbon dating. The database contains the sample laboratory code, radiocarbon age and calibrated calendar date calculated using CALIB 5.0.1 (Stuiver *et al.*, 2005). The following options were chosen: a laboratory multiplier

effect of 1; 95% confidence limits and their dataset IntCal04 (which is confined to 0-26 000 cal BP). This dataset is recommended for most non-marine samples and is based on dendrochronologically dated tree-ring samples which cover the period from 0 - 12 400 cal BP. For the time interval 12 400 – 26 000 cal BP, data from marine records (corals and foraminifera) are converted to the atmospheric equivalent with a site-specific marine reservoir correction to provide terrestrial calibration.

In instances where marine samples (such as shells and foraminifera) have been dated, the dataset Marine04 has been employed. The marine calibration dataset incorporates a time-dependent global ocean reservoir correction of about 400 years but to accommodate local effects, the difference Delta R in reservoir age of the local region of interest and the model ocean should ideally be determined (Stuiver and Reimer, 2004). Where possible, this was achieved through reference to Harkness (1983) (Table 4.1).

Whilst some of the radiocarbon dated samples included in the database are corroborated by pollen chronostratigraphic data, the vast majority have no such supporting evidence. This means factors such as rootlet penetration of the sample or isotopic fractionation may lead to either over or under estimation of the true sample age. As Peltier *et al.* (2002) pp 450 note; "Without multiple-dated samples, sometimes on different constituents of the sediment, or other proxy measurements of age, it may be difficult to identify the effects of these contaminants". It is also the case that if erosion has occurred between a terrestrial-marine contact in the stratigraphic column, any age estimate for the date of transition between the two environments may substantially over estimate the true age of the event. To guard against this, detailed microfossil analysis can be employed to demonstrate a continuous transition across the stratigraphic contact. This has been carried out for all samples contained within the primary index point tier. It should be noted that for the case of limiting dates, an erosive contact is not important since no attempt is being made to fix the time of a transition between a marine and terrestrial environment.

4.2.2(iii) Altitude

As mentioned in chapter two, a particular issue that has introduced additional complication into comparing Irish RSL records is the multitude of datums to which various sample altitudes have

been related to. For the purposes of this study, all heights are standardized to OD (Belfast). Importantly however, whilst 0m OD (Belfast) may be a mean sea level for Belfast, values for msl vary spatially as a result of *inter alia* differing tidal regimes, the topography of the geoid, coastline configuration and prevailing meteorological conditions (Van de Plassche, 1986). Consequently, values of local mean sea level were obtained through reference to the Admiralty tide tables (1997). Where values for msl were absent, mean tide level was calculated through summing readings of MHWST, MHWNT, MLWNT and MLWST then averaging the result.

Type of contact	Reference water level	Indicative range	Source
Base of basal peat	MTL to MHWST	?80cm	Shennan (1982; 1986; 1994)
	MHWST+20cm	60cm	Zong and Tooley (1996)
Fenwood/carr below marine clastics	MHWST-20cm	40cm	Zong and Tooley (1996)
+ Reedswamp below marine clastics		20cm	Shennan (1994)
Carr above marine clastics	MHWST+80cm	60cm	Zong and Tooley (1996)
	M^1-10cm to M^1-20cm	20cm	Shennan (1994)
-above Phragmites or saltmarsh	MHWST	20cm	Shennan (1982; 1986)
Reedswamp above marine clastics	MHWST+60cm	40cm	Zong and Tooley (1996)
	M^1-10 to M^1-20cm	≥20cm	Shennan (1994)
- above saltmarsh	M^1-20cm	20cm	Shennan (1986)
Intercalated peat	M^1-10 to MHWST-20cm	70cm	Shennan (1994)
Organic clay silt (salt marsh) below marine clastics	MHWST-20cm	40cm	Zong and Tooley (1996)
Within basal peat (middle of layer)	MHWST to MTL	>80cm	Shennan (1994)
	infer from stratigraphy	70cm	Shennan (1986)
Fish trap (Archaeological evidence)	MTL	50cm	Parnell *(pers comm.)*
'Peat' above/below organic silt	MHWST	50cm	Tooley (1978); Shennan, (1982)

Table 4.2 Reference water levels (and associated indicative ranges) of sea-level indicators used in the database

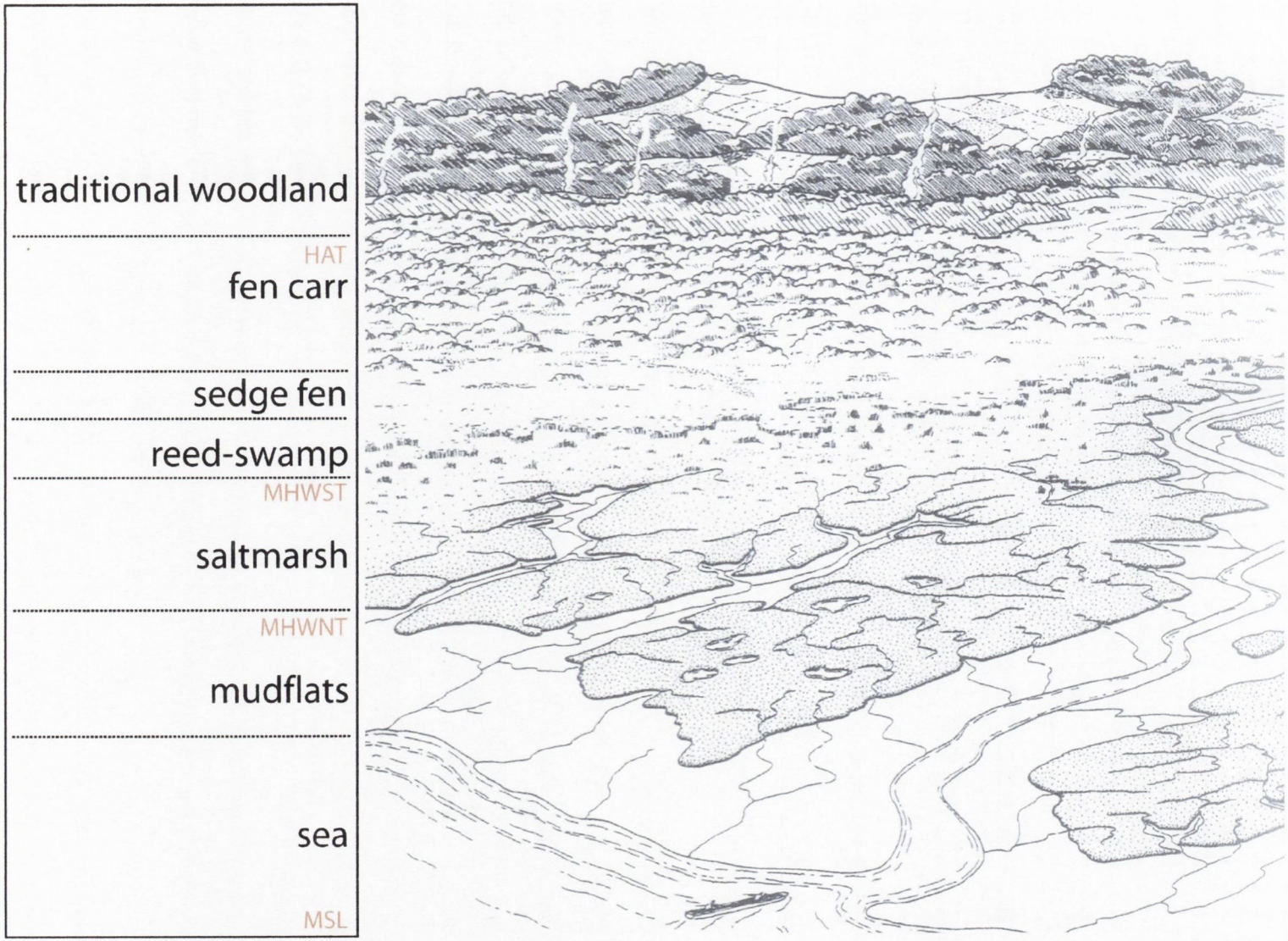


Figure 4.4 Vegetation succession in the coastal environment. (Adapted from O'Sullivan, 2001)

Indicative Meaning

Where available, the database provides information on the inferred indicative meaning, the nature of the evidence from which this is derived, and the local tidal parameters used in reconstruction. For Primary Index Points, all indicative meanings are derived from a combination of lithostratigraphic and biostratigraphic data. Table 4.2 provides a list of commonly employed sea-level indicators with their associated indicative meanings and original publications. The contemporary distribution of many of these sea-level indicators can be seen in Figure 4.4.

With regards to the limiting dates, all samples from terrestrial environments are said to have formed no lower than MHWST. However, when considering freshwater peat deposits, this may or may not be the case and a degree of ambiguity exists within the literature as to the maximum seaward limit that freshwater peat horizons can develop. Shennan *et al.* (1995a) note that fen peats, representing terrestrial or freshwater vegetation, have their seaward limit between MHWST and HAT. However, in a later paper, (Shennan *et al.*, 2000b, pp 282) stated that "such (fen) peats may form in a wide vertical range depending on local palaeogeography, especially their relationship to active tidal channels and the groundwater table. Therefore, the indicative range (for these dates) is from above MHWST, where freshwater peat formation is controlled by tidal influence, to just below MTL, where groundwater level is the controlling factor (Godwin, 1940; Shennan, 1982)". Thus in this latter paper, the maximum elevation of tide-level is calculated as the midpoint between MHWST and MTL. These differences in interpretation arise as a result of the varying influence of local parameters, the nature of the over/underlying sediments and also the position of the dated sample in the stratigraphic column. The latter point is significant since sea-level indicators taken from basal peats are commonly related to the position of the terrestrial water table and not to sea level. It is the case that the relationship between basal peats and tide levels can be heavily influenced by groundwater table gradients, the floodbasin effect and also river water levels (Van de Plassche, 1982). Thus from all of the above it is apparent that there is no fixed seaward limit for terrestrial peat growth. Instead, local parameters have a strong influence, enabling in certain extreme cases, for freshwater peat to form as low as MTL. A thorough understanding of the local environment in which a dated peat sample is taken from is mandatory in order to be able to state with certainty that the indicative meaning for a given freshwater peat lies below MHWST. Such details are commonly lacking in the Irish literature thus in the Irish sea-level

database, a conservative estimate of MHWST is given as the most seaward limit for terrestrial peat inception. Whilst all dated deposits from terrestrial environments have been said to have formed at or above MHWST, where deposits have been described as 'inter-tidal organics' they are interpreted as having formed at or above MHWNT since this generally marks the boundary between organic-rich inter-tidal deposits (e.g. salt-marsh) and inter-tidal to sub-tidal minerogenic sediments. Conversely, shells incorporated within wholly marine deposits are seen as representing deposition below MHWNT. (It should be emphasized however, that sea level data sourced from shells is particularly ambiguous since in most instances it is not known whether the samples were *in situ*, in which case they provide lower limits to sea level, or whether they were deposited by wave and wind action, in which case they provide upper limits only (Lambeck *et al.*, 1998)).

Codes	Description
0	No tendency yet given
1	Good positive tendency of sea-level movement. From a transgressive contact, or a stratigraphic or biostratigraphic change, indicating a positive sea-level tendency. There is no indication of erosion
2	Poor positive tendency of sea-level movement. From an apparent transgressive contact, but where either erosion is suspected or possible, or there is no supporting biostratigraphic evidence from the site
3	Base of basal peat dated ('basis peat')
4	Sample from within the basal peat, but not from the base, or from the transgressive contact (i.e. not 1,2 or 3)
5	Good negative tendency of sea-level movement, from a regressive contact, or a stratigraphic or biostratigraphic change indicating a negative sea-level tendency. There is no indication of a hiatus in sedimentation
6	Poor negative tendency of sea-level movement. From an apparent regressive contact, but where either a hiatus is expected, or there is no supporting biostratigraphic evidence from the site
7	Sample from any part of a sequence beyond the limit of marine sedimentation
8	Sample from middle/whole of an intercalated peat where no single tendency can be identified. It represents a minimum age for a negative tendency followed by a positive tendency
9	Sample cannot be reliably related to a specific sea-level tendency in terms of either age, altitude or indicative meaning. But is within marine limit

Table 4.3 Numerical codes for sea level tendency values. (Taken from Shennan, 1994)

4.2.2(iv) Sea Level tendency

Established sea-level literature often makes reference to sea level tendencies (Shennan, 1980, 1982, 1983; Tooley, 1982). A positive sea level tendency is assigned to a sea-level indicator where

it indicates an increase in marine influence (and *vice versa* for a negative sea level tendency) irrespective of its cause. It is important to recognise that sea level tendencies are not synonymous with a rise or fall in RSL and changes in tendency may occur in the absence of sea-level movement. Shennan (1994) defined 9 separate sea level tendencies to account for the entire range of stratigraphic contacts (Table 4.3) and where possible, these are included in the database along with its mode of establishment. However, in general there are currently insufficient data points to perform meaningful tendency analysis (Shennan *et al.*, 1983).

4.3 Data Limitations

Users of the sea-level database must be aware of some fundamental limitations in the data it contains. In the first instance, the database is compiled from published material and assumes that the information included within these publications is accurate and reliable. Full bibliographic details are provided in Appendix II, and the user is advised to refer to the original publications for further details of any data they select to employ.

More specific limitations are associated with the differing classes and fields of data, reflecting wider limitations associated with existing sea-level methodology. Below, the key points to be aware of are highlighted, along with reference to publications which deal with these issues in more detail.

4.3.1 Location, Data Distribution and Regions

The quality of a local record of RSL change is dependent upon the accuracy, precision and distribution (in time and space) of sea level data. This is highly variable around the coast of Ireland with the result that sea-level changes in certain parts of the country are still associated with considerable uncertainty.

The combination of sea level index points from different sites into regional plots assumes that inter-site differences in RSL change are not substantial. This assumption becomes increasingly prone to error as the regions become larger, or the individual records are analysed at higher resolutions. It will also be more suspect in areas with complex hydrographic regimes and coastal geometries since significant inter-site variability in tidal parameters (and hence indicative meanings) may be produced. The current 21 regions reflect areas with broadly similar modelled RSL histories derived

from recent attempts to model the glacial isostatic adjustment (GIA) process (Chapter 5). These regions are therefore suitable for input to geophysical models with similar, comparatively coarse, spatial resolutions. Conversely, care should be taken in extrapolating the general results from these broad regions to detailed, site specific studies.

4.3.2 Inferred Age

The reliable interpretation of age data rests upon the assumption that the dated sea-level indicator reflects the timing of the sea-level change of interest. This assumption has been explored in section 3.7.

4.3.3 Altitude

Errors in altitude can be introduced during sample collection as a consequence of limitations in the precision of the surveying equipment used, uncertainties inherent within the benchmark network, and errors associated with the measurement of lithostratigraphic contacts in the field (see Shennan, 1982, 1986). An additional potential source of error comes from uncertainties associated with establishing the indicative meaning of the sample (see below). For material recovered by hand coring, estimates of the magnitude of these various error sources have been compiled by Shennan (1982, 1986). These estimates, in addition to information pertaining to the indicative meaning are summarised in Tables 4.2 and 3.4.

In the database, a standardised vertical error term after Shennan (1982, 1986) is derived from equation 4.1:

$$\sqrt{(a^2 + b^2 + c^2)}$$

(Eq. 4.1)

(a = Indicative range, b = error from boundary ID, c = depth measurement error)

Where a published estimate of total altitudinal error exceeds that calculated from Eq. 4.1, the former is adopted. Some entries in the lower quality tiers lack precise levelling data, having been related, for example, to 'around MHWST'. In these instances a default 'indicative' error term of ±1m is assigned. It should be noted that some of the data presented by Sinnott (1999) contained within the primary index point tier came from sites which were not within surveying distance of a levelled

benchmark. In such instances, altitude was gauged via surveying to the level of high water then relating back to Admiralty Tide Tables. This procedure is less accurate than surveying to a local benchmark but the absence of associated error estimates specific to this method meant such index points were assigned the same error estimates as those points which had been conventionally levelled. In the case of dated samples where the indicative meaning is ambiguous or where the environment of deposition is contested (e.g. Limiting Dates series II), no attempt has been made to allocate a vertical error term. Consequently, the absence of an error term indicates an error of unquantified magnitude rather than the absence of any vertical error.

In addition to the altitude errors described above, a further unquantified error may be introduced by post-depositional compaction of the sediment column, either under its own weight or as a consequence of subsequent loading by water or an overlying sediment burden. Whilst the potential influence of compaction on Holocene RSL records is long established and widely acknowledged (e.g. Jelgersma, 1961; Terzaghi and Peck, 1967; Greensmith and Tucker, 1971, 1973; Tooley, 1978; Heyworth and Kidson, 1982; Shennan, 1986; Allen 1996; Haslett *et al.*, 1998a; Shennan *et al.*, 2000b; Edwards, 2006a), it has commonly been set aside due to the lack of a formal means for correcting for its influence and the difficulty in calculating changing sediment pore water pressure through time (Allen, 2000b; Shennan and Horton, 2002; Brain, 2006). Compaction serves to lower the reconstructed altitude of former RSLs and will be greatest where index points are established from thick intercalated sedimentary sequences. Index points should therefore be interpreted in light of the stratigraphic data supplied in the database.

Indicative Meaning

Each primary index point contained within the database has an associated indicative meaning which has been constructed with reference to contemporary tidal characteristics. Since the majority of data points in the database are collected some distance from an established tide gauge, site-specific tidal parameters have to be extrapolated or interpolated from the closest available records. This will introduce an error term which may only be accurately corrected by on-site data-logging of tidal conditions.

The indicative meaning method inherently assumes that significant changes in tidal range have not occurred at the study site. In reality, many regions will have experienced alterations in tidal parameters which will introduce error into reconstructions (e.g. Austin, 1991; Shennan *et al.*, 2000c). These errors are likely to be greatest in macro-tidal regions such as the Shannon Estuary and Dundalk Bay, or in areas that have experienced pronounced changes in coastal geometry over time. Spatial variation in the contemporary tidal range theoretically should not compromise the accuracy of the indicative meaning, since it is always expressed as a function of tidal parameters instead of a large constant (Shennan *et al.*, 2002). However, where the tidal range may be extremely large or small, the accuracy of the indicative meaning and associated indicative range may be called into question (e.g. Smith and Morgan, 1989; Scaife and Long, 1995 in the Bristol Channel/Severn).

Lastly, some researchers have questioned the level of precision inferred by Shennan (1982; 1986; 1994) for lithostratigraphic sea level indicators (e.g. Horton *et al.*, 2000). It has been suggested the values for the indicative range included in Table 4.2 are too precise and should be increased. However, a detailed study reassessing the quantitative estimates for the indicative meaning and associated indicative range made by Shennan has yet to be undertaken.

	Number of data points			
Region	*Primary Index Points*	*Secondary Index Points*	*Limiting Dates (Type I)*	*Limiting Dates (Type II)*
(1) Dublin		1	3	3
(2) North Wexford	1			
(3) South Wexford	11	1	3	1
(4) East Cork	10	1		
(5) West Cork	13	1	3	1
(6) Kerry	1		10	
(7) South Clare		1	4	
(8) Middle Shannon		3		
(9) Inner Shannon			1	8
(10) Galway			1	
(11) Connemara			17	
(12) North Mayo				8
(13) Sligo			1	
(14) West Donegal	8		10	1
(15) North Donegal	3		3	3
(16) Lough Swilly			1	7
(17) Derry			4	6
(18) North Antrim			5	4
(19) North Down		1	10	5
(20) South Down	2	2	4	4
(21) Dundalk			4	12
	49	**11**	**84**	**63**

Table 4.4 Frequency distribution of Irish RSL data in each geographical region of the database

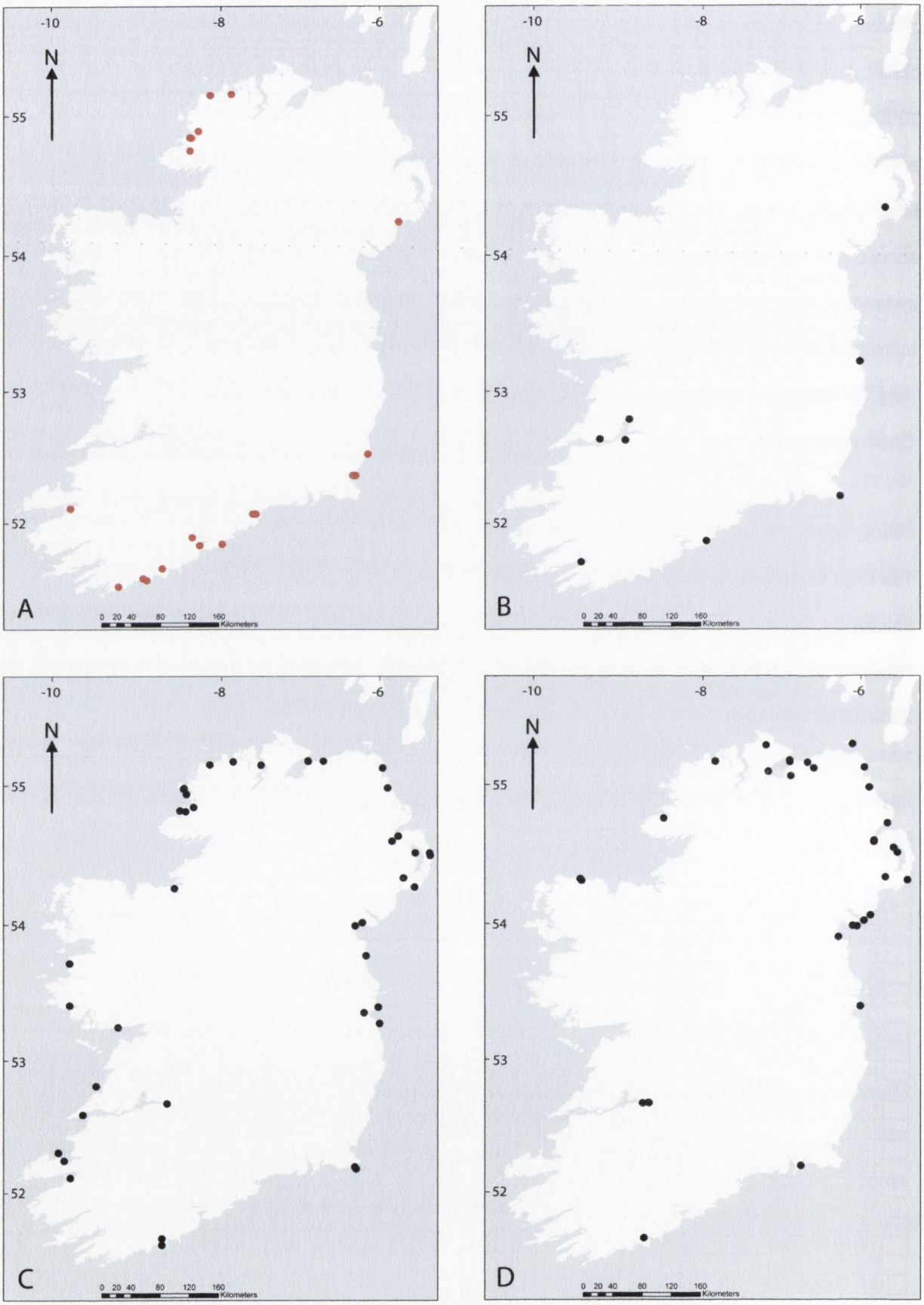


Figure 4.5 The spatial distribution of Irish RSL data included in the database. **A** Primary index points; **B** Secondary index points; **C** Limiting Dates Type I; **D** Limiting dates Type II

4.4 Sea-Level Changes Around Ireland – An Update

4.4.1 Data Distribution

The spatial distribution of sea level data from each data class is presented in Figure 4.5 and summarised in Table 4.4. A clear spatial bias in the quality and quantity of data is apparent. The highest quality sea level data (Primary Index Points) is concentrated along the southern Irish coast. This is in marked contrast to much of the eastern and western coastlines which are virtually devoid of reliable sea level index points. This regional bias primarily reflects the availability of sedimentary sequences suitable for establishing sea level index points. In comparison with the UK, surprisingly few intercalated sequences of terrestrial and marine sediments have been reported from the east and west coast of Ireland, and many of these are localised and of limited thickness (e.g. back-barrier deposits). Similarly, unlike a number of studies from Scotland (Shennan *et al.*, 1993, 1995, 2000a), there is a lack of isolation basin data from the north of Ireland. It is likely that this also reflects the relative scarcity of suitable sites, since modelling results suggest potential sites will be restricted to specific stretches of the Ulster coastline (Lambeck, 1996; Lambeck and Purcell, 2001). Figure 4.6 shows the temporal distribution of different classes of sea level data in 1000 year intervals. From this figure it is clear that, in addition to the spatial variability outlined above, there is also a marked bias in the ages of available data with most of the sea level index points dating to the mid or late Holocene. This reflects the fact that southern Ireland has experienced rising RSL over the past 10 000 years and the rapid rate of early Holocene eustatic sea-level rise means early Holocene coastal sequences will now be found at considerable depth or in offshore contexts.

4.4.2 Example Data and Their Implications

Figure 4.7 shows example age-altitude plots from five regions distributed around the coast of Ireland. These plots show the variability in data quality in space and time and reveal the general patterns of RSL change for these regions. As noted by previous workers, a generally coherent and consistent pattern of change is revealed by the data, showing the contrast between long-term RSL rise in southern Ireland with the more complicated, non-monotonic changes in north-east Ireland (e.g. Carter, 1982a; Carter *et al.*, 1989a). However, inspection of the classes of data used reveal that, for many areas, the pattern of RSL change is only broadly delimited by limiting dates, and does not fix the altitude of past sea level in time and space. This is particularly the case for the RSL

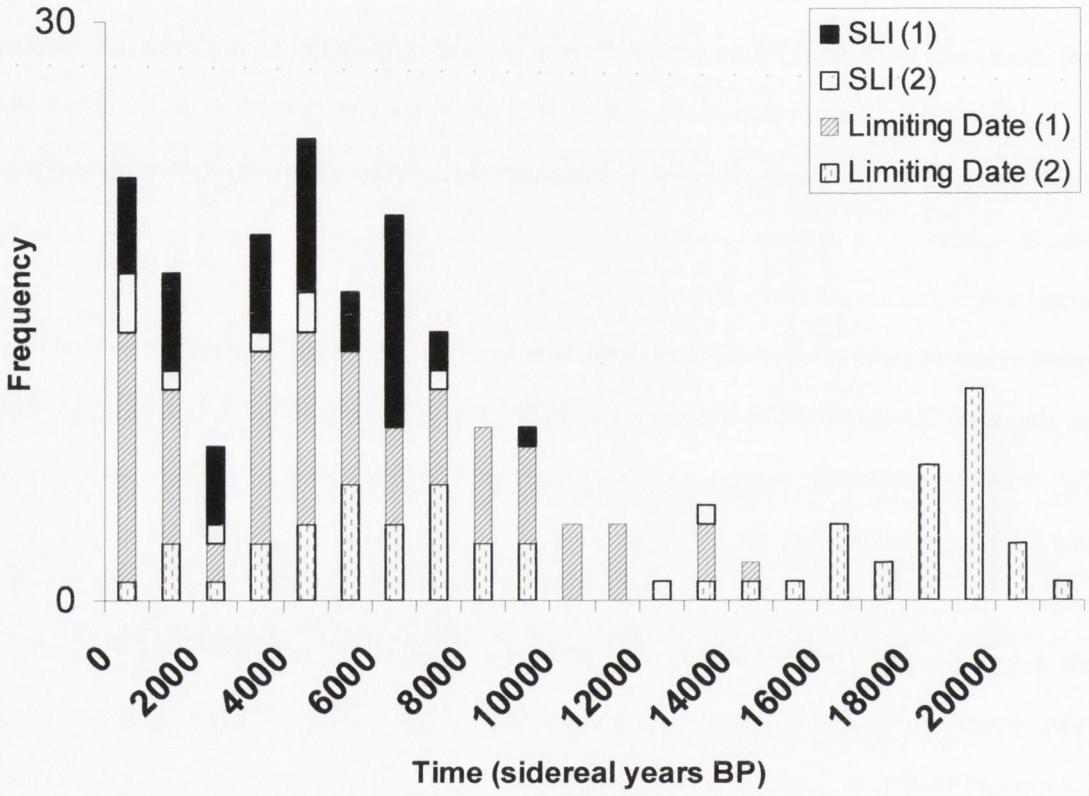


Figure 4.6 The temporal distribution of Irish RSL data included in the database

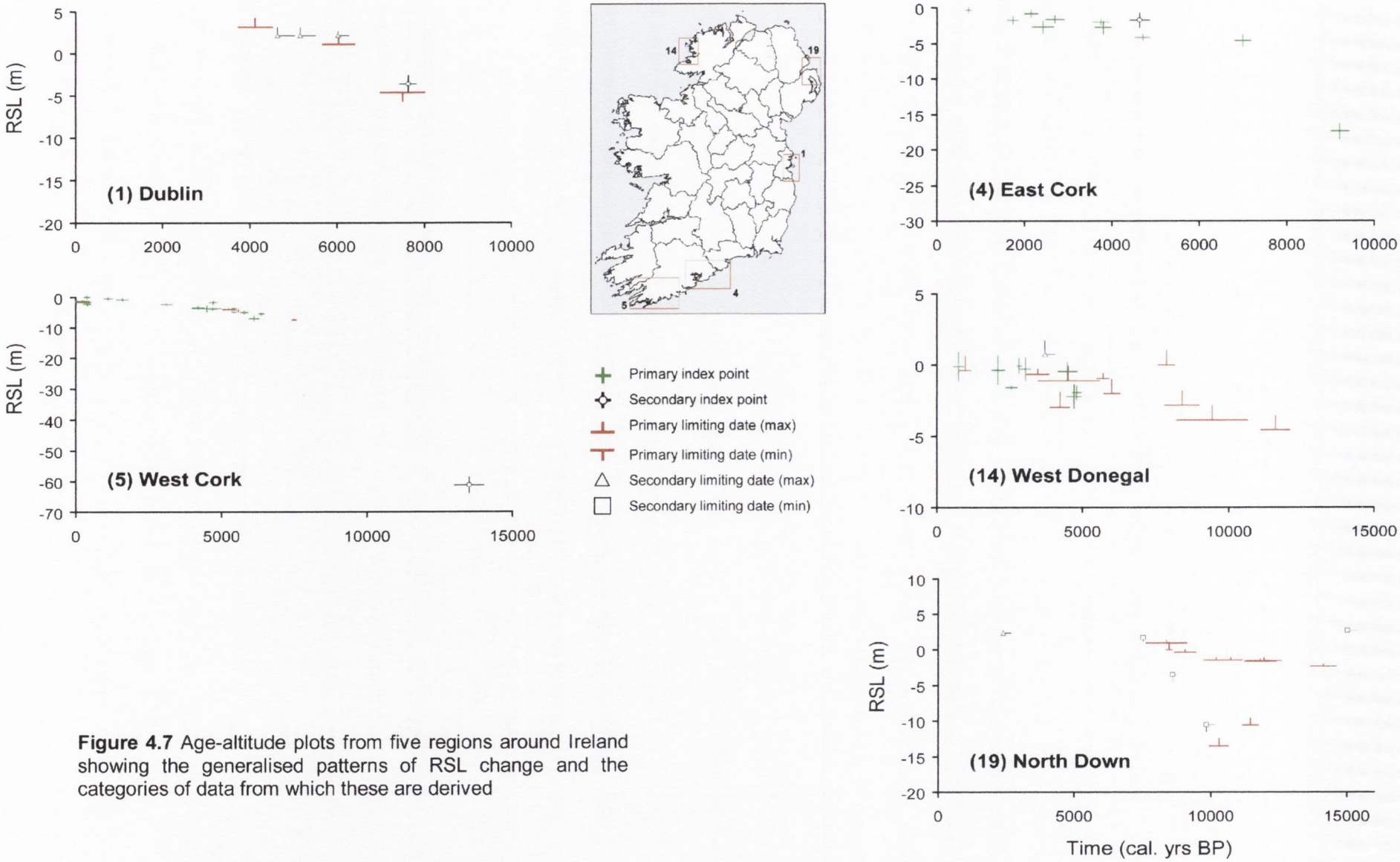


Figure 4.7 Age-altitude plots from five regions around Ireland showing the generalised patterns of RSL change and the categories of data from which these are derived

curves for the north of Ireland compiled by Carter (1982a) which have frequently been referred to in papers considering aspects of coastal palaeoenvironmental change (e.g. Lambeck, 1996). Since these hypothesised 'sea-level curves' are primarily constructed from limiting dates, the magnitude and timing of specific features (e.g. the Holocene highstand or early Holocene lowstand) cannot be accurately constrained and are potentially erroneous (e.g. Orford *et al.*, 2003).

In the past, uncertainties in the limitations of the Irish RSL data (augmented by the disparate nature of the data itself) have resulted in some of the best data from Ireland being underutilized. A case in point is that of the glacial rebound modelling efforts of Lambeck (1996) for Ireland. Here, large vertical error components have been added to the empirical data in order to yield a better fit with model RSL predictions. Consequently, the data are of little value in constraining model input parameters such as ice sheet thickness. Taken in the overall context of the available data, this approach by Lambeck was justified yet the database presented here should alleviate such problems.

4.5 SUMMARY

- In the past, efforts to model sea/land-level behaviour from Ireland have been hampered by uncertainties in the reliability of field data (e.g. Lambeck, 1996). The database presented here is a first attempt at compiling a standardized sea-level database for Ireland, which is consistent with established protocols for the evaluation of sea level information.

- The available data are of varied quality, yet limitations associated with this are mitigated by applying a rigorous appraisal of the errors inherent within the data and the consequent adoption of a quality tier system.

- For all of the 'primary index point' entries in the database, the sampling design employed by the individual researcher was specifically defined to test particular hypotheses relating to trends in RSL. Yet most of the records contained in the database come from studies of different aspects of Quaternary environmental change but these still contribute to the analysis of RSL change by constraining maximum/ minimum altitudes of sea level in both time and space.

- However, of the 200 records in the database, less than 25% occupy the primary SLIP tier and a large proportion of the existing data must be considered as 'limiting' in nature. In the past, attempts have been made to treat these limiting data as index points, which in some instances has resulted in misleading conclusions being reached.

- Arguably the biggest weakness in the dataset at present is the spatial and temporal bias of the index points. The highest quality data are almost exclusively found along the south coast and are of mid/late Holocene age.

CHAPTER FIVE

Postglacial Relative Sea Level Observations From Ireland and Their Role in Glacial Rebound Modelling

5.1 INTRODUCTION

Recently, a number of researchers have modelled observations of glacio-isostatic adjustment (GIA) from the British Isles (e.g. Lambeck, 1991; Lambeck, 1993a; Lambeck, 1993b; Lambeck, 1995; Lambeck., 1996; Peltier *et al.*, 2002; Shennan *et al.*, 2000a; 2002; 2006a). These efforts have been facilitated by the availability of long time-series records of RSL change in Britain which, through the process of geophysical modelling, can be distilled to infer information regarding former ice extent, thickness and melting history, as well as viscosity structure within the mantle. However, in the past, the fragmentary nature of the Irish RSL record, allied with a paucity of observational constraints on the Irish Ice Sheet, has deterred researchers from using these data to constrain either the ice history or sub-surface viscosity structure for this region. Notable exceptions to this are the works by Lambeck (1996) and Lambeck and Purcell (2001), which utilize existing records of RSL to help reconstruct the vertical limits of the Irish ice sheet.

Since these publications, there have been a number of pertinent developments that mean it is now timely to re-evaluate the contribution Irish observational data can make to modelling GIA in the British Isles. Firstly, the body of available Irish sea level data has grown since the work of Lambeck and co-workers. This is augmented by the recent work of Shennan *et al.*, (2006b) which delivers new, high resolution RSL records from sites immediately adjacent to Ireland (Islay and Knapdale) that date back to c. 16 000 BP. Data from these two sites provide both a useful and unique constraint on model reconstructions of the Irish Ice Sheet.

Secondly, observations of postglacial RSL from Ireland have now been screened and categorized in a new Irish sea-level database (chapter 4). A similar database for the UK (Shennan and Horton,

2002), has been integral to previous GIA modelling studies from Britain (e.g. Lambeck, 1995; Shennan *et al.*, 2002; Shennan *et al.*, 2006a).

Thirdly, new data on the spatial extent (e.g. O'Cofaigh and Evans, 2001; Hiemstra *et al.*, 2006; Sejrup *et al.*, 2005) and thickness (e.g. Rae *et al.*, 2004; Bowen *et al.*, 2002) of the Irish ice sheet since the last glacial maximum (LGM) are now available. Furthermore, a number of Accelerator Mass Spectrometry (AMS) radiocarbon dates derived from marine muds have, for the first time, delivered chronological control on the post-LGM deglacial phase (e.g. McCabe and Haynes, 1996; McCabe and Clark, 1998, 2003; McCabe *et al.*, 2005). In addition, field evidence now indicates that ice from the British Isles coalesced with Fennoscandian ice during the Late Devensian (e.g. Sejrup *et al.*, 1994, 2005; Svendsen *et al.*, 2004; Carr *et al.*, 2006), producing a more laterally extensive ice sheet than envisaged by Lambeck (1996).

Finally, Shennan *et al.* (2006a) have recently developed a revised GIA model for the British Isles (hereafter referred to as BIM-1), which satisfies key glaciological, seismic and sea-level constraints from both near- and far-field localities. This new model incorporates a more realistic assessment of ice sheet mass (and associated crustal loading) than previous studies by accounting for the effects of topography/bathymetry on ice thickness (Milne *et al.*, 2006). Importantly, whilst this model has an Irish ice sheet component, it is not calibrated using any of the Irish RSL data. Consequently, the new dataset of Irish RSL reconstructions can provide an independent test of its Earth and ice components.

In this chapter, the RSL simulations of BIM-1 are evaluated against the independent Irish dataset. It is demonstrated that significant misfits exist between observations and predictions for certain portions of the Irish coastline, indicating limitations in one or more components of BIM-1. The source of these limitations is investigated by examining the effects of modifying the Irish ice sheet and Earth parameters of this model. In light of these results, a revised model of the Irish ice sheet is produced that satisfies both RSL and glaciological field constraints. The results also indicate that minor modifications of the Earth model are required to reconcile RSL predictions with observations from Ireland and the west coast of Britain.

5.2 Modelling Relative Sea-Level Change

Glacio isostatic modelling can be seen as a means of simulating RSL changes forced by mechanisms associated with deglaciation. Such models are primarily based on the mathematical analysis of the deformation of a viscoelastic Earth produced by loading of the surface. Observations of RSL provide key constraints on such models and (because of the intimate relationship they have with GIA and eustatic sea level), through the process of geophysical modelling they can be distilled to infer information regarding former ice extent, thickness and melting history, as well as data pertaining to the physical structure of the Earth and its response to loading (e.g. Milne *et al.*, 2005). Geophysical models that simulate RSL change associated with the build up of mid-high latitude ice sheets have a long history of development (e.g., Walcott, 1972; Peltier, 1974; Farrell and Clark, 1976; Peltier and Andrews, 1976; Nakada and Lambeck, 1987; Mitrovica and Peltier, 1991; Johnston, 1993; Milne and Mitrovica, 1996; Milne *et al.*, 1999) with their initial inception in the early/mid 1970's introducing something of a conceptual revolution in RSL research (Pirazzoli, 1996). On the basis of the mathematical analyses presented within these papers, for the first time it became possible to predict RSL histories that ought to be observed at any point on the Earth's surface and to compare these predictions with dated records of the history of RSL change since the LGM (Peltier, 2002).

The general theory pertaining to the application of geophysical models for the prediction of RSL change, and their validation *via* the use of radiocarbon-dated SLIPs has been extensively discussed in the literature (e.g. Lambeck 1996; Peltier *et al.*, 2002). In essence, models of GIA consist of three key elements: An ice loading model (to define the global distribution of grounded ice thickness over time), an Earth model (to simulate deformation of the solid Earth to surface loading) and an algorithm to compute associated changes in sea level (which, as well as being a key observable, is also an important component of the GIA loading model). The resultant change in RSL with respect to the present at a location φ and time t is summarised in equation 5.1:

$$\Delta\zeta(\varphi, t) = \Delta\zeta_e(t) + \Delta\zeta_r(\varphi, t) + \Delta\zeta_i(\varphi, t) + \Delta\zeta_w(\varphi, t)$$

(Eq. 5.1)

after Lambeck (1996)

Where: $\Delta\zeta_e(t)$ *is defined as the eustatic change in sea level at time t with respect to the present (defined as the change in ocean volume divided by ocean area);* $\Delta\zeta_r(\varphi, t)$ *describes the change in sea level that would occur on a rigid Earth, resulting from the gravitational attraction of the ice and water;* $\Delta\zeta_i(\varphi, t)$ *is the glacio-isostatic term reflecting modification of the Earth's surface under the changing ice load;* $\Delta\zeta_w(\varphi, t)$ *describes the sea-level change resulting from the hydro-isostatic influence of changing water load.*

In the past, GIA-related data have been employed in conjunction with an adopted model of the Late Pleistocene ice-sheet histories to help infer Earth viscosity structure (e.g. Peltier and Andrews 1976; Sabadini *et al.*, 1982; Wu and Peltier 1983; Nakada and Lambeck 1989; Ivins *et al.*, 1993; Mitrovica *et al.*, 1994; Kaufmann and Wolf 1996; Mitrovica and Forte 1997). The inverse of the above procedure involves adopting a realistic Earth model (which has been inferred from seismic and other GIA related data) to elucidate patterns of crustal loading caused by ice sheet growth. It is this *modus operandi* which is of interest to this study and follows many previous modelling attempts for elsewhere in formerly glaciated regions (e.g. Tushingham and Peltier 1991; Lambeck 1993a,b; Shennan *et al.*, 2002).

Some authors have expressed concern about the ability of geophysical models to accurately reproduce patterns of RSL change (e.g. McCabe *et al.*, 2005). It has been argued that models of this type are over-reliant on Holocene sea level observations to constrain predictions of RSL change during the earlier deglacial phase (e.g. McCabe, 1997), whilst others have suggested that such models are unable to capture detailed variations in RSL change during the Holocene period (e.g. Smith, 2005). However, such criticisms are often based upon controvertible indicators of former sea level which as yet cannot provide unequivocal data with which to test these suggestions.

Within this chapter, the validity of the Irish component of BIM- 1 is tested (then modified). The key components of this model are outlined below.

5.2.1 Earth Model

The Earth model employed in this investigation is a spherical, self-gravitating, compressible, Maxwell visco-elastic body. It was initially developed by Tushingham and Peltier (1991) and has formed the basis of several subsequent geophysical modelling studies (e.g. Gehrels *et al.*, 2004; Milne *et al.*, 2002, 1999; Shennan *et al.*, 2006a). The elastic and density structure are taken from seismic constraints (Dziewonski and Anderson 1981) and are depth-parameterised with a resolution of between 10 and 25 km. The viscosity structure is more crudely parameterised into three layers which correspond to the lithosphere (in which the viscosity is set to a very high value of 1×10^{43} Pa s), the upper mantle (from the base of the lithosphere to the 670 km seismic discontinuity) and the lower mantle (from 670 km depth to the core-mantle boundary). In this study, the same viscosity parameters that have resolved discrepancies between far-field RSL data and model predictions (Bassett *et al.*, 2005) and delivered a good fit to the observational evidence of postglacial RSL change in Britain (Shennan *et al.*, 2006a) are employed. These values are 1) a lithospheric thickness of 71km, 2) an upper mantle viscosity (v_{UM}) of 5×10^{20} Pa s, 3) a lower mantle viscosity (v_{LM}) of 4×10^{22} Pa s.

In previous GRM investigations from the British Isles region, values for both upper and lower mantle viscosity have remained broadly consistent between studies. However, significant differences have emerged in the depth of the adopted lithosphere with values ranging from as low as 65 km (e.g. Lambeck, 1995, 1996; Shennan *et al.*, 2000a) up to 120 km (e.g. Peltier, 1986). To a certain extent, such discrepancies have arisen because of the way glacial rebound models essentially act as an interpolator between field observations: The interpolation parameters are specified as mantle-response functions and an ice height scale factor, and since both are often relative 'unknowns', a degree of trade off between Earth and ice parameters becomes possible (Lambeck and Purcell, 2001). Until the recent work of Shennan *et al.* (2006a), the best estimate for (regional) lithospheric thickness came from the work of Peltier *et al.* (2002); Shennan *et al.* (2002); Peltier (2004). The authors (*idem*) used both RSL and glaciological evidence from Britain to infer a lithospheric thickness of c.90 km which was also in good agreement with the independent seismic analyses of DeLaughter *et al.* (1999). However, Shennan *et al.* (2006a) have shown that the failure to account for the affects of terrain and the nature of ice build up prior to the LGM invalidates the inference of lithospheric thickness made by Peltier and co worker (see below for more details).

Several authors have considered whether the three layer Earth model used in this investigation accurately depicts the Earth's internal structure and have instead suggested that more subtle variations exist which can only be adequately depicted in a multi layered model. However, the work of Lambeck *et al.* (1996, 1998) and Lambeck and Johnston (2000) strongly suggests that a three-layer model will suffice and that increasing the number of layers to five yields only a marginal improvement in the overall comparison of predictions of (postglacial) sea-level change with observations. Indeed, (at least in terms of the British Isles data) 'for all predictive purposes of sea-level change and shoreline evolution, three-layer or five-layer models are essentially equivalent' (Lambeck *et al.,* 1998; pp 135).

Virtually all efforts at addressing the GIA problem through geophysical modelling have focused on incorporating an Earth model that accurately depicts the radial stratification of the Earth's mantle/lithosphere. However, it has been suggested that for the Earth response, there may be no single set of rheological parameters that is appropriate globally because of the possibility that lateral variations in Earth structure exist (Lambeck *et al.*, 2002). Models employing lateral variation in lithospheric thickness and viscosity structure have sometimes been attempted, but often at the expense of dropping a number of other key parameters of the already complex radially symmetric models (Lambeck *et al.*, 1998). It should be noted that Peltier *et al.* (2002); pp 453 state that from their modelling experience, "at present... no systematic errors have yet been identified that argue unambiguously for the importance of lateral viscosity heterogeneity of the mantle in properly reconciling the totality of the available data related to the process of GIA." Such lateral variation probably does exist however, but the assumption of radially symmetry is operationally adequate in view of the present observational uncertainties for both sea level and the ice loads (Lambeck, 2004)

In previous GRM exercises, there has been a degree of debate over whether to operate in the calibrated, 'Newtonian' timescale or to adopt radiocarbon ages (e.g. Lambeck 1993a,b, 1995, 1998; Lambeck *et al.*, 1996; Peltier *et al.*, 2002; Peltier, 1991). Lambeck, (1998) notes that time enters into the formulation of the rebound models in three ways: through dated constraints on the geographical extent of former ice cover, through dated sea-level indicators defining the position of

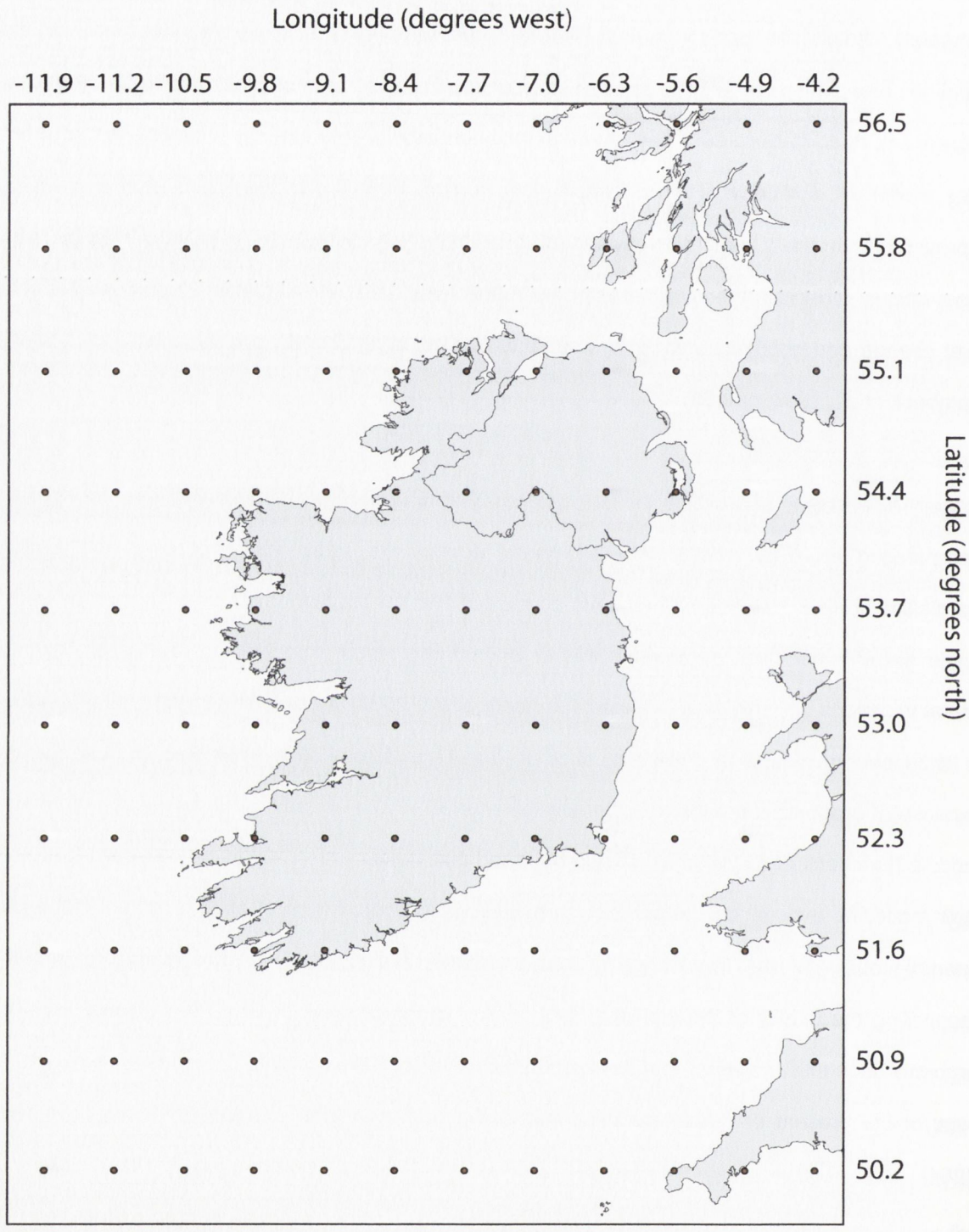


Figure 5.1 Grid cell resolution for the Ice model component of BIM- 1

palaeo sea level and finally, through the mantle rheology. Because of the non linearity associated with the radiocarbon timescale, it has sometimes been argued that inferences about ice sheets drawn from rebound models adopting the radiocarbon timescale are either incorrect or at best inexact (e.g. Bard *et al.*, 1990b; Peltier 1991). Many previous modelling efforts have not been consistent in their usage of timescales (e.g. Shennan *et al.*, 2000a; Lambeck 1993a,b, 1995, 1996, 1998): The effects of this have been concisely outlined by Peltier *et al.* (2002), who show that errors become magnified the further back in time RSL predictions are made. Because of this, the Newtonian timescale has been employed here.

5.2.2 Sea-Level Model

To generate sea-level predictions based on an input Earth and ice model, the most recent, generalised, form of the sea-level equation is solved (Mitrovica and Milne 2003; Kendall *et al.*, 2005). By applying this most recent version of the theory, the influence of shoreline migration and sea-level changes in areas of ablating marine-based ice are accounted for. This theory also takes into account the influence of GIA-induced perturbations in Earth rotation on sea-level change (e.g. Milne and Mitrovica 1996). All predictions were generated using a spherical harmonic truncation of degree and order 256.

5.2.3 Ice Model

The adopted ice model is comprised of a relatively low (spatial) resolution global model with a higher resolution (0.7 x 0.7 degree) local model for the BIIS (Figure 5.1). The global component of the ice model is the same as that used by Shennan *et al.* (2006a) and follows the analyses of Bassett *et al.* (2005) which gives a close fit with far-field observations of RSL dating from the time of the LGM to c. 9000 BP (Figure 5.2). These far field RSL records are derived from both fossil coral data (Barbados, Tahiti and Huon) and dated organic material (Bonaparte Gulf and Sunda Shelf) and can be combined to form a good approximation for the 'global' eustatic signal between the LGM and the start of the Holocene.

However, both the precision and accuracy of the coral datasets and the dated sediment core records have been questioned. For example, the Sunda Shelf evidence presented by Yokoyama *et al.* (2000) has been challenged and inconsistencies in their interpretation of the various sediment

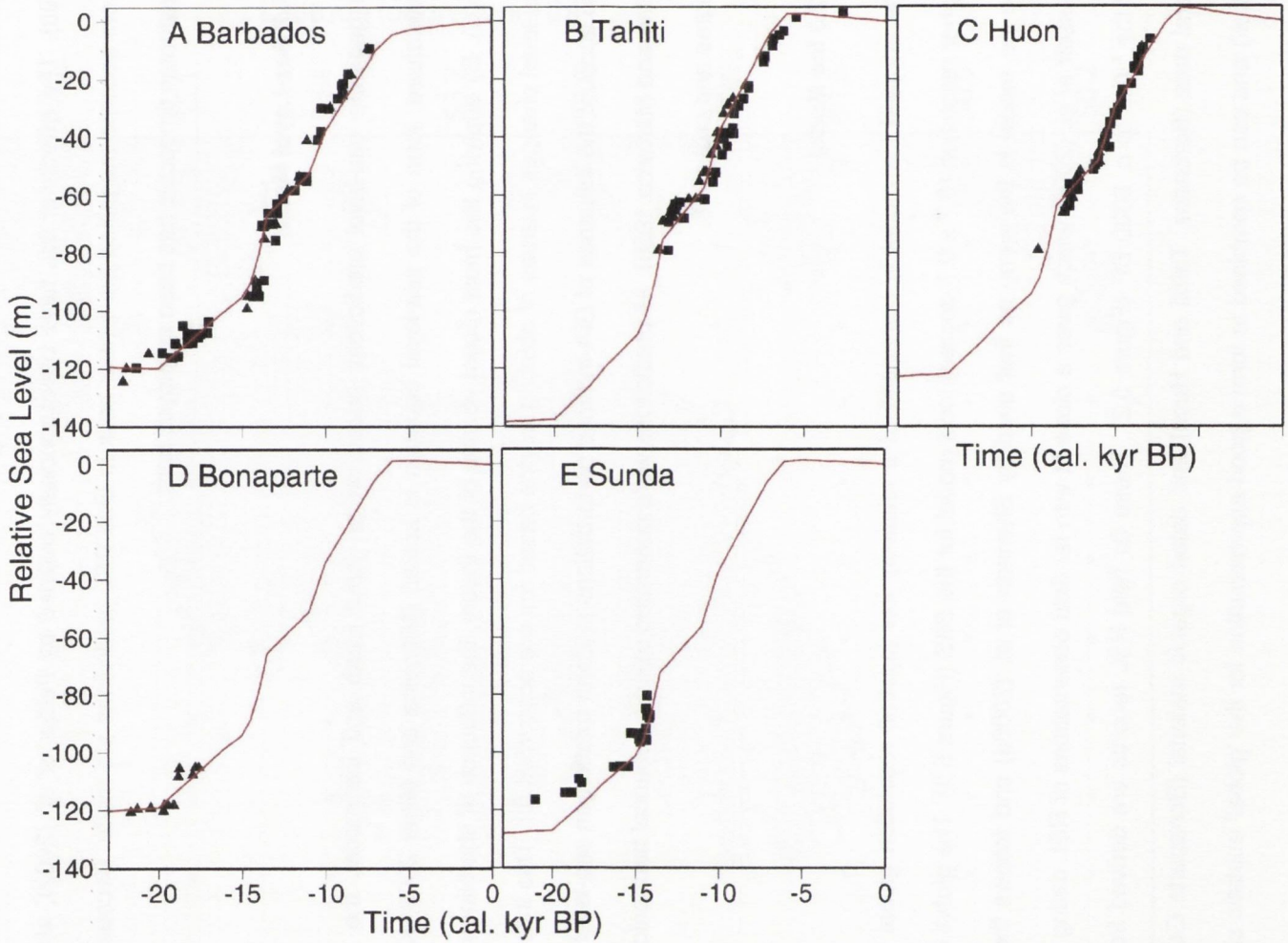


Figure 5.2 Observed fit between RSL predictions derived from the best fit global ice model of Bassett *et al.* (2005) and far field RSL records from (A) Barbados (Fairbanks, 1989; Bard *et al.*, 1990); (B) Tahiti (Bard *et al.*, 1996); (C) Huon (Chappell and Polach, 1991; Cutler *et al.*, 2003); (D) Bonapart Gulf (Yokoyama *et al.*, 2000) and (E) Sunda Shelf (Hanebuth *et al.*, 2000). Triangles and squares represent U-Th and calibrated ^{14}C dated samples, respectively. (Adapted from Bassett *et al.*, 2005)

cores that make up this record have been highlighted (Shennan and Milne, 2003). At Barbados differences of opinion exist over an appropriate correction term for regional tectonic uplift (e.g. Radtke *et al.*, 1988; Peltier and Fairbanks, 2006) and attention has also been drawn to the fact that the dated coral samples are not taken from a single coral bank but from several closely spaced cores through different coral colonies (e.g. Lambeck *et al.*, 2002). Furthermore, sea-level reconstructions based upon dated coral samples are invariably of low resolution and even *Acropora palmata* (the coral species with the narrowest vertical distribution) has an indicative range of 5m (Lighty *et al.*, 1982).

Despite the above problems, these records have successfully been used to detect marked changes in the rate of meltwater discharge since the LGM. Of particular interest is the discovery of Meltwater Pulse 1A (MWP 1A) in the Barbados sea-level record, a rise in sea level of c.24 m in less than 1000 years between 14 - 15 000 BP (Fairbanks, 1989). Whilst a degree of debate exists over the hemispheric origins of this meltwater event (e.g. Clark *et al.*, 1996, 2002; Peltier, 2005, Tarasov and Peltier, 2005), its detection in other far field records from Tahiti (Bard *et al.*, 1996) and the Sunda Shelf (Hanebuth *et al.*, 2000) mean its existence is now accepted *nemine contracidente*, and is a proven feature of postglacial eustatic sea-level history (Peltier, 2002). Furthermore several recent GRM studies have shown its inclusion to be vitally important in reconciling model predictions with observations in certain near field localities (e.g. Shennan *et al.*, 2002). Significantly, the rebound model of Lambeck (1993a,b) (which formed the basis for the later GRM study for Ireland) does not incorporate this pronounced meltwater event at 14 - 15 000 BP and instead, compensates by adding increased meltwater input during the Holocene period (see discussion below). Some workers have also argued for the existence of an additional meltwater pulse at c.19 000 BP (e.g. Yokoyama *et al.*, 2000; Clark *et al.*, 2004). However, its existence remains highly controvertible (see Peltier and Fairbanks, 2006) and as such, it has not been incorporated in the global ice model employed in this study.

Whilst the eustatic model of Bassett *et al.* (2005) gives an excellent fit with far field observations of RSL in the period prior to c.9 000 BP, the Holocene component of this model remains to be calibrated and so is not used to constrain this part of the postglacial record. In fact, the nature of eustatic sea-level rise throughout the Holocene has elicited much attention with uncertainty over

the time at which ice melting ended during the late Holocene (Table 5.1). In many of the original modelling studies, it was assumed that melting ceased abruptly at c. 7000 BP with no further eustatic contribution up to the present. However, this assumption was later questioned (e.g.

Author(s)	RSL data source	Time interval (cal. kyr BP)	Eustatic contribution (m)	Suggested source
Nakada and Lambeck (1989)	Australia/ Pacific Islands	6 - present	1.5 – 2	Antarctic
Flemming *et al.* (1998)	Global	7 - 1	3 – 5	Antarctic/ temperate glaciers
Lambeck & Bard (2000)	French Mediterranean coast	7 - 2	3	
Lambeck (2002)	Australia	7 - 2	3	
Peltier (2002)	Caribbean	7 – 4	2	
Milne *et al.* (2005)	Caribbean/ South America	7 - present	1	

Table 5.1 Estimates of Late Holocene eustatic sea-level rise and suggested source region(s)

Nakada and Lambeck, 1988; Lambeck and Nakada, 1990) and more recent studies suggested eustatic sea level over this period increased at an average rate of about 0.5 mm yr^{-1}, for a total of about 3 m (Lambeck *et al.*, 1998; Lambeck, 2002; Flemming *et al.*, 1998). Such figures remain controvertible however: Peltier *et al.* (2002) and Shennan *et al.* (2002) argue that a late Holocene melting 'tail' of the strength suggested by Lambeck and others is untenable with far-field observations of the mid-Holocene highstand and go on to suggest that melting had largely ceased by 4000 BP. In the analyses in this chapter, the findings of Nakada and Lambeck (1989) are adopted who suggest a late Holocene eustatic meltwater contribution of c. 2m between 6000 and 2000 BP. Far-field corrections to the Bassett *et al.* (2005) eustatic model in the period 7000 - 3000 BP are made as follows: -1 m at 7000 BP, -2 m at 6000 BP, reducing to -0.2 m at 3000 BP. (These corrections are identical to those utilized by Shennan *et al.* (2006a), facilitating comparison between findings from the two studies).

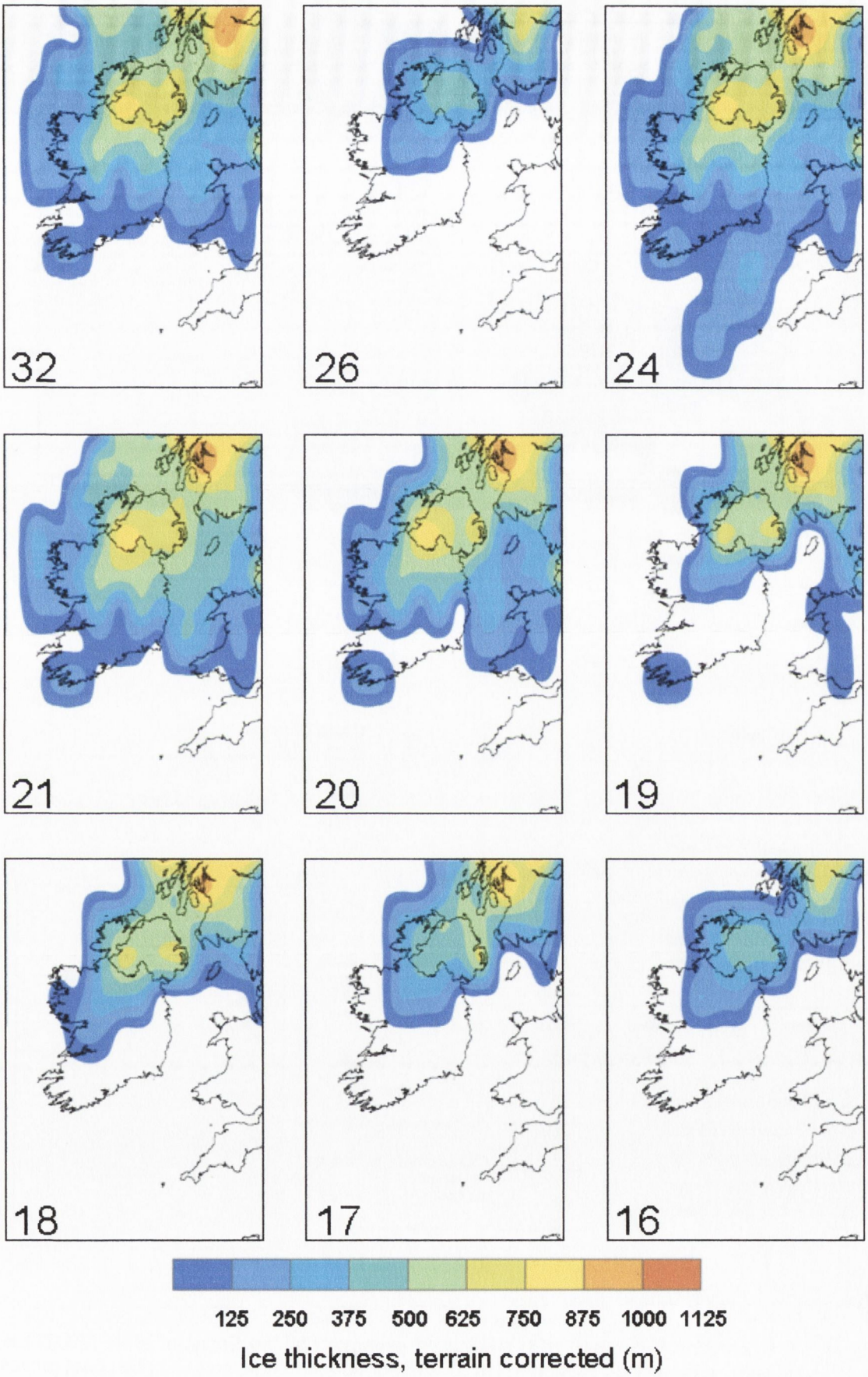


Figure 5.3 Time-space evolution of the Irish component of BIM- 1 (32 – 16 000 cal. BP)

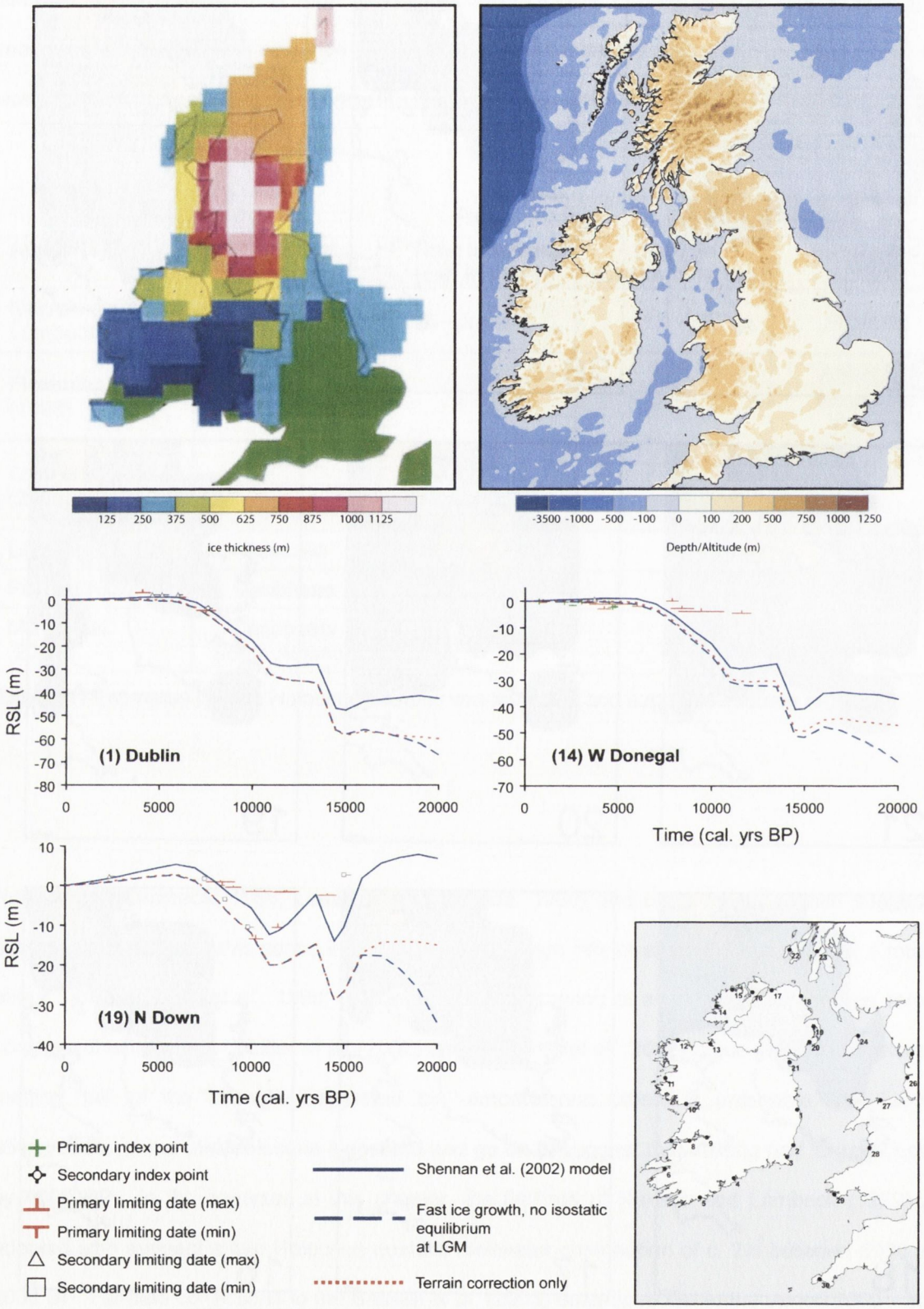


Figure 5.4 A comparison between RSL predictions generated by the Shennan *et al.* (2002) ice model (solid line) and a revised version which a) is out of equilibrium at the LGM (ice sheet growth after 34 000 cal. BP) (long dashed line) and b) is corrected for underlying terrain (dotted line). Also shown is the LGM ice thickness/extent of the Shennan *et al.* (2002) ice model (top left) along with a topographic/bathymetric map of the British Isles (top right)

In order to accurately model GIA observations from Ireland and adjacent regions, a high-resolution spatial and temporal description of the local to regional ice load leading up to and following the LGM is required (Lambeck, 1993a). The space-time evolution of the Irish component of BIM-1 is illustrated in Figure 5.3 (and outlined in more detail in section 5.5.1). As Shennan *et al.* (2006a); pp 590 note: 'Important differences of opinion exist on whether it is acceptable to assume that the ice sheets, oceans and solid Earth were in isostatic equilibrium at the Last Glacial Maximum. For example, Peltier (1998) argues that they were, while Lambeck and co-workers (2002) argue that isostatic equilibrium was not achieved and therefore pre-Last Glacial Maximum changes in ice sheet dimensions are important.' The significance of taking into account pre LGM ice build up in GRM models is demonstrated in Figure 5.4.

In the BIIS model employed in this investigation, overall advance and retreat phases are based on the findings of Sejrup *et al.* (1994, 2005); Bos *et al.* (2004) and Evans *et al.* (2005). Prior to the LGM, extensive ice accumulation began at the end of the Alesund Interstadial (c.34 000 BP) leading to coalescence of the British and Fennoscandian ice sheets. (Adopting this pre LGM ice sheet configuration causes the Earth to be out of equilibrium at the LGM Figure 5.4). This ice advance episode was interrupted by retreat between 26 - 25 000 BP before rapid expansion of the BIIS at the LGM. LGM ice thickness over northern Britain is constrained by Ballantyne *et al.* (1998a) whose trimline evidence suggests a maximum of c.1000 m of ice existed over Scotland. Deglaciation began at c.21 000 BP and the removal of ice from both Ireland and Britain was largely complete by 15 000 BP (e.g. Ballantyne, 1997).

Lastly, the recent analyses of Milne *et al.* (2006) and Shennan *et al.* (2006a) have emphasized the importance of correcting for the effect that underlying topography has on overall ice sheet mass. Prior to these studies, all GRM investigations from the British Isles assumed the ice lay on a flat terrain thus (in places) greatly overestimating total ice volume. The affect of accounting for terrain is demonstrated in Figure 5.4. (The combined influence of accounting for both terrain and pre LGM ice build up may be seen in Figure 5.5).

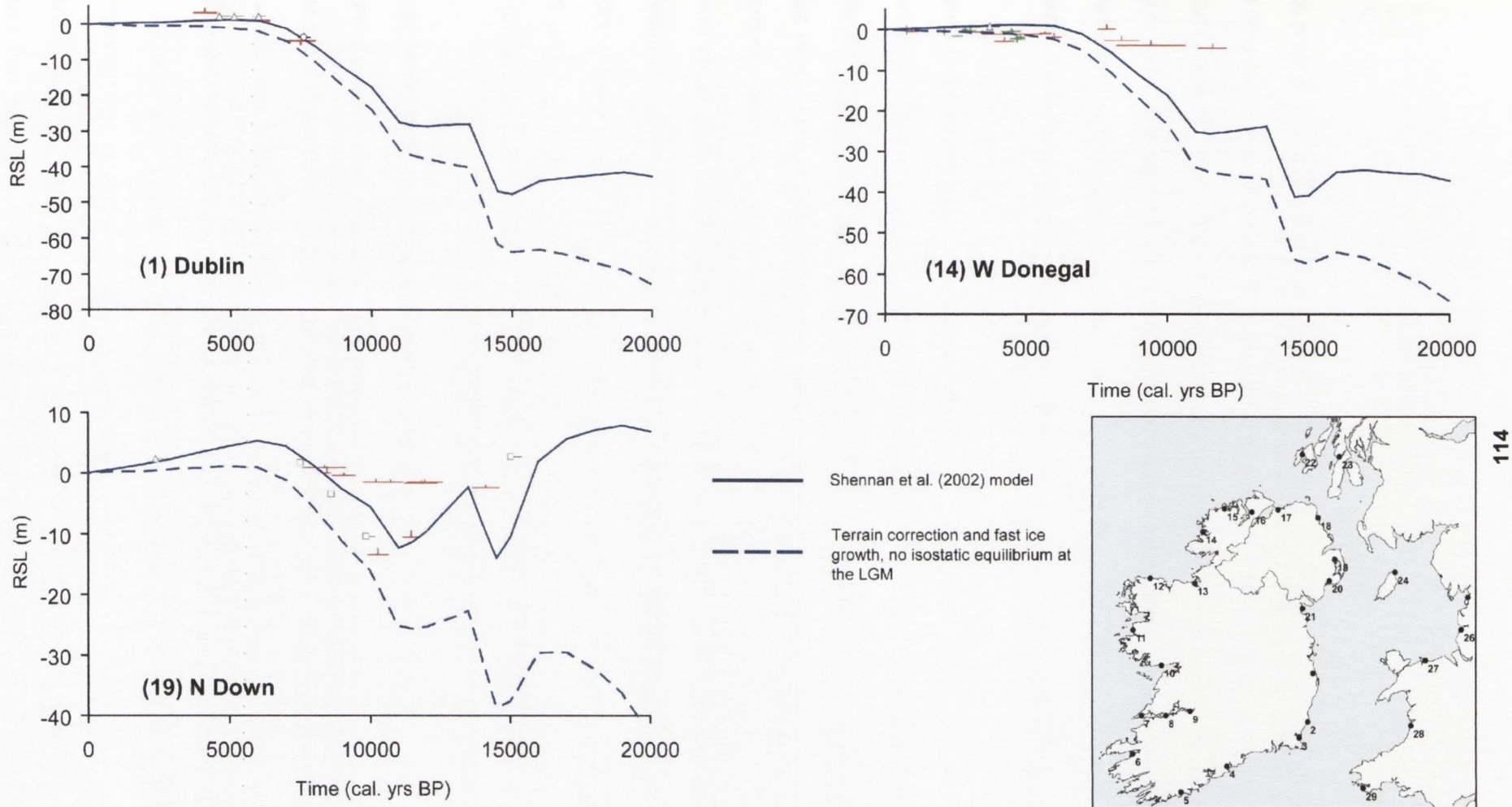


Figure 5.5 A comparison between RSL predictions generated by the original Shennan *et al.* (2002) ice model (solid line) and a revised version which accounts for both terrain *and* the effects off ice sheet growth after 34 000 cal. yrs BP (long dashed line)

Site	Name	Lat.	Long.	No. data	Reference(s)
01	Dublin	53.138	-6.123	0 + 7	Mitchell (1976)
02	North Wexford	52.521	-6.232	1	Sinnott (1999)
03	South Wexford	52.315	-6.412	11 + 5	Sinnott (1999); Mitchell (1977); Maloney (1985); Carter & Orford (1982); Orford & Carter (1982b); Dresser (1980)
04	East Cork	51.949	-7.935	10 + 1	Sinnott (1999); Carter *et al.* (1989); Mitchell (1977); Mitchell (1976)
05	West Cork	51.607	-8.906	13 + 5	Sinnott (1999); Devoy *et al.* (in press); Buzer (1980); Stillman (1968)
06	Kerry	52.107	-9.916	1 + 10	Carter *et al.* (1989)
07	South Clare	52.601	-9.721	0 + 5	O'Sullivan (2001); Devoy *et al.* (1996); Pearson (1979)
08	Mid Shannon	52.601	-9.213	0 + 3	O'Sullivan (2001)
09	Inner Shannon	52.658	-8.705	0 + 9	O'Sullivan (2001)
10	Galway	53.238	-9.305	0 + 1	Mitchell (1976)
11	Connemara	53.679	-9.900	0 + 10	Devoy *et al.* (1996); Jessen (1949)
12	North Mayo	54.320	-9.540	0 + 8	McCabe *et al.* (1986, 2005)
13	Sligo	54.258	-8.607	0 + 1	Burenhuit (1980)
14	West Donegal	54.885	-8.431	8 + 11	Harmon (2006); Shaw (1985); Shaw & Carter (1994); Gehrels (2004); Smith & Pilcher (1973); Gaulin *et al.* (1983); Pearson (1979); Telford (1978); Carter (1982)
15	North Donegal	55.160	-7.979	3 + 6	Shaw (1985); Shaw & Carter (1994)
16	Lough Swilly	55.119	-7.399	0 + 8	Shaw & Carter (1994); Colhoun *et al.* (1973); McCabe & Clark (2003)
17	Derry	55.15	-6.851	0 + 10	Woodman (unpub.*); Bazley *et al.* (unpub.*); Carter & Wilson (1990); Pearson (1979); Battarbee I. (1982)
18	North Antrim	55.052	-6.014	0 + 8	Mitchell & Stephens (1974); Prior I. (1981); Smith *et al.* (1974)
19	North Down	54.55	-5.658	0 + 15	Stephens & Collins (1960); Smith *et al.* (1974); Morrison & Stephens (1961); Morrison & Stephens (1965); Manning *et al.* (1970); Pearson & Pilcher (1975); Harkness & Wilson (1979); McCabe & Clark (1998)
20	South Down	54.29	-5.774	2 + 10	Singh & Smith (1973); Carter (1982); Hubbs, Bien & Suess (1965); Dresser *et al.* (1973); McCabe & Clark (1998); Harmon (2006)
21	Dundalk	53.947	-6.340	0 + 16	Mitchell (1971); Penney (1983); Jessen (1949); McCabe *et al.* (2005); McCabe & Clark (1998); McCabe & Haynes (1996); Clark *et al.* (submitted+)
22	Islay	55.808	-6.343	8	Shennan *et al.* (2006b)
23	Knapdale	55.784	-5.561	9 + 7	Shennan *et al.* (2006b)
24	Isle of Man	54.395	-4.388	12 + 2	Tooley (1977); Roberts *et al.* (2006)
25	Morecombe Bay	54.085	-2.958	21 + 3	Tooley (1978a); Birks (1982); Zong & Tooley (1996)
26	Lancashire	53.685	-2.990	42	Tooley (1978a,b, 1985); Huddart (1992)
27	North Wales	53.300	-3.748	9	Tooley (1978a); Kidson & Heyworth (1982); Bedlington (1993)
28	Mid Males	52.473	-4.058	14 + 3	Heyworth & Kidson (1982)
29	SW Wales	51.658	-5.066	1	Heyworth & Kidson (1982)
30	Cornwall	50.129	-5.482	7	Healy (1995)

* cited in Carter (1982a) + cited in McCabe *et al.* (2005)

Table 5.2 Source(s) of RSL data from Ireland and UK (west coast) sites

			RSL observation			GRM (v_{UM} 5 x 10^{20} Pa s)						GRM (v_{UM} 4 x 10^{20} Pa s)		
SL database tier	Location	Radiocarbon lab code	Age (cal. yrs BP)	RSL Alt. (m)	Error (m)	BIM-1 (m)	GRM A (m)	GRM B (m)	GRM C (m)	GRM D (m)	GRM E (m)	BIM-1 (m)	GRM E (m)	Reference
SLIP (I)	(4) E Cork	Q-2379	9205±199	-17.41	1.05	-23.0	-23.5	-22.9	-23.8	-23.8	-22.0	-23.4	-22.7	Carter *et al.* (1989)
SLIP (I)	(14) W Donegal	SRR-2606	4703±255	-2.2	0.82	+1.50	+0.36	+1.19	+0.18	+0.39	+0.79	+0.42	-0.23	Shaw and Carter (1994)
SLIP (I)	(14) W Donegal	Beta-195704	4807±154	-1.96	0.5	+1.52	+0.32	+1.20	+0.16	+0.37	+0.79	+0.41	-0.25	Harmon (2006)
SLIP (I)	(15) N Donegal	Beta-42528	4122±284	-0.94	0.78	+2.36	+1.38	+2.08	+1.26	+1.47	+1.76	+1.27	+0.69	Shaw (1985)
SLIP (I)	(15) N Donegal	Beta-34314	5072±240	-1.36	0.82	+2.87	+1.52	+2.48	+1.36	+1.61	+1.99	+1.44	+0.66	Shaw and Carter (1994)
SLIP (I)	(24) Isle of Man	AA29333	5382±104	+2.09	0.5	+4.82	+3.80	+4.86	+3.69	+3.84	+4.23	+3.26	+2.78	Roberts *et al.* (2006)
SLIP (I)	(24) Isle of Man	AA29329	6392±99	+2.11	0.5	+4.78	+3.56	+4.87	+3.37	+3.55	+4.07	+3.06	+2.34	Roberts *et al.* (2006)
Lim. Date (I)	(14) W Donegal	UB-2676	5697±205	-0.95	0.34	+1.53	0	+1.13	-0.22	+0.06	+0.58	+0.17	-0.70	Shaw (1985)
Lim. Date (I)	(15) N Donegal	Gif-5482	6580±301	-1.125	1.00	+2.33	+0.32	+1.77	+0.10	+0.45	+1.00	+0.47	-0.75	Shaw (1985)
Lim. Date (I)	(16) L Swilly	Gif-5483	6625±314	-1.925	2.00	+4.66	+2.55	+4.21	+2.35	+2.68	+3.20	+2.49	+1.15	Gaulin (1983)
Lim. Date (I)	(17) Derry	Beta-34315	10026±346	-6.8	0.34	-2.47	-7.16	-3.27	-7.47	-6.86	-5.97	-5.59	-9.03	Carter and Wilson (1990)
Lim. Date (I)	(18) N Antrim	UB-984	9022±381	-1.54	0.34	+2.48	-1.07	+2.09	-1.30	-0.86	-0.17	-0.83	-3.37	Prior *et al.* (1981)
Lim. Date (I)	(18) N Antrim	UB-985	13600±334	-2.24	0.34	+3.76	-6.22	+2.96	-6.43	-5.43	-4.12	+3.18	-5.61	Prior *et al.* (1981)
Lim. Date (I)	(19) N Down	St-3066	10283±366	-13.5	1.00	-4.98	-9.67	-5.13	-10.1	-9.55	-8.17	-7.95	-11.19	Manning *et al.* (1970)

Table 5.3 Empirical RSL observations from Ireland/west Britain alongside glacial rebound model predictions of RSL. All Limiting date (type I) observations represent maximum altitudes for RSL. See text for glacial rebound model descriptions.

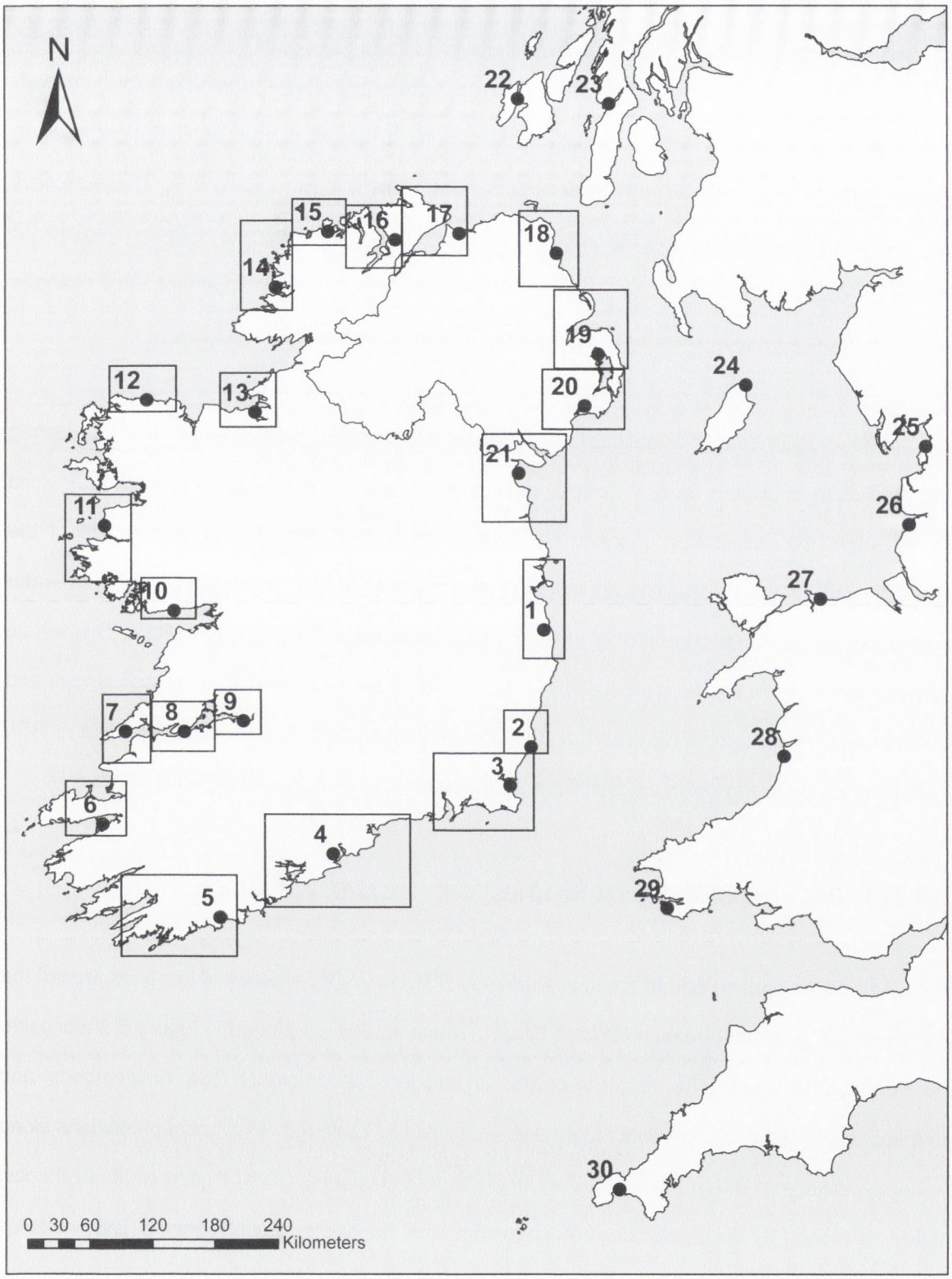


Figure 5.6 Location map of the 30 RSL time-series records

5.3 Reconstructing Relative Sea-Level Change

Available RSL data from Ireland has been screened, classified and compiled into a database (Chapter 4). At present, lengthy suites of index points from single sites in Ireland are rare: Instead, SLIPs are grouped into 21 distinct geographical regions to permit more meaningful comparison between RSL observations and GRM predictions (see Table 5.2 and Figure 5.6). These records from Ireland are presented alongside available RSL data from the west coast of Britain (regions 22 – 30) (Appendix VI). Their inclusion in any GRM analyses from Ireland is important since changes to the evolution and mass of the Irish ice sheet will impart a regional GIA signature.

Both the index point data and limiting data contained with the RSL database are useful for testing RSL predictions produced by GIA models, since sea-level curves should pass through SLIPs and remain within the constraints imposed by limiting dates. Post-depositional compaction of the sediment column, either under its own weight or as a consequence of subsequent loading by water or an overlying sediment burden may lower the reconstructed altitude of former RSL. This will be greatest where index points are established from thick intercalated sedimentary sequences and consequently, when comparing predictions with observations from such sequences, best fit RSL predictions should lay towards the top of the SLIP scatter (e.g. Shennan *et al.*, 2002, 2006a).

5.4 Testing the BIM-1 GIA Model for the British Isles

The BIM-1 GIA model is used to produce estimated RSL curves for a series of locations around the Irish coast and adjacent areas in western Britain. These curves are plotted in Figure 5.7 alongside the geologically-based RSL reconstructions. A selection of empirical RSL observations and corresponding BIM-1 RSL predictions are also presented in Table 5.3. The model predictions along Ireland's east coast (Dublin (1) and S. Wexford (2)), south coast (E. Cork (4)) and south west coast (Inner Shannon (9)) are in reasonable agreement with the observational dataset. This indicates that, away from the former centre of loading by the Irish ice sheet, BIM-1 is effective in reproducing the dominant patterns of change. Further to the north, however, significant discrepancies begin to emerge between the model predictions and the empirical RSL data.

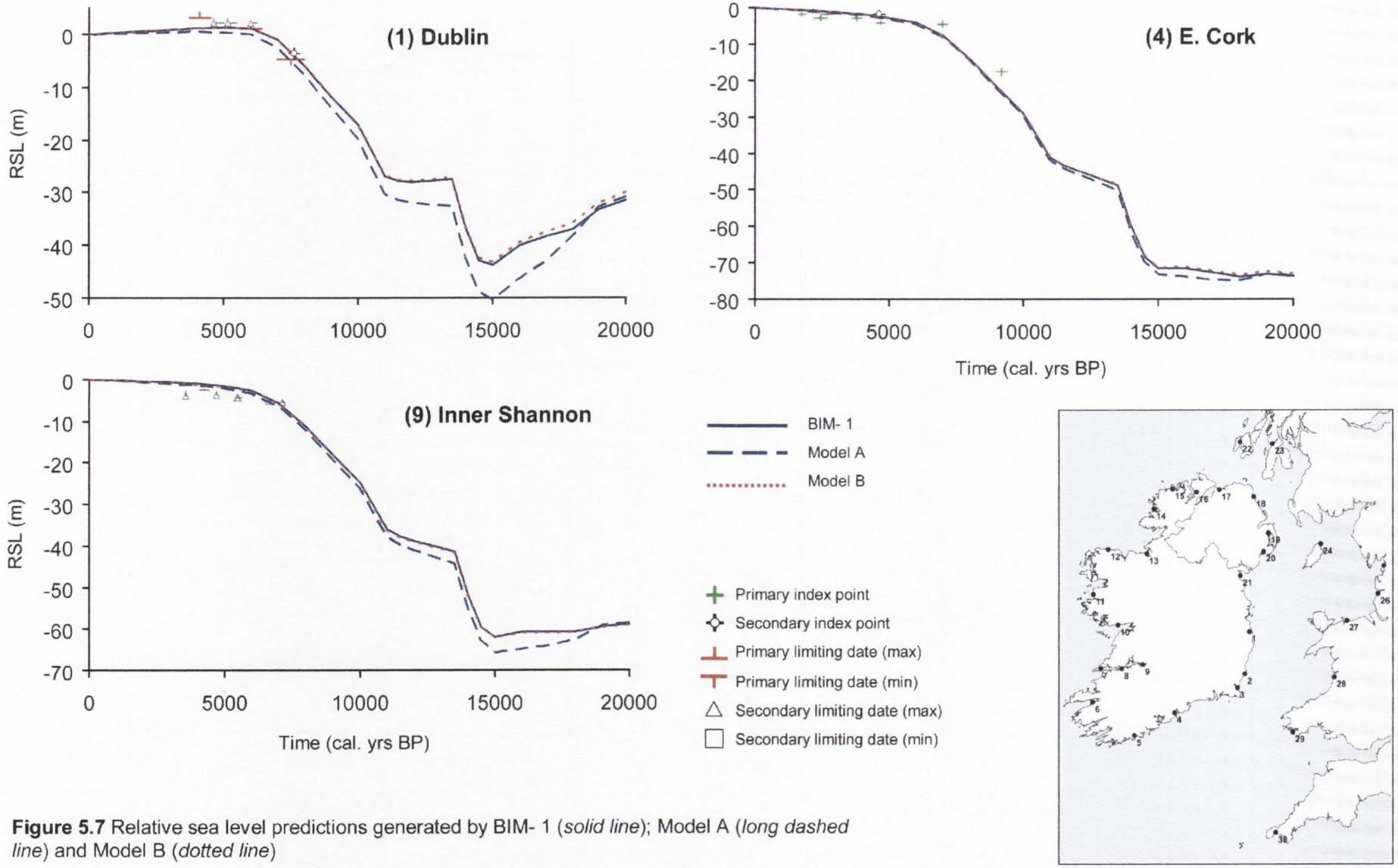


Figure 5.7 Relative sea level predictions generated by BIM- 1 (*solid line*); Model A (*long dashed line*) and Model B (*dotted line*)

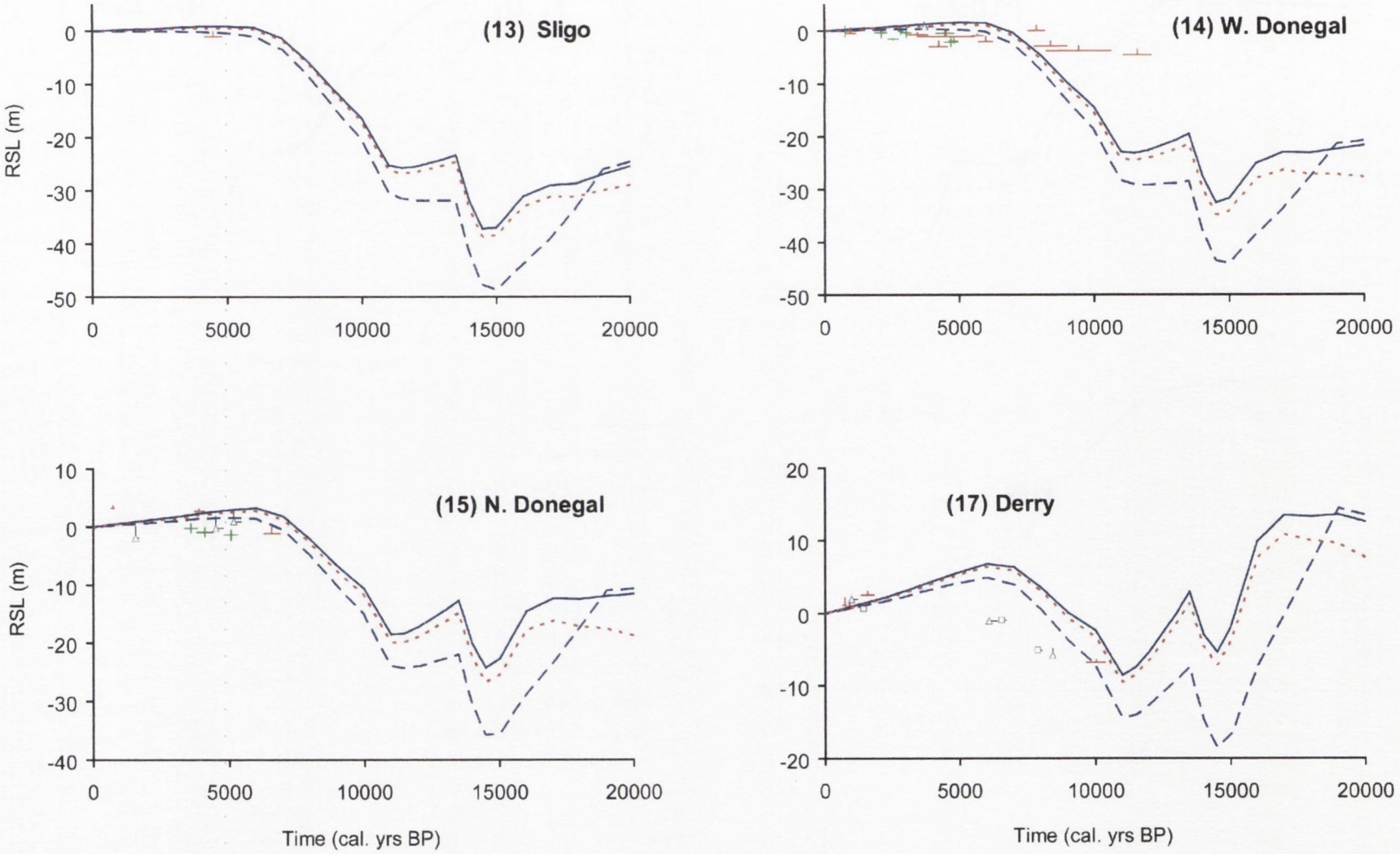


Figure 5.7(cont) Relative sea level predictions generated by BIM- 1 (*solid line*); Model A (*long dashed line*) and Model B (*dotted line*)

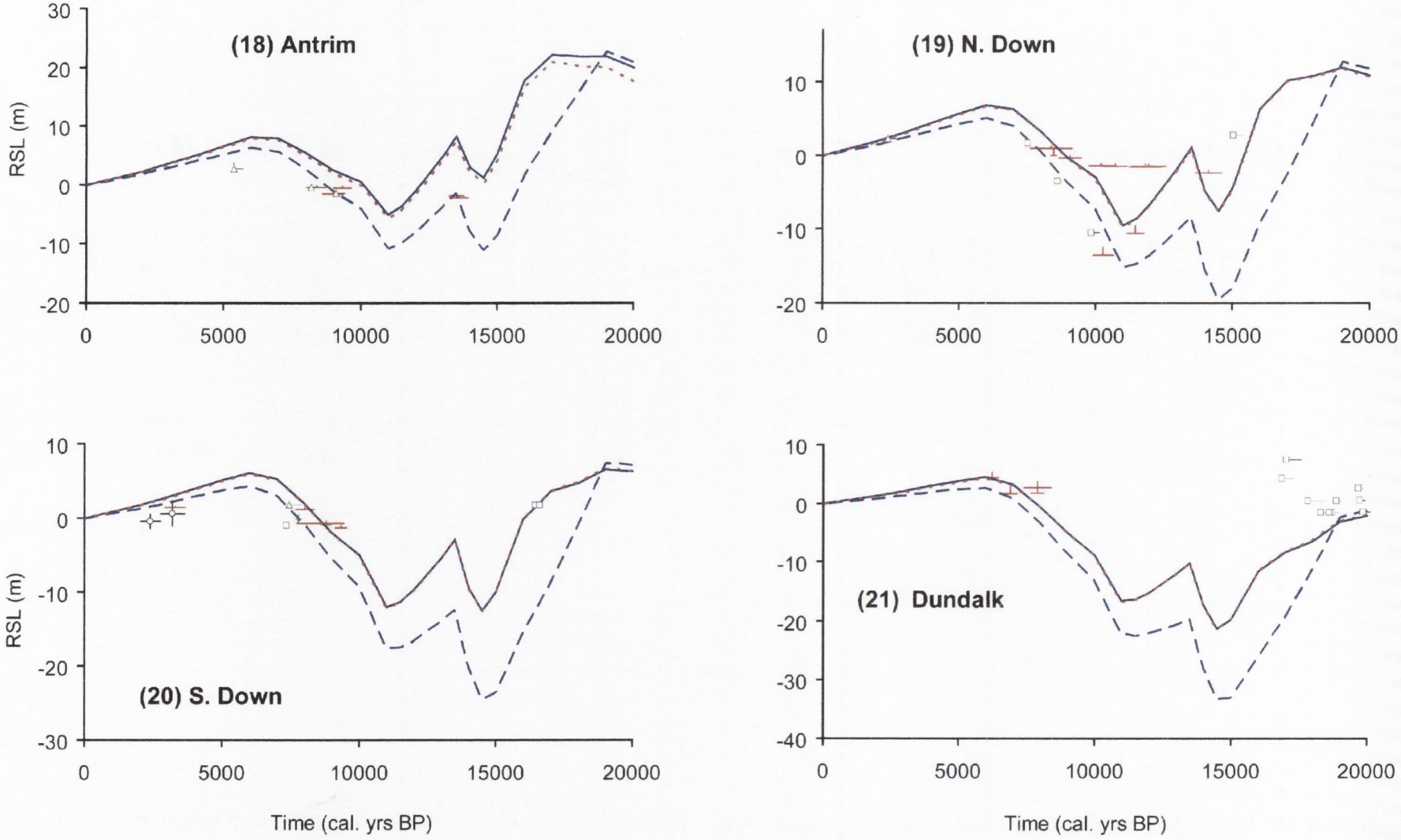


Figure 5.7(cont) Relative sea level predictions generated by BIM- 1 (*solid line*); Model A (*long dashed line*) and Model B (*dotted line*)

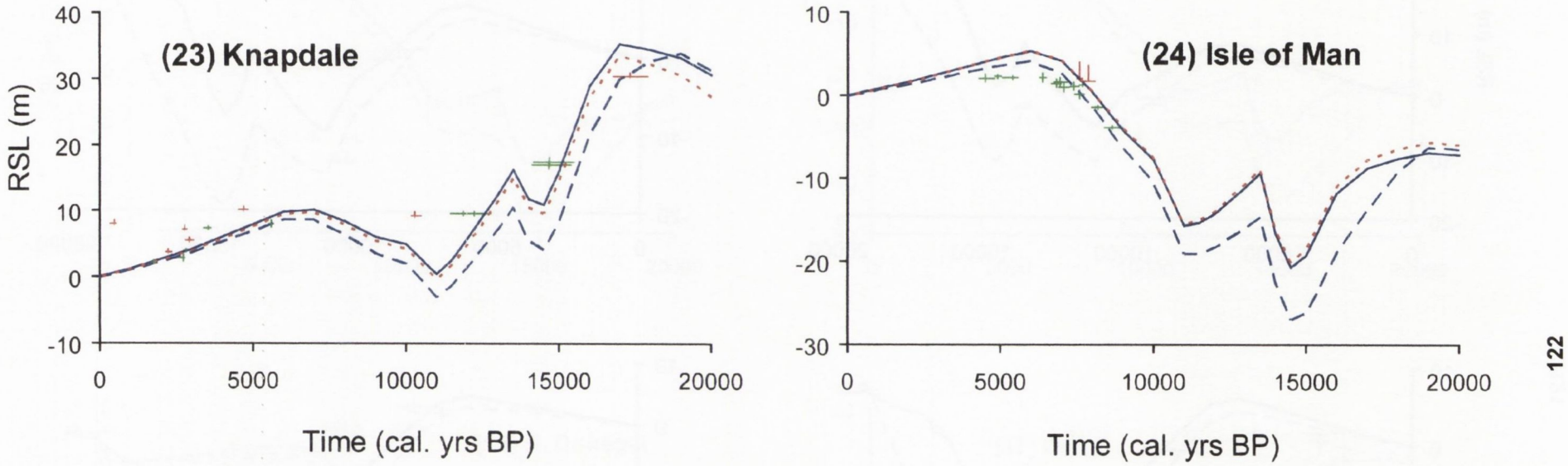


Figure 5.7(cont) Relative sea level predictions generated by BIM- 1 (*solid line*); Model A (*long dashed line*) and Model B (*dotted line*)

The first of these misfits is the tendency for BIM-1 to underestimate the maximum height of Late Devensian RSL along the north Ulster coastline, as indicated by abundant raised shoreline evidence in this region. Limitations associated with dating and quantifying the indicative meaning of raised shorelines, coupled with the lack of primary index points for this time period, prevents a precise quantification of this discrepancy. Nevertheless, these shorelines provide strong evidence of higher sea levels, both to the west and south of Antrim, than indicated by the model (Stephens and Synge, 1965; Synge and Stephens, 1966; Synge, 1977a, b; Mitchell, 1977). For example, in Derry (17) (Figure 5.7), a Lateglacial RSL of up to +25 m (compared to a modelled estimate of c. +13 m) is suggested by Stephens and Synge (1965), whilst in North Down (19), Stephens and McCabe (1977) find evidence of raised shorelines at heights of +30 m (compared to a modelled estimate of just over +10 m).

The second misfit concerns the presence of a Holocene highstand (commonly around 6000 BP), where ongoing crustal rebound starts to outpace dwindling sea-level rise. The model delivers predictions of Holocene RSL higher than present as far south and west as Sligo (13), as indicated by the 0 m isobase in Figure 5.8. In west Donegal (14), BIM-1 predicts RSL of approximately +2 m above present at 6000 BP. Whilst there are very few locations in Ireland where the magnitude of the Holocene highstand can be tightly constrained, the work of Shaw (1985), Shaw and Carter (1994), Gehrels (2004) and Harmon (2006) in Donegal has enabled the spatial signature of the 0 m Holocene isobase to be reasonably well defined. The index point data provided by the authors above shows that RSL occupied a position around -2 m at c.5000 BP in west Donegal (14) (Figure 5.7) and suggests that the 0 m threshold lies somewhere along the north Donegal coastline (i.e. further to the north and east than suggested by BIM-1). On the Isle of Man (24) (Figure 5.7), similar discrepancies exist with the Holocene RSL data presented by Roberts *et al.* (2006). Here, the model predicts values in excess of +5 m above present whereas the data constrain sea level to have been no more that 2.5 m above present during this period.

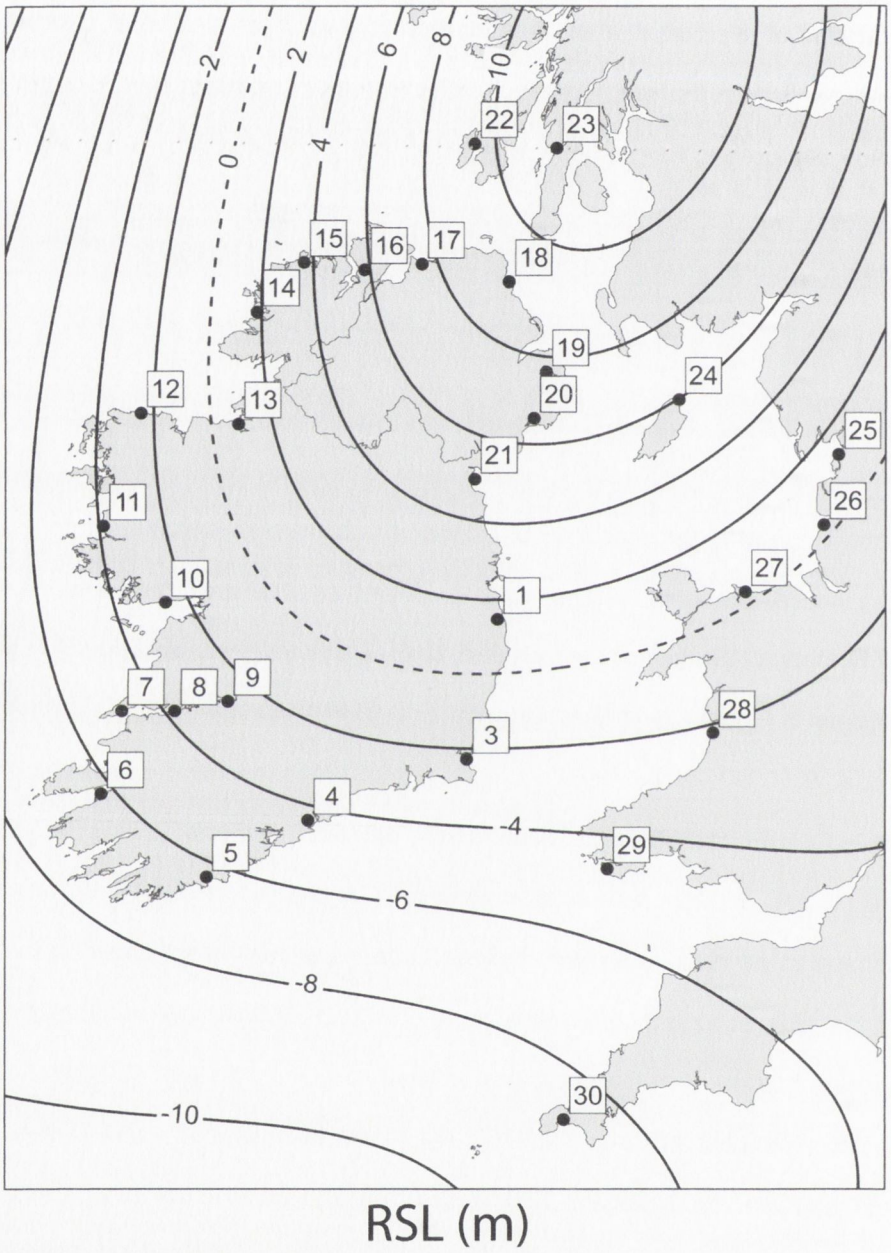


Figure 5.8 BIM- 1 RSL isobase map at 6 000 cal. BP

5.5 Examining The Possible Causes of Poor Model Performance

The RSL data from Ireland demonstrate that BIM-1 performs poorly in areas that have experienced the greatest loading from the Irish ice sheet, but that it can predict RSL change with reasonable skill in those areas that have experienced thinner, or shorter-duration ice cover. In light of this, a series of sensitivity analyses are performed to examine the Irish ice sheet component of BIM-1 and its influence on the pattern and magnitude of predicted RSL change.

5.5.1 The Role of the Irish Ice Sheet

In order to predict the timing and magnitude of glacial rebound, it is necessary to know the size (thickness and lateral extent) of the ice sheet. In addition, it is important to understand when, how fast and by how much this size has changed through time. This choice of input parameters inevitably means that there is a degree of model solution non-uniqueness due to the trade-off between variables, and this is a recognised limitation inherent in GIA modelling studies (e.g. Mitrovica and Peltier, 1995). However, both the observational RSL data and glaciological field evidence (below) provide a stringent set of independent conditions that must be met, and any adjustment to the nominal BIM-1 ice sheet model are made in light of these constraints.

5.5.1(i)a Pre LGM (>24 000 BP)

During the Derryvree Cold phase, Ireland is assumed to have been ice free, in accordance with radiocarbon dated organic silts from northwest Ireland which yield an age of 30 500 ^{14}C years BP (Colhoun *et al.*, 1972). Although no calibration curve for the time interval 26 000 – 50 000 cal BP can be recommended as yet (e.g. van der Plicht *et al.*, 2004; Bard *et al.*, 2004; Balter, 2006), an approximate sidereal age of 34 000 BP has been (tentatively) assigned to this date. The evidence presented from Derryvree is consistent with other dated faunal, floral and organic sediments from the region that evidence ice free conditions around this time (e.g. Mitchell, 1976, 1981; Stuart and van Wijngaarden-Bakker, 1985; Sejrup and Knudsen, 1993; Woodman *et al.*, 1997; Bos *et al.*, 2004; Evans *et al.*, 2005). The Derryvree cold phase was abruptly terminated by rapid ice sheet growth and expansion after c. 34 000 BP which lasted through until c. 27 000 BP. Before the onset

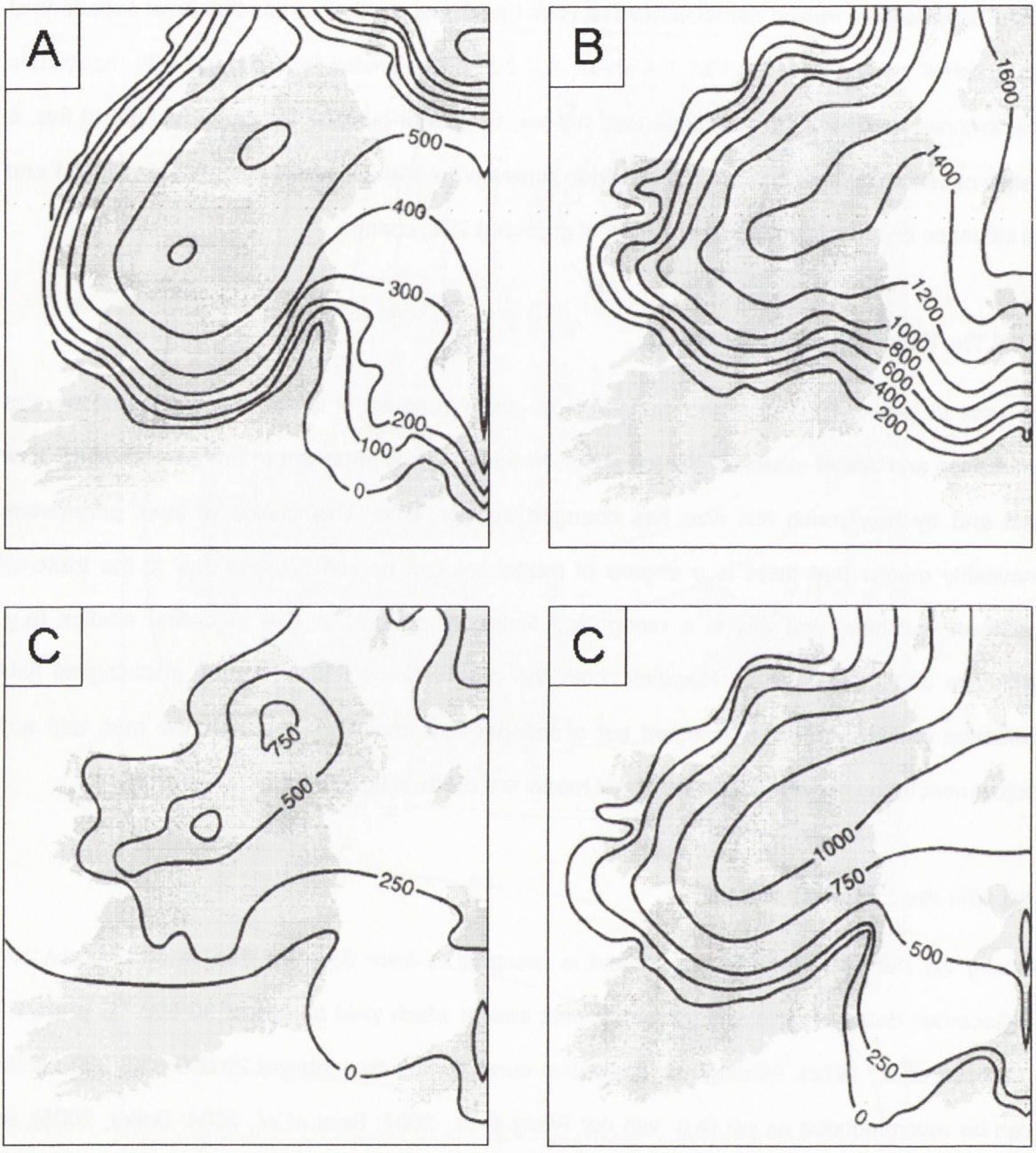


Figure 5.9 LGM ice extent and thickness from **A** Lambeck (1996); **B** Boulton *et al.* (1977); **C** Boulton *et al.* (1985); **D** Boulton *et al.* (1991)

of the LGM at 24 000 BP, a brief retreat phase ensued between 26 000 and 25 000 BP in agreement with the inferences of Sejrup *et al.* (1994).

5.5.1(i)b The LGM (24 000 – 21 000 BP)

During the LGM, a laterally extensive ice sheet is assumed to have existed over much of Ireland and out onto the adjacent shelf region (e.g. Evans *et al.*, 1980; Warren, 1992; McCabe, 1995; Clark and Meehan, 2001; Sejrup *et al.*, 2005). This contrasts with other models which envisage an ice sheet whose maximum spatial extent at the LGM to the north and west broadly followed the present day coastline (e.g. Hull, 1878, Stephens and Synge, 1965). To the south, ice is now considered to have extended beyond the traditional South of Ireland End Moraine (SIEM) limit (O'Cofaigh and Evans, 2001; Evans and O'Cofaigh, 2003; Hegarty, 2004; Ballantyne *et al.*, 2006). Recent evidence also indicates ice extending out into the Celtic Sea south of the Scilly Isles (Scourse and Furze, 2001; Hiemstra *et al.*, 2006).

When considering the vertical limits of the Irish Ice sheet, there is a relative paucity of constraints provided by trimline data within Ireland when compared to the UK (e.g. Ballantyne *et al.*, 1998a, b; McCarroll and Ballantyne, 2000). In the south west of Ireland (Macgillycuddy's Reeks), Rae *et al.* (2004) use trimline evidence to infer an ice sheet thickness of c.370 m whilst in the east (Wicklow Mountains), Ballantyne *et al.* (2006) suggest that only peaks above c.725 m remained ice free at the LGM. In Donegal, Ballantyne *et al.* (submitted) use the same technique to infer that at its maximum thickness, the ice surface in this region descended from an ice-shed slightly over 700 m above present sea level to c.500 m along the west and northwest coasts of Donegal, and to c.400 m along the northern coast of Inishowen in north Donegal. These findings support the inferences made by McCabe (1997) and Clark and Meehan (2001) for similar magnitude ice thickness in this locality and question the accuracy of the Lambeck (1996) ice model in this region (Figure 5.9a). Elsewhere, the evidence becomes more equivocal although maximal ice thickness in the Knockmealdown mountains (Lewis, 1976), Galtee mountains (McCabe, 1985) and Connemara mountains (Coudé, 1977) is unlikely to have been as great as over Ulster since the highest peaks in these mountain ranges are proposed to have been nunataks at the LGM.

The above work enables estimates for maximal LGM ice thickness to be made for a number of locations around Ireland, albeit with varying degrees of confidence. However, nearly all of these studies are geographically biased towards coastal localities, a pattern that manifests from the fact that the general relief of Ireland is essentially that of a bowl with a low lying centre and uplands around its periphery (Clark and Meehan, 2001). Accordingly, empirical constraints on ice thickness are non existent for much of central (inland) Ireland and existing estimates for these parts are instead derived from various numerically based modelling studies.

The steady state model for the BIIS developed by Boulton *et al.* (1977) represents one of the first attempts to employ mathematical analyses to estimate the vertical dimensions of the last ice sheet in this region. This model, which incorporated Synge's (1969) ice limit for Ireland allied with Weertman's (1961) climatic analysis, was used to determine the surface height profile of the ice sheet. A maximum value of 1500m was suggested for Ireland (Figure 5.9b) and these overall findings were broadly substantiated by the later work of Denton and Hughes (1981). Boulton *et al.* (1985, 1991) present revised versions of the Boulton *et al.* (1977), model incorporating satellite imagery and ground surveys along with geological and chronological evidence (Figure 5.9c and 5.9d). A revision of basal boundary conditions yields a much thinner ice sheet, with maximum ice thicknesses for Ireland of 750 m and 1000 m respectively.

Lastly, Paterson (1994) suggests that (assuming a laterally-unconstrained, perfectly plastic ice sheet resting on a horizontal bed), ice sheet thickness (h) is related to distance from the ice margin (L) in the manner outlined in equation 5.2:

$$h = (2\tau L/\rho g)^{1/2}$$

(Eq. 5.2)

where ρ is ice density (900 kg m^{-3}) and g is gravitational acceleration (9.8 m s^{-2}).

The value of basal sheer stress (τ) for extant ice sheets moving over a rigid bed generally falls within the range 50-100 kPa (Paterson, 1994). Even if an ice sheet of modest lateral extent (with limits similar to the present day coastline) is assumed to have existed over Ireland at the LGM,

theoretical maximal ice thickness could have reached between c. 1500 m ($\tau = 50$ kPa) and c. 2100 m ($\tau = 100$ kPa).

All the above underlines the degree of uncertainty surrounding the overall vertical dimensions of the Irish ice sheet and emphasizes the fact that there exists considerable scope for adjustment in the BIM- 1 LGM ice thickness over much of central Ireland.

5.5.1(i)c Retreat from the LGM (20 000 – 15 000 BP)

The onset of deglaciation began after 21 000 BP with (rapid) ice retreat taking place up until 19 000 BP. By this time, much of west and northwest Ireland was said to be ice free (e.g. Clark, 2006) with the ice margin in the Irish Sea occupying a position to the north of the Isle of Man (e.g. McCabe and Clark, 2003; Thomas *et al.*, 2004). Various authors had argued that the Isle of Man was ice free by c. 21 000 years ago, pointing towards the early evacuation of ice from the Irish Sea Basin (Shotton and Williams, 1971, 1973; Stephens and McCabe, 1977). However, the recent work of Thomas *et al.* (2004) and Roberts *et al.* (2006) questions the validity of the radiocarbon dates used to constrain this early retreat episode, instead suggesting that ice could have remained on the Isle of Man until much later in the deglacial cycle.

Interrupting this ice retreat from the LGM was a large scale ice re-advance which affected much of the country (e.g. McCabe *et al.*, 1998; McCabe and Clark, 2003; McCabe *et al.*, 2005). This 'Killard Point' or 'Heinrich 1' event involved ice re-advance as far south as the Shannon and onto the north of the Isle of Man in the Irish Sea Basin (Thomas *et al.*, 2004). However this event was brief and ice retreat followed shortly after, with the region ice free by 15 000 BP (e.g. McCabe *et al.*, 2005). During the Younger-Dryas event (c. 13 000 – 11 500 BP) ice was confined to mountain cirque glaciation (Murray-Gray and Coxon, 1991; Coxon, 1997) and was on a significantly smaller scale than its equivalent in the Scottish Highlands (e.g. Thorp, 1986).

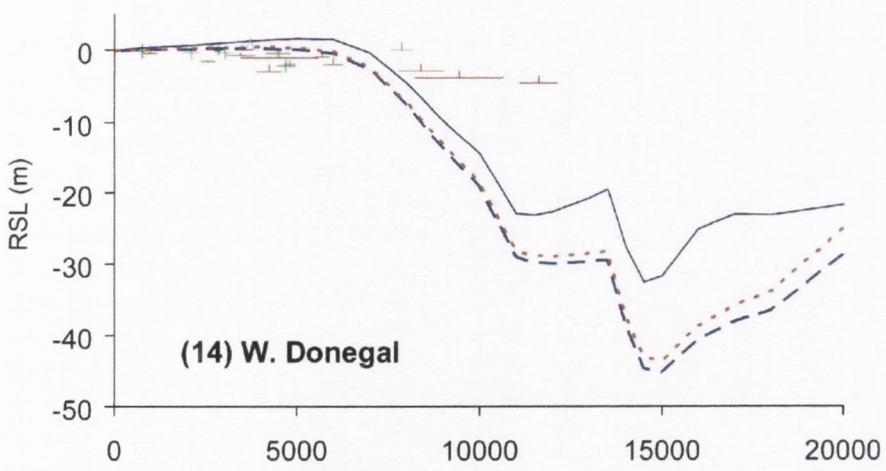
The underestimation of maximal Late Devensian RSL by BIM-1 indicates the need for greater crustal depression at this time. This requires either an earlier, or larger LGM ice build up, and/or a more persistent ice sheet during deglaciation (e.g. slower melting). In contrast, the overestimation of the Holocene highstand around 6000 BP indicates that the model predicts too much glacial

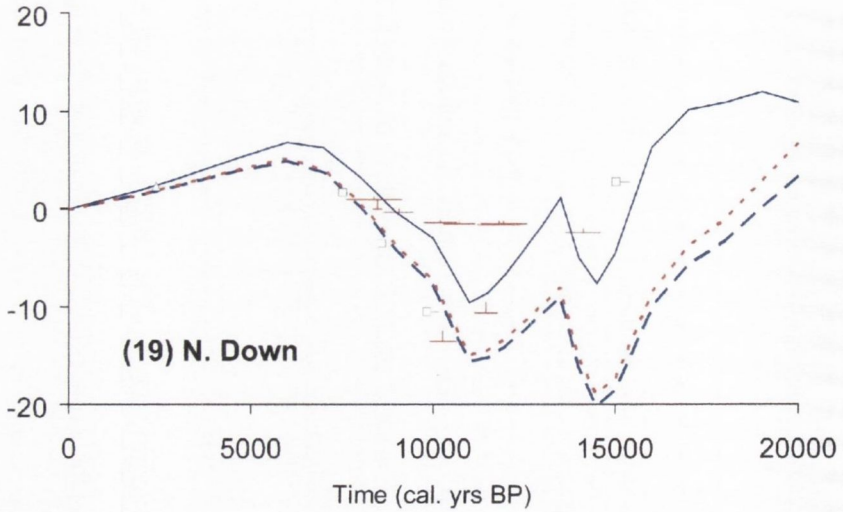
rebound by this time. This requires a reduction in LGM ice volume and/or persistence accompanied by more rapid melting during the deglacial phase. The latter therefore places clear constraints on the modifications acceptable to produce the required Late Devensian RSLs.

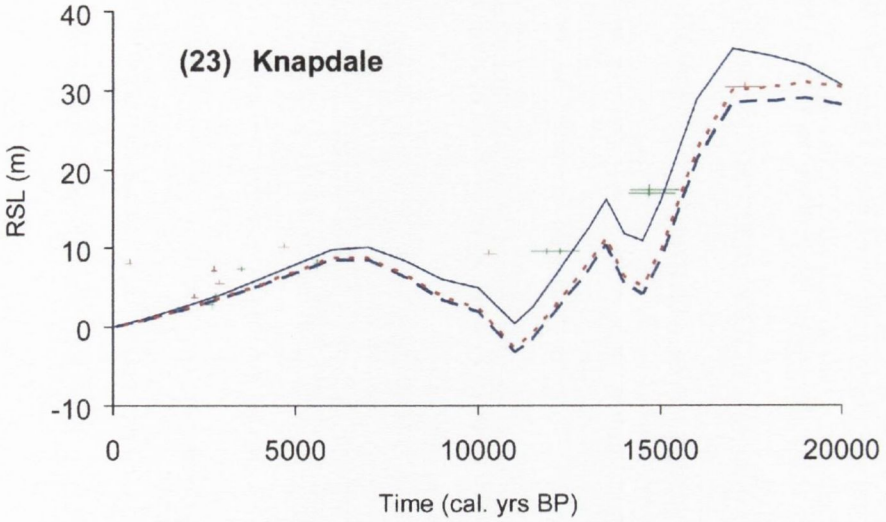
5.5.1(ii) Sensitivity analysis

In order to demonstrate the effects of varying ice volume at the time of the LGM and during the deglacial phase, two sets of sensitivity analyses are conducted using BIM-1, each with differing modifications to the Irish ice sheet component. In the past, researchers working outside the GIA modelling community have sometimes expressed concern at the accuracy of GIA ice model components at the local scale and the impact these inaccuracies may have on predictions of RSL (e.g. McCabe, 1997; McCabe *et al.*, 2005). For this reason, the influence of local-scale ice movements, such as the re-advance associated with the 'Killard Point' stadial included in the nominal ice model for BIM-1 are examined first. This is achieved by considering the contrasting RSL changes that arise if Ireland were completely ice-free after 19 000 BP (Model A). A second model (Model B) is also considered in which the spatial extent of Irish ice to the west and northwest of Ireland at the time of the LGM is reduced, with (onshore) ice limits similar to those proposed by Stephens and Synge (1965), and Bowen *et al.* (2002).

The results of these two experiments are plotted in Figure 5.7 alongside the original RSL predictions of BIM-1. It is clear from this that in the south and southwest of Ireland (away from the centre of crustal loading), the effects of removing all ice from Ireland after 19 000 BP (Model A) are small. However, in the north of Ireland the changes are more pronounced. The greatest divergence from the nominal BIM-1 ice model occur at c. 15 000 BP with a reduction in RSL predictions of c.10-12 m. During the Holocene there is also an appreciable difference between BIM-1 and Model A predictions, with the latter commonly reducing the amplitude of the Holocene highstand by c. 2-3 m. The greatest difference between BIM-1 predictions and the RSL predictions from Model B (removal of west coast ice at the LGM) occurs, perhaps unsurprisingly, at sites located in NW Ireland. However, even in west Donegal (14), where the largest change in RSL is experienced, the







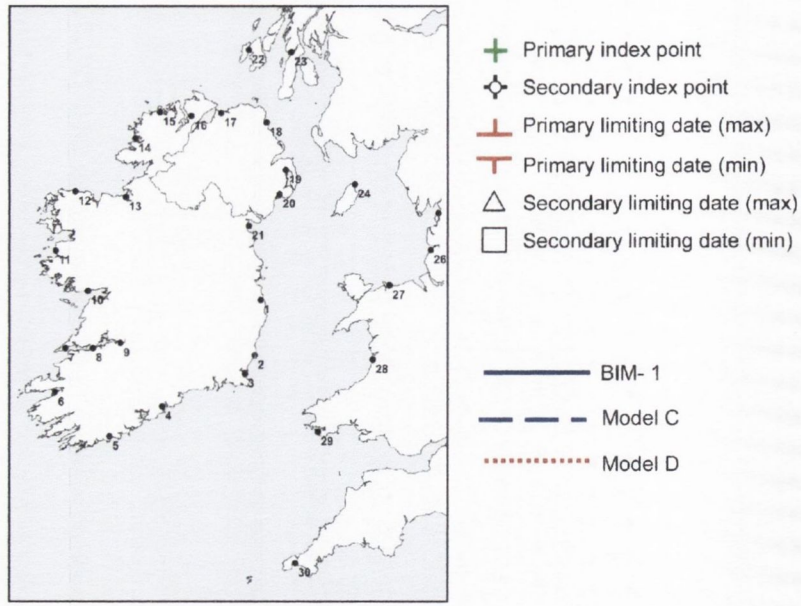


Figure 5.10 Relative sea level predictions generated by BIM- 1 (*solid line*); Model C (*long dashed line*) and Model D (*dotted line*)

effect of total offshore ice removal only contributes to c. 5 – 6 m of RSL lowering at 20 000 BP. At all sites, the influence on the Holocene GIA signature is not significant.

5.5.1(iii) Improving the fit with empirical RSL evidence

The results of these sensitivity analyses clearly demonstrate the requirement for a very thin ice sheet during the deglacial phase, since significant reductions in the amplitude of the Holocene highstand (and the associated geographical location of the Holocene 0 m isobase) cannot be generated by local-scale changes in LGM ice thickness alone (Model B). Whilst the results from Model A significantly improve the fit with Holocene RSL evidence, the position of the Holocene 0m isobase in Donegal is still too far to the west.

A further set of analyses are performed to examine the effects of changing the duration of ice loading, initiating deglaciation shortly after 22 000 BP (e.g. McCabe *et al.*, 2005) as opposed to after 21 000 BP in BIM-1. In the first analysis (Model C), the same pre-LGM and LGM ice sheet as BIM- 1 are adopted and, in accordance with the earlier sensitivity analyses, a very thin Irish ice sheet of approximately half the ice thickness of BIM-1 during the deglacial phase is used. The second analysis (Model D) is identical to Model C except that, at the time of the LGM, a more laterally extensive ice sheet of similar morphology to that proposed by Sejrup *et al.* (2005) is invoked. In this model, offshore ice thicknesses approach c. 650 m off NW Donegal with ice extending out to the continental shelf break. Ice thickness in the northern Irish Sea Basin are also increased to a similar thickness in order to reconcile glaciological evidence suggesting that Snaeffel (621 m) on the Isle of Man was transgressed by ice at the LGM (e.g. Bowen, 1973; Boulton *et al.*, 1977).

Both Figure 5.10 and Table 5.3 show a comparison between RSL predictions generated by the nominal ice model for BIM-1 and the revised ice models C and D. Once again, away from the centre of ice loading over the north of Ireland, the disparity between RSL predictions from the three models becomes insignificant. However, around the north of Ireland coastline the differences become more apparent. Predictions from Model C deliver a far better fit with the Holocene observational RSL data than BIM- 1 and the Holocene 0 m isobase no longer lies to the west of Sligo. A brief comparison between Model C and Model A (all Irish Ice removed from BIM -1 after 19

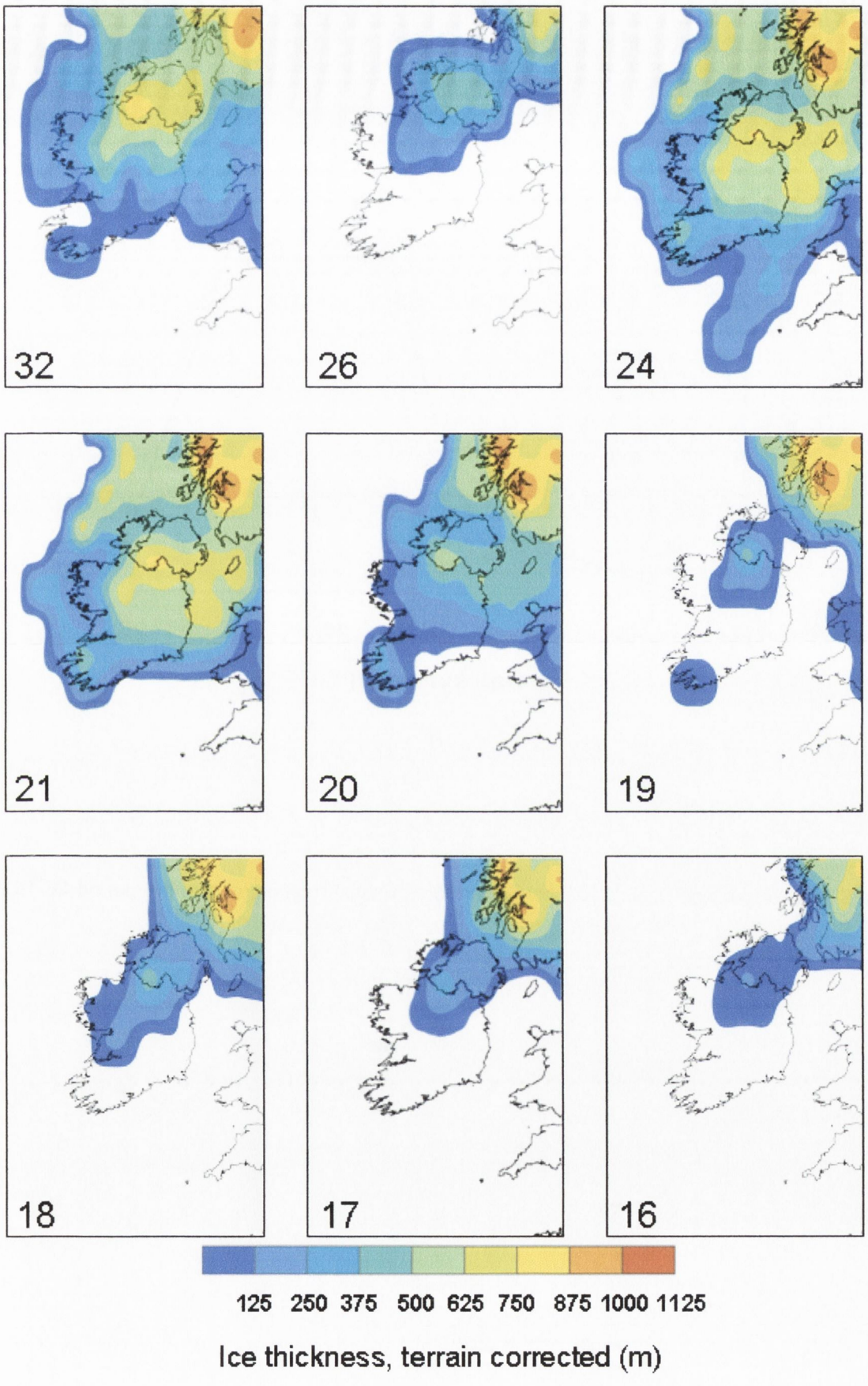


Figure 5.11 Time-space evolution of the 'best fit' ice model (Model E) (32-16 000 cal. BP)

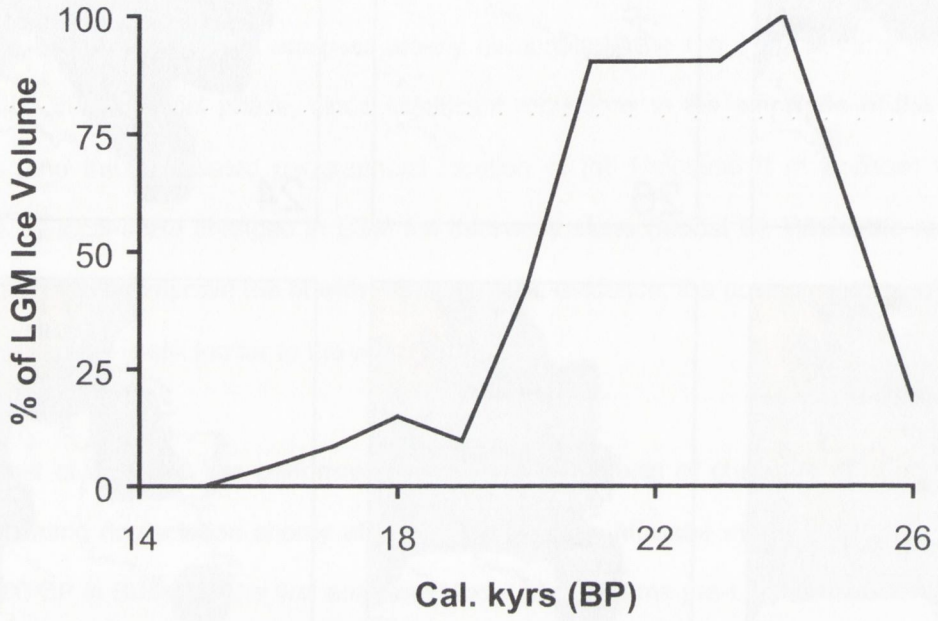


Figure 5.12 Percentage volumetric change of Model E (Irish component) in the period 26-15 000 cal. BP

000 BP) reveals that after c. 15 000 BP, predictions from the two models are largely identical with differences in the Holocene of no more than a few decimetres.

The RSL predictions produced by Model D (laterally extensive, thicker LGM ice) show even greater agreement with RSL data. The geographical extent of the Holocene highstand in the north is restricted, whilst concomitantly increasing maximal Late Devensian RSL predictions to heights above those delivered by Model C. However, even this increase is insufficient to satisfy the geomorphological RSL evidence and limiting data (minimum) provided by dated 'glaciomarine' muds from around the north and eastern Ulster coastline. Neither do predictions from Model D yield a particular good fit with the high resolution sea level index point data from Islay (22) and Knapdale (23) in southeast Scotland. In fact, in the north of Ireland, maximal Late Devensian RSL predictions derived from this ice model are no higher than those produced by BIM- 1 since the increase in RSL predictions induced by thicker ice is offset by the effects of rapid and earlier deglaciation.

5.5.1(iv) A best-fit Irish ice model

The analyses conducted in section 5.5.1(iii) demonstrate that whilst changes to the ice sheet in the deglacial phase have greatly improved the fit with the Holocene observational evidence, significant misfits remain between model predictions and geomorphological RSL evidence from the Late Devensian. Specifically, the maximal predictions of Late Devensian RSL are consistently below the geomorphological evidence of RSL for this time period. Because of this, it would appear that early deglaciation shortly after 22 000 BP is incompatible with the Irish RSL data. In light of these results, a revised Irish ice sheet component for BIM-1 is developed that is constrained by both glaciological and RSL data (Model E, Figure 5.11). This model uses the same pre LGM ice build up, and LGM duration (24 000 – 21 000 BP) as the nominal Irish Ice component of BIM-1. This best fit model differs from BIM-1 by having thicker ice in the Irish Sea and to the northwest of Ireland at the LGM (similar to Model D). In addition, ice is thickened across the Wicklow Mountains, Irish Midlands and over Kerry, in accordance with the trimline data of Rae *et al.* (2004) and Ballantyne *et al.* (2006). During the deglacial phase (20 000 to 16 000 BP), an almost identical ice sheet to that described in Model C is adopted, including rapid onset of deglaciation at 21 000 BP (Figures 5.11 and 5.12).

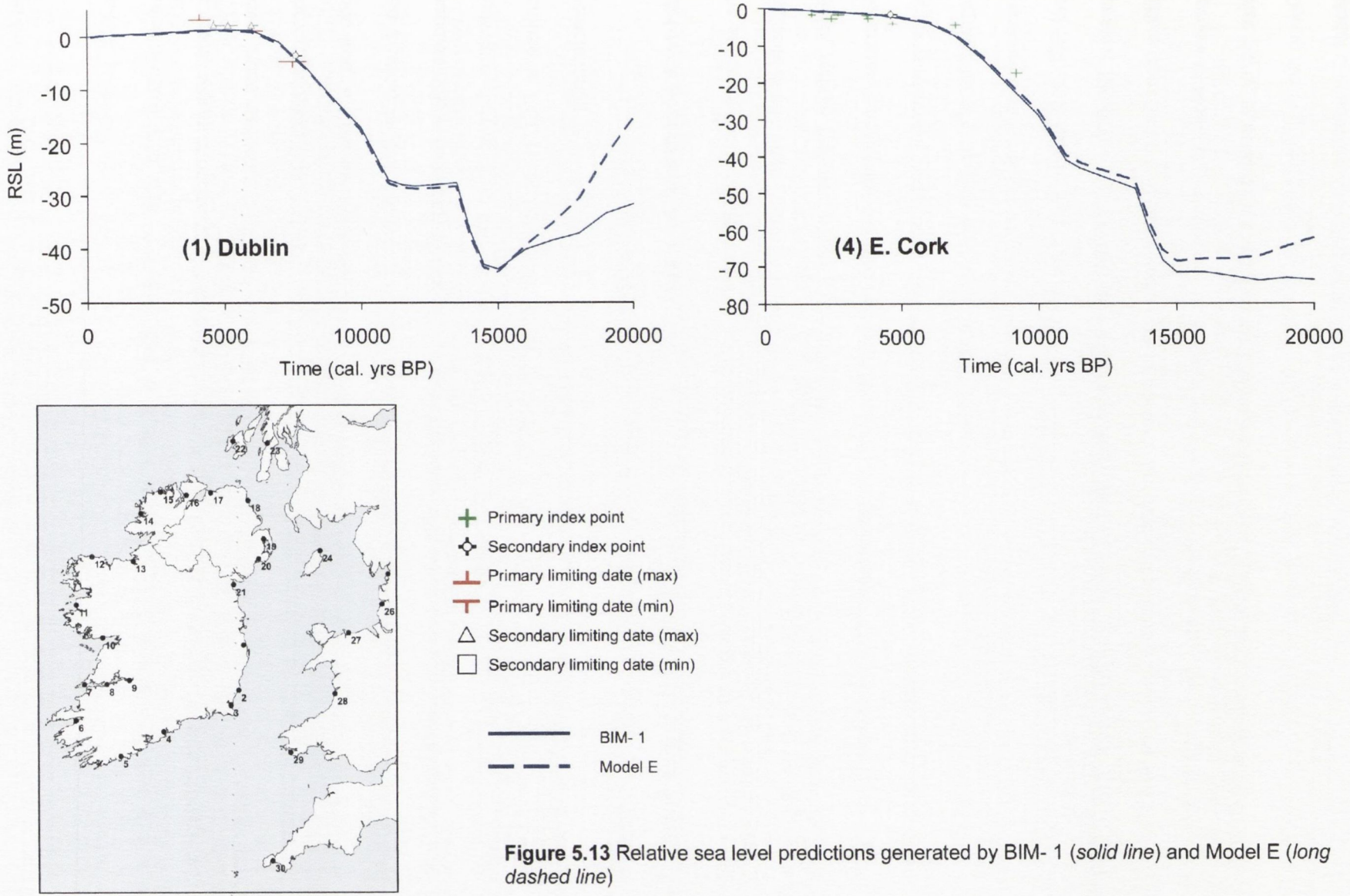


Figure 5.13 Relative sea level predictions generated by BIM- 1 (*solid line*) and Model E (*long dashed line*)

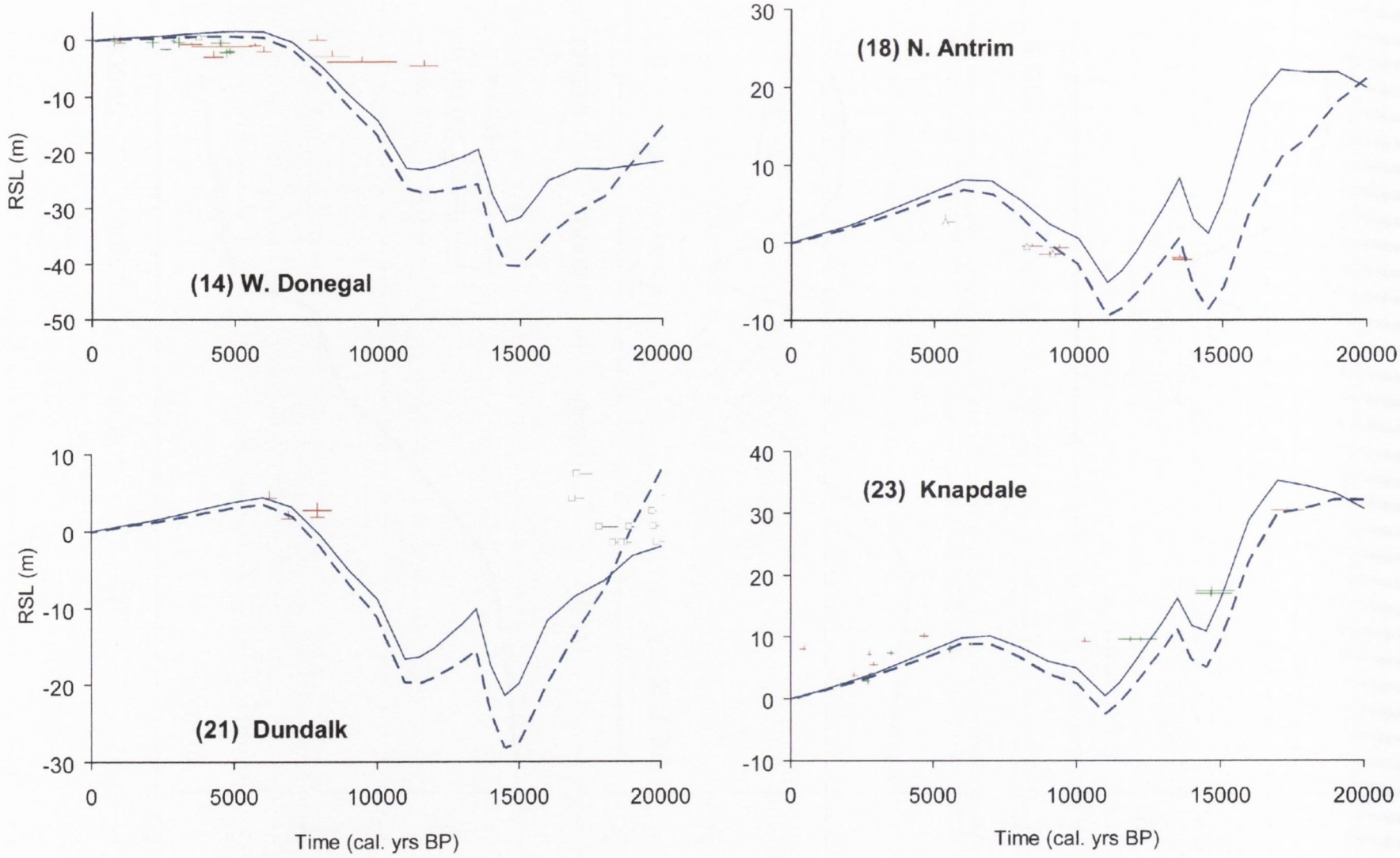


Figure 5.13(cont) Relative sea level predictions generated by BIM- 1 (*solid line*) and Model E (*long dashed line*)

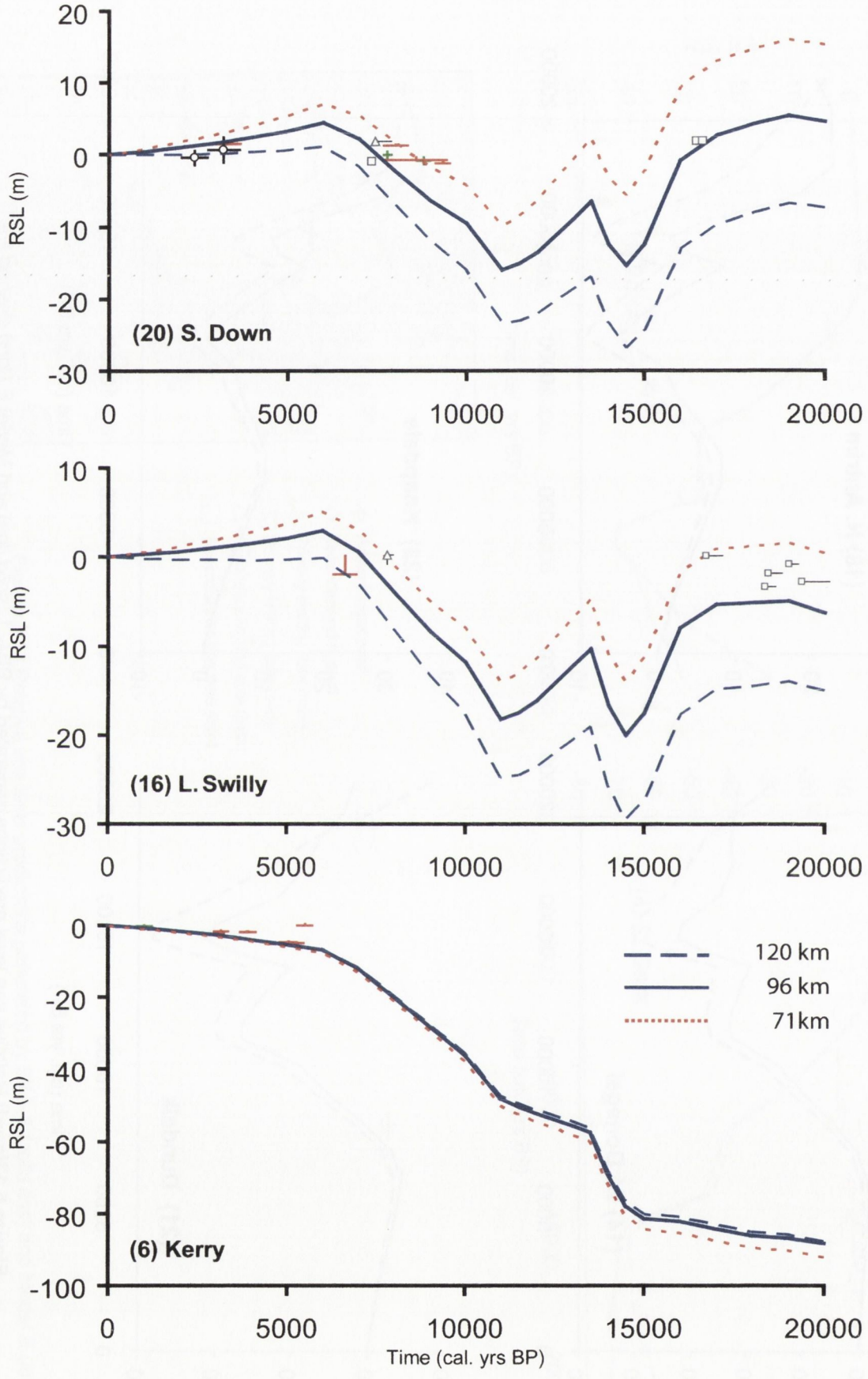


Figure 5.14 Relative sea level predictions generated by the Shennan *et al.* (2002) ice model run with a lithosphere thickness of 120, 96 and 71km

Despite significant differences between the Irish ice sheet component of the nominal BIM-1 model and the revised 'best fit' ice model (Model E), there are no discernible differences between the RSL predictions of the two models for the majority of sites along Ireland's east, south and west coasts (see Figure 5.13 and Table 5.3). The only exception to this is in Dublin (1), where the significantly greater volume of ice in the Irish Sea Basin associated with Model E results in more crustal uplift and higher predicted Holocene RSLs. Similarly, for most sites along the western coast of Britain, there is good agreement between RSL predictions from the two models.

Model E also produces a greatly improved fit with RSL data from the north of Ireland. The problematic Holocene highstand from Sligo (13) no longer exists and is greatly lowered in west Donegal (14). At all sites along the Ulster coastline, RSL predictions from Model E now yield a far better fit with the limiting data (maximum) in both the Holocene and back into the Late Devensian. This is achieved whilst also generating sufficient crustal depression to satisfy most of the raised shoreline data and a lot of the limiting data (minimum) from the AMS radiocarbon dated 'glaciomarine' muds.

It is apparent that at both Islay (22) and Knapdale (23), the fit with the index point data is not as good as with BIM-1. This misfit is caused by the thin Irish ice sheet during the deglacial phase which has enabled increased crustal rebound to occur and a subsequent lowering of RSL. However, local scale adjustments in both Scottish ice thickness and more specifically the rate of ice retreat should provide a solution to these model misfits.

5.5.2 The Role of the Earth Model

The Earth model used in these analyses (and outlined in section 5.2.1) is parameterised into three layers which correspond to the lithosphere, the upper mantle and the lower mantle. When considering patterns of GIA in regions once occupied by small ice sheets such as the BIIS, changes in the depth/viscosity of the lithosphere and upper mantle are of greatest significance. Adjustments to the lower mantle would appear less important since even the rebound associated with the much larger Fennoscandian ice-sheet is only weakly sensitive to this model parameter (Mitrovica and Peltier, 1991; Peltier, 1996).

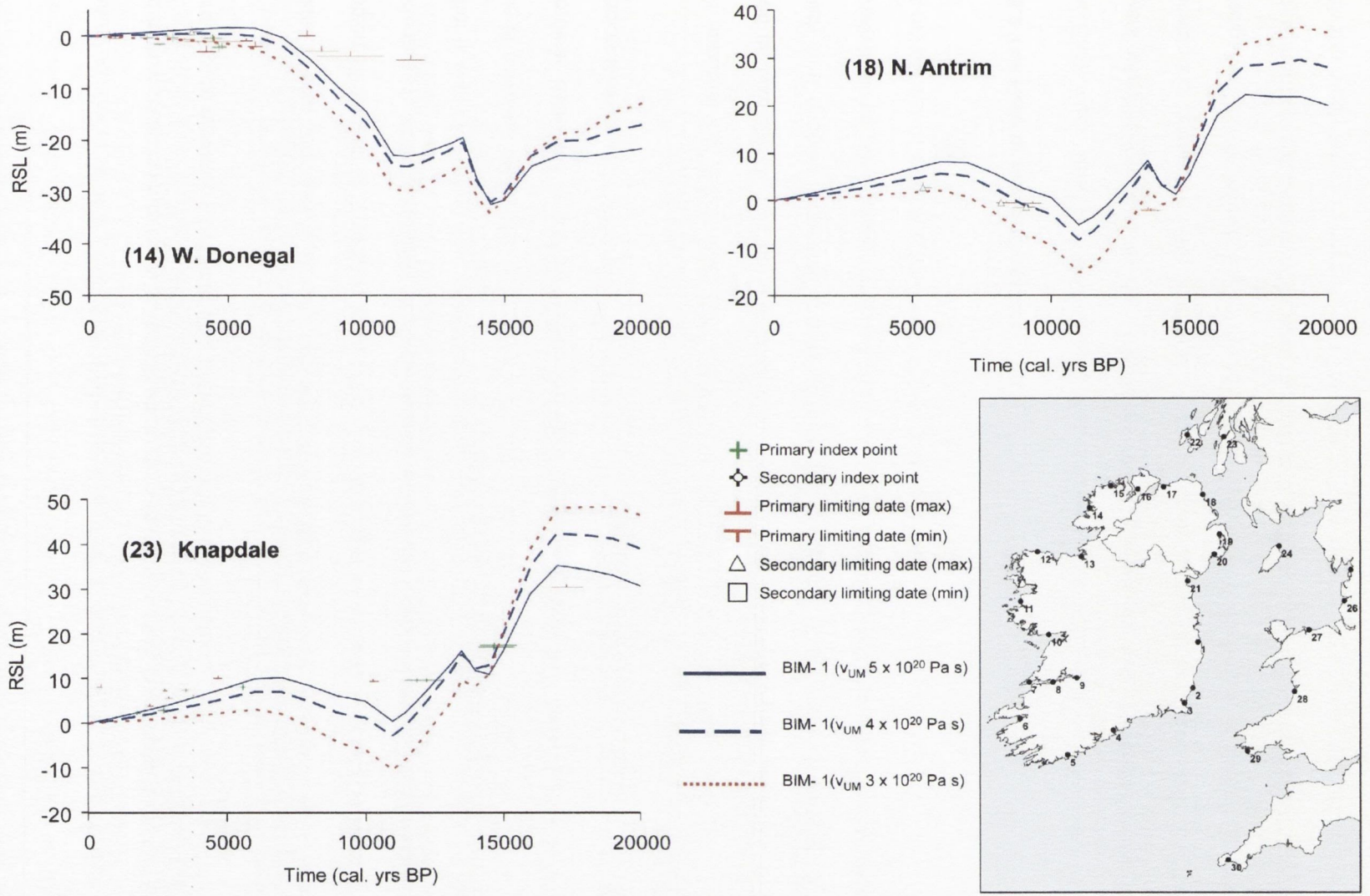
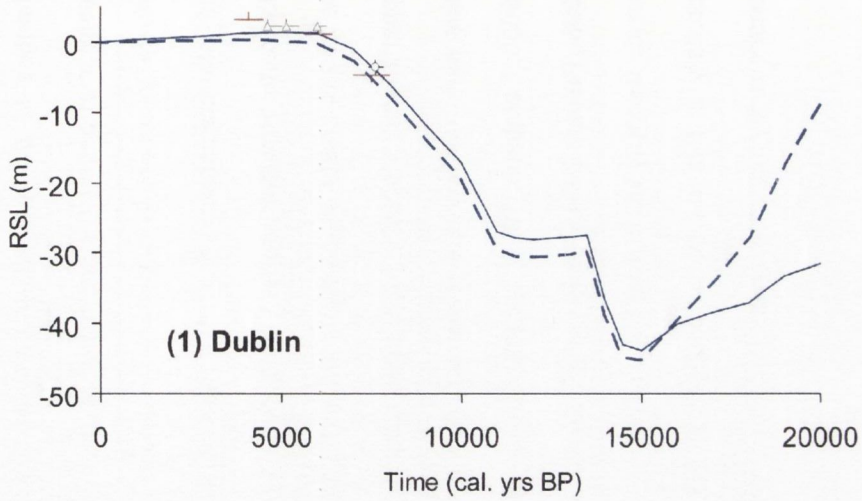
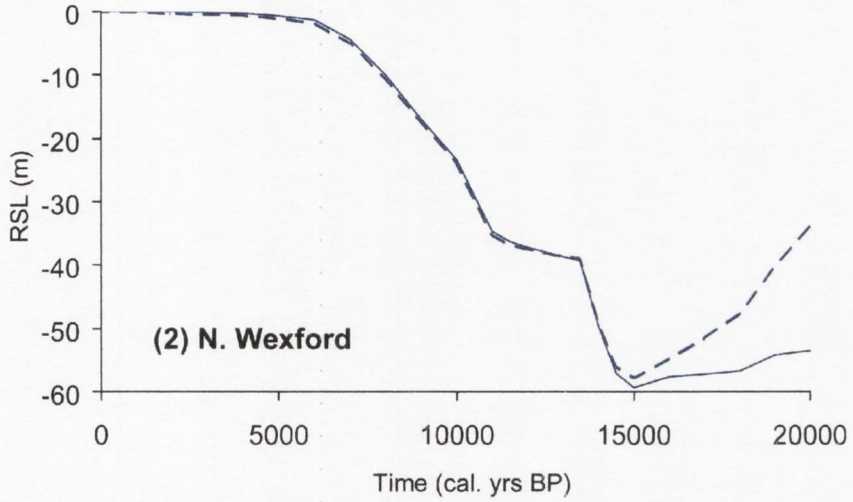


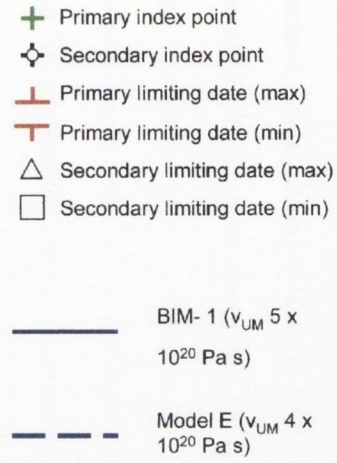
Figure 5.15 Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*), 4 x 10^{20} Pa s (*long dashed line*) and 3 x 10^{20} Pa s (*dotted line*)

Lithospheric thickness controls the magnitude of crustal depression in response to loading and any changes to this will either uniformly raise or lower predictions of RSL throughout the entire postglacial period (Figure 5.14). It follows that changes in the depth of the lithosphere will be unable to reconcile differences between predictions from BIM-1 and the RSL evidence since maximal predictions from BIM- 1 are generally too low in the Late Devensian and too high in the Holocene. However, upper mantle viscosity is of more relevance here since this parameter controls the rate of crustal response to loading-unloading episodes (e.g. Lambeck, 1993a,b). Estimates for upper mantle viscosity have been consistent between various modelling studies from the British Isles with most workers preferring either a value of 4×10^{20} Pa ^{14}Cs (Lambeck, 1993a; Lambeck, 1993b; Lambeck, 1995; Lambeck *et al.*, 1996; Shennan *et al.*, 2000a) or 5×10^{20} Pa s (Peltier, 1998; Peltier, 2004; Peltier *et al.*, 2002; Shennan *et al.*, 2006a). Reducing the values of upper mantle viscosity will increase the response rate of the crust to loading and unloading episodes. The corollary of this will be a reduction in the rate of crustal deformation during the Holocene and consequently induce a lowering of RSL predictions during this period.

To examine the impact of changes in upper mantle viscosity, a final set of sensitivity analyses are performed on BIM-1 by producing RSL predictions generated with reduced upper mantle viscosity values of 3×10^{20} Pa s and 4×10^{20} Pa s. These are compared with the original RSL predictions from BIM-1 (v_{UM} 5×10^{20} Pa s) and the results presented in Figure 5.15 and Table 5.3. In the south and west of Ireland, where the eustatic contribution to RSL change appears to dominate over the glacio-isostatic component, the reduction in upper mantle viscosity has negligible impact on the RSL predictions. However, toward the centre of loading in the north the impact of these changes becomes far more apparent. At all sites around the Ulster coastline, reduced upper mantle viscosity raises maximal predictions of RSL by, in the case of the v_{UM} 3×10^{20} Pa s model, in excess of 10 m. Conversely, the magnitude (and geographical range) of the Holocene highstand is reduced: in west Donegal (14) it is lowered by c.2m by adopting the v_{UM} 4×10^{20} Pa s model, whilst the v_{UM} 3×10^{20} Pa s model has the effect of removing the highstand altogether. In fact, when adopting the low viscosity (v_{UM} 3×10^{20} Pa s) Earth model, the greatest magnitude of Holocene highstand in Ireland is no more than +2.5 m.







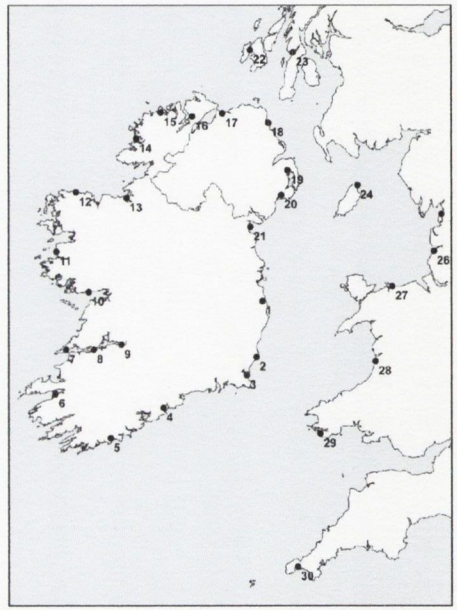


Figure 5.16 Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

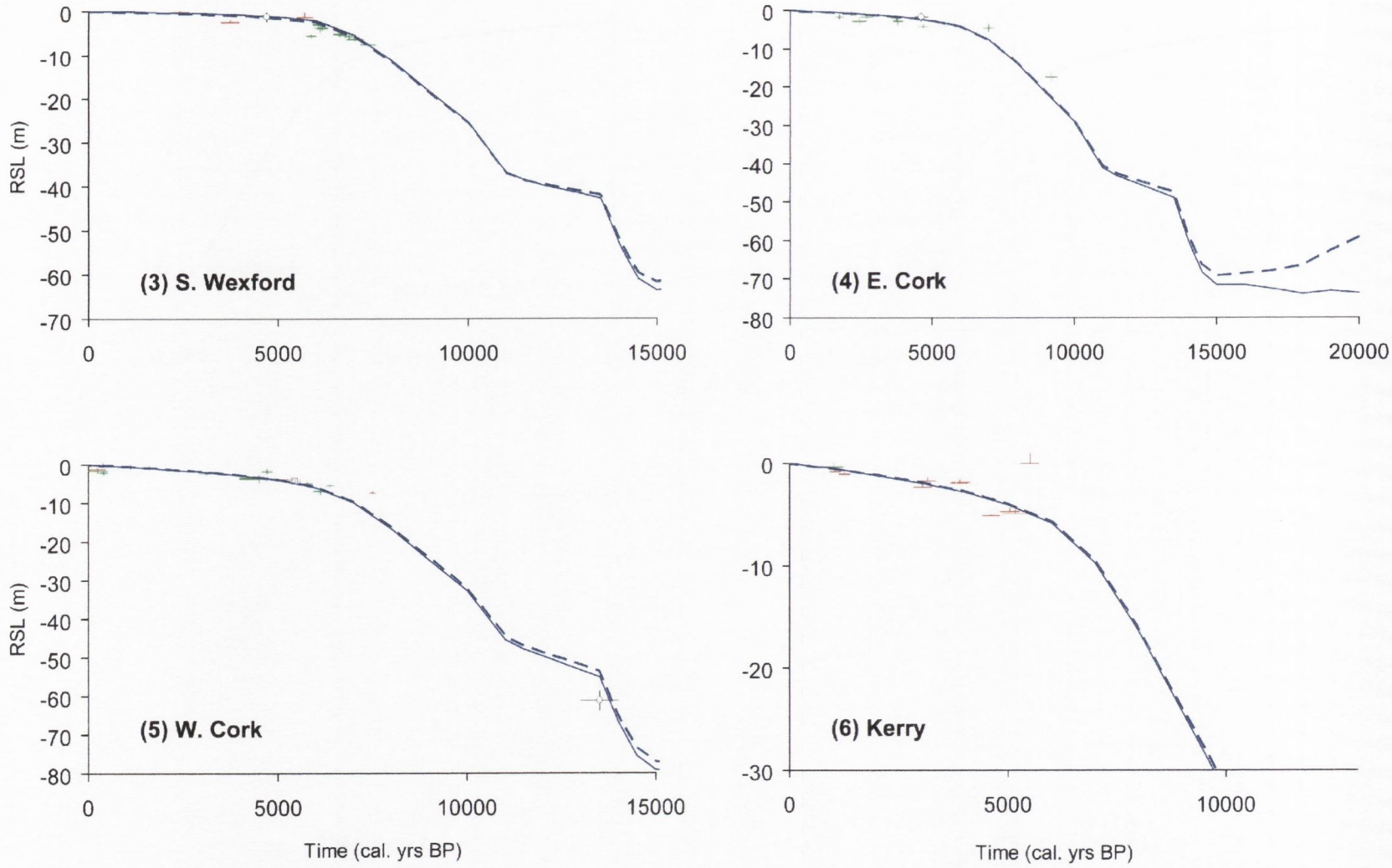


Figure 5.16(cont) Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

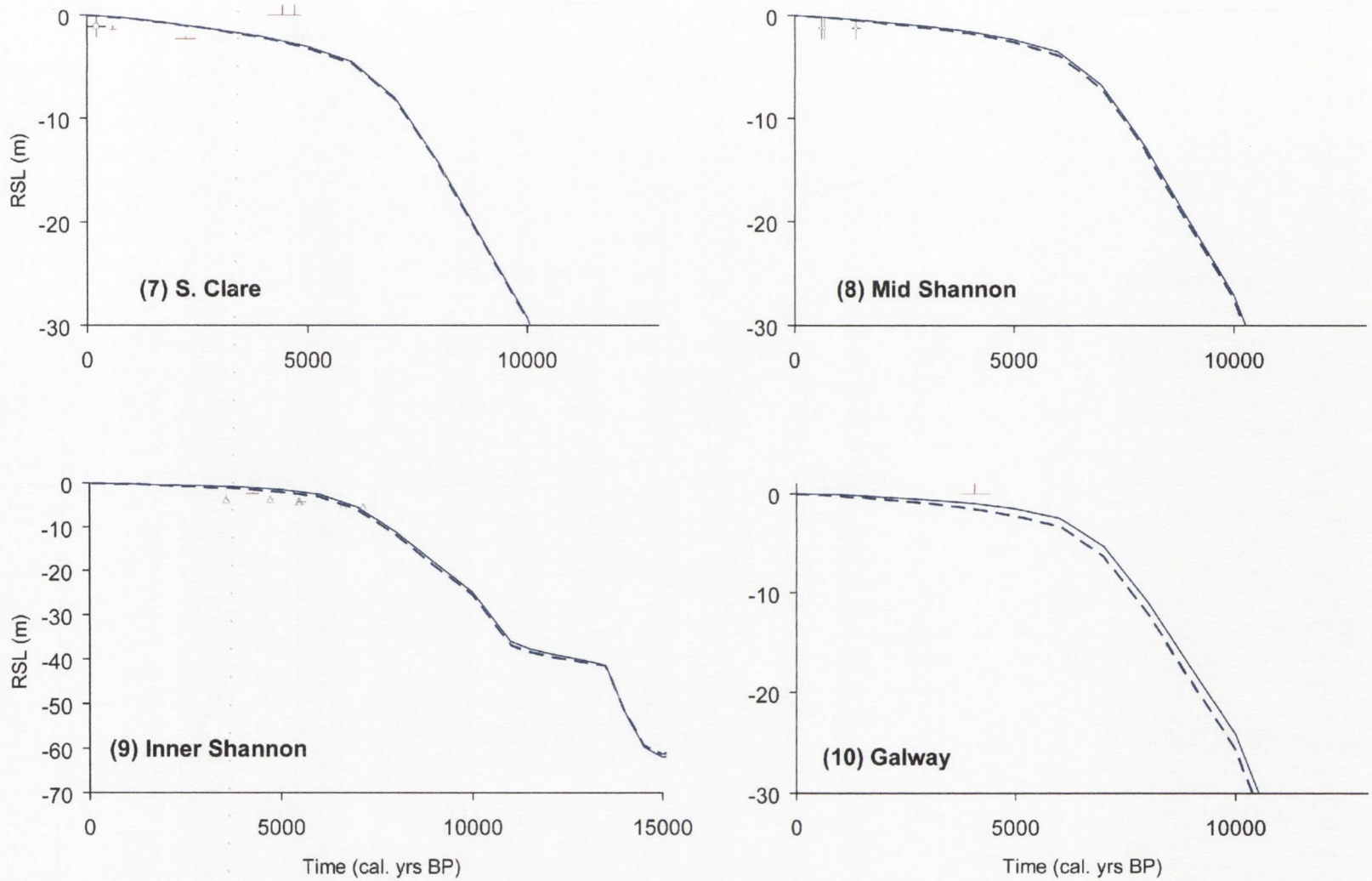


Figure 5.16(cont) Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

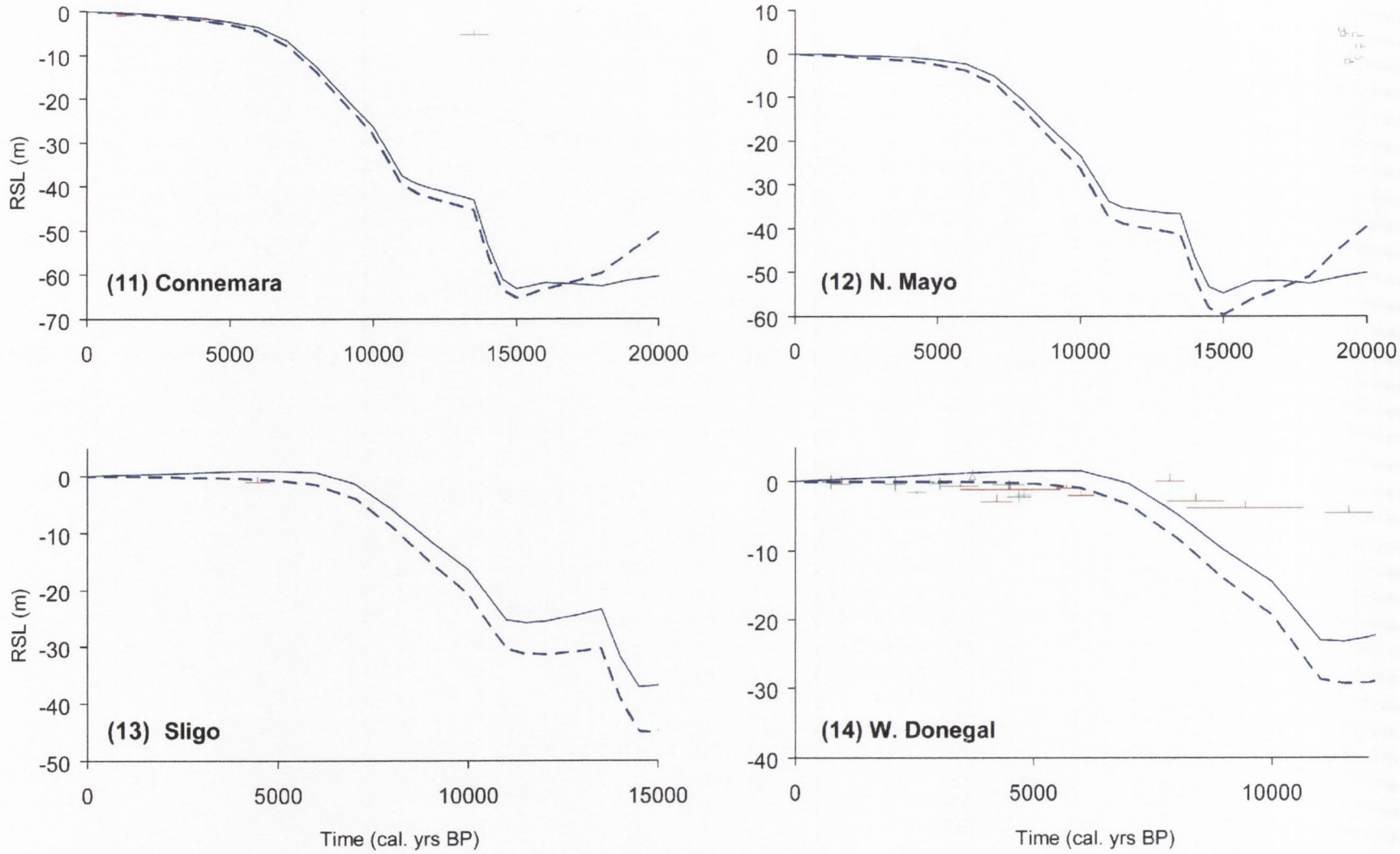


Figure 5.16(cont) Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

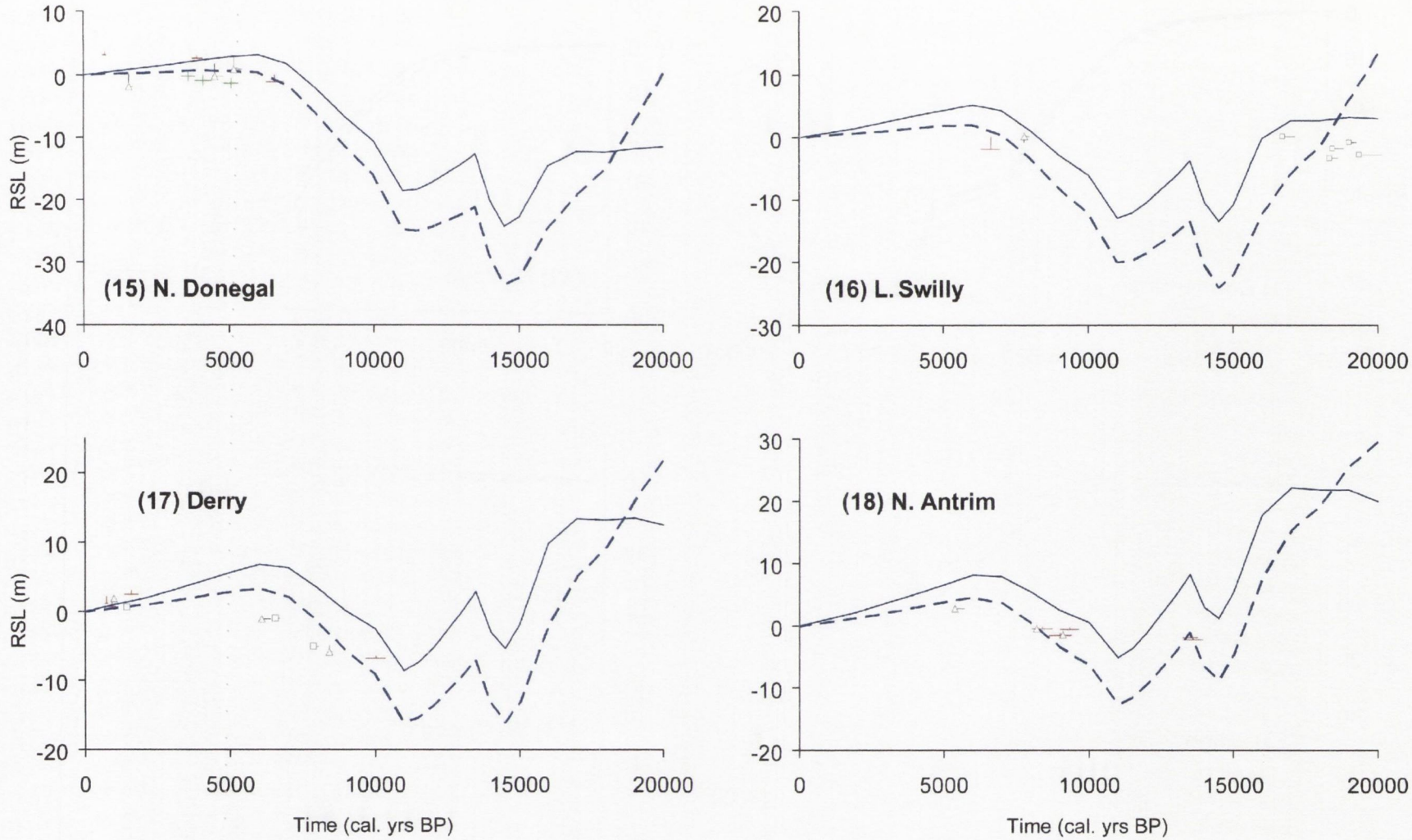


Figure 5.16(cont) Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

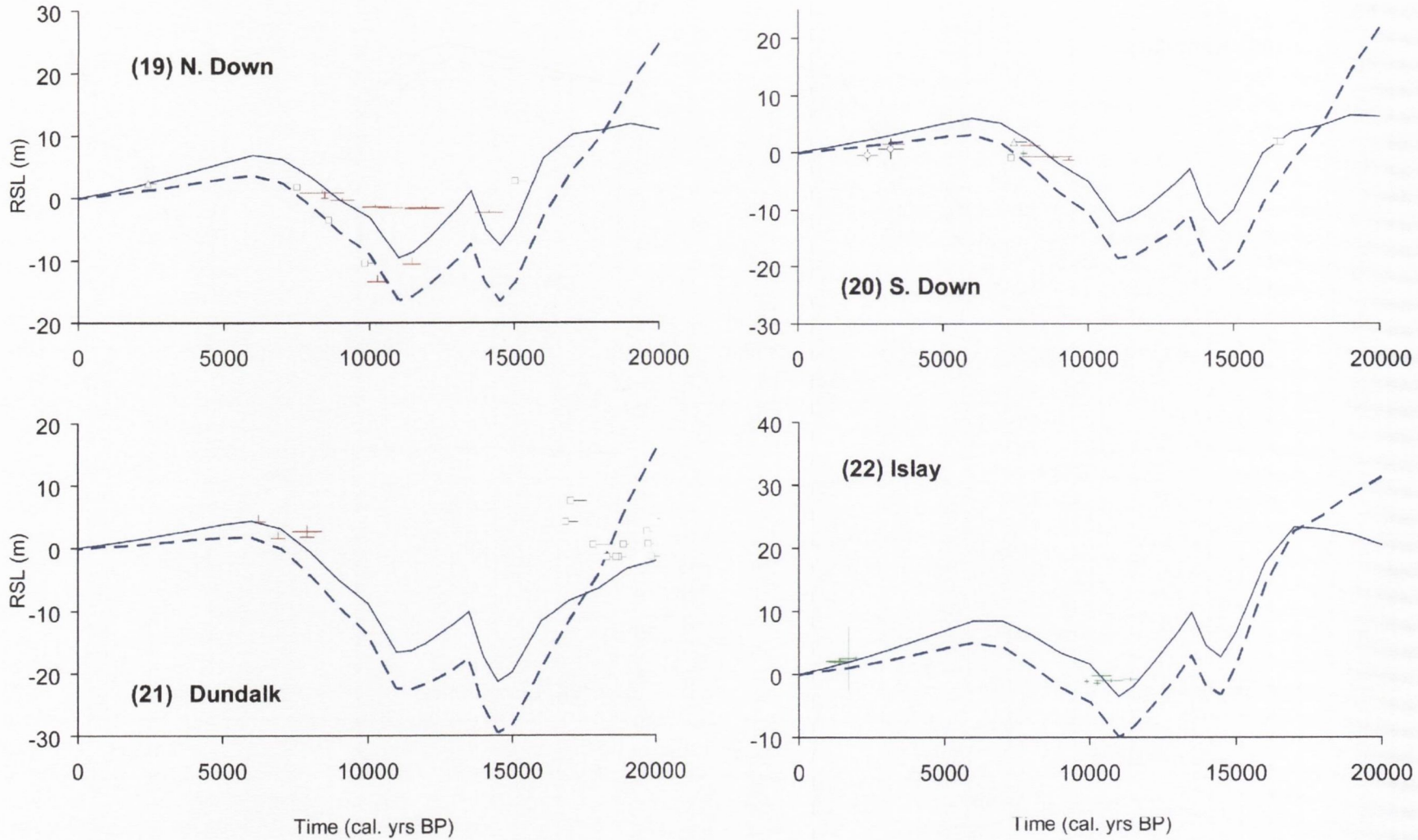


Figure 5.16(cont) Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

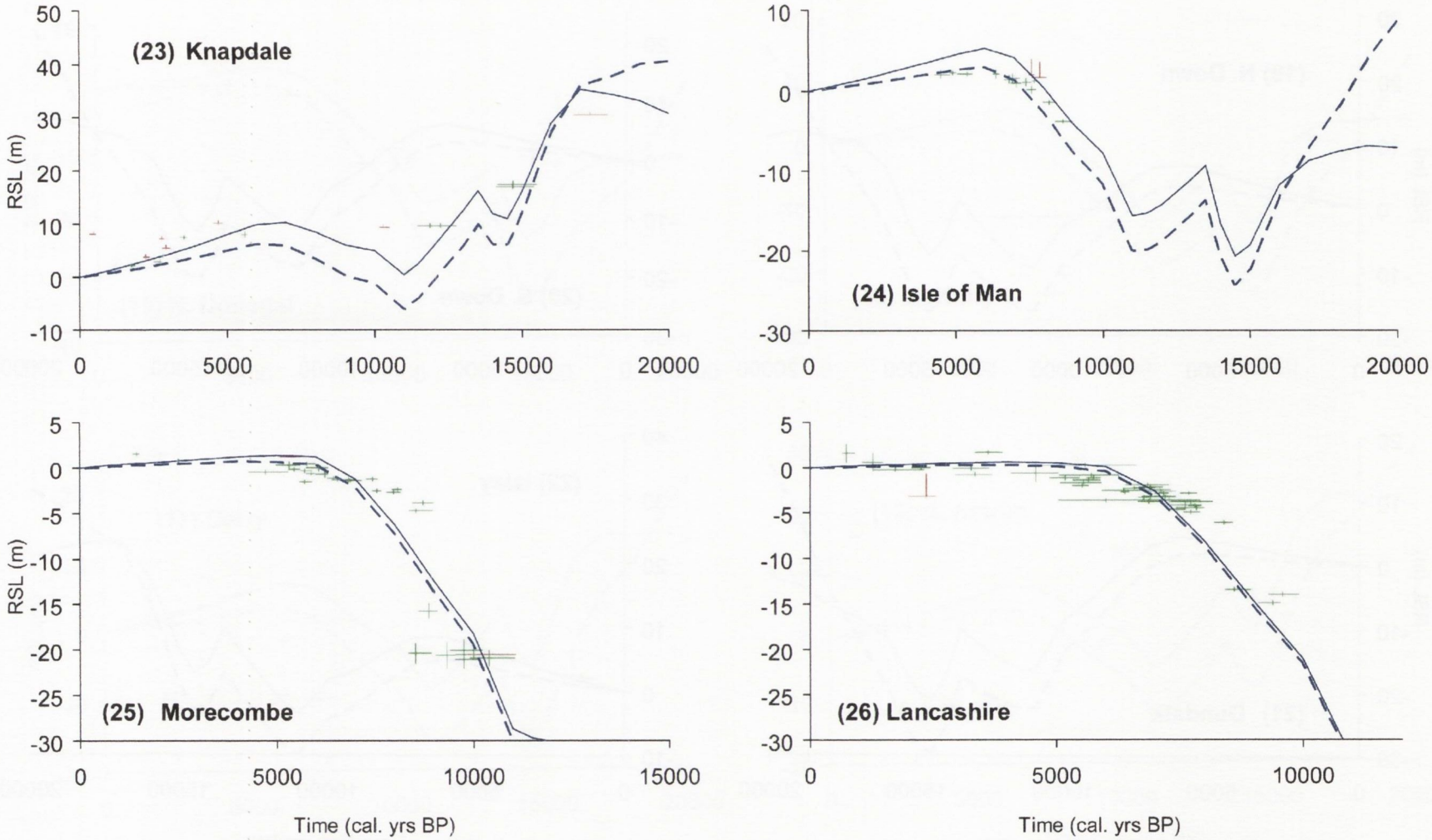


Figure 5.16(cont) Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

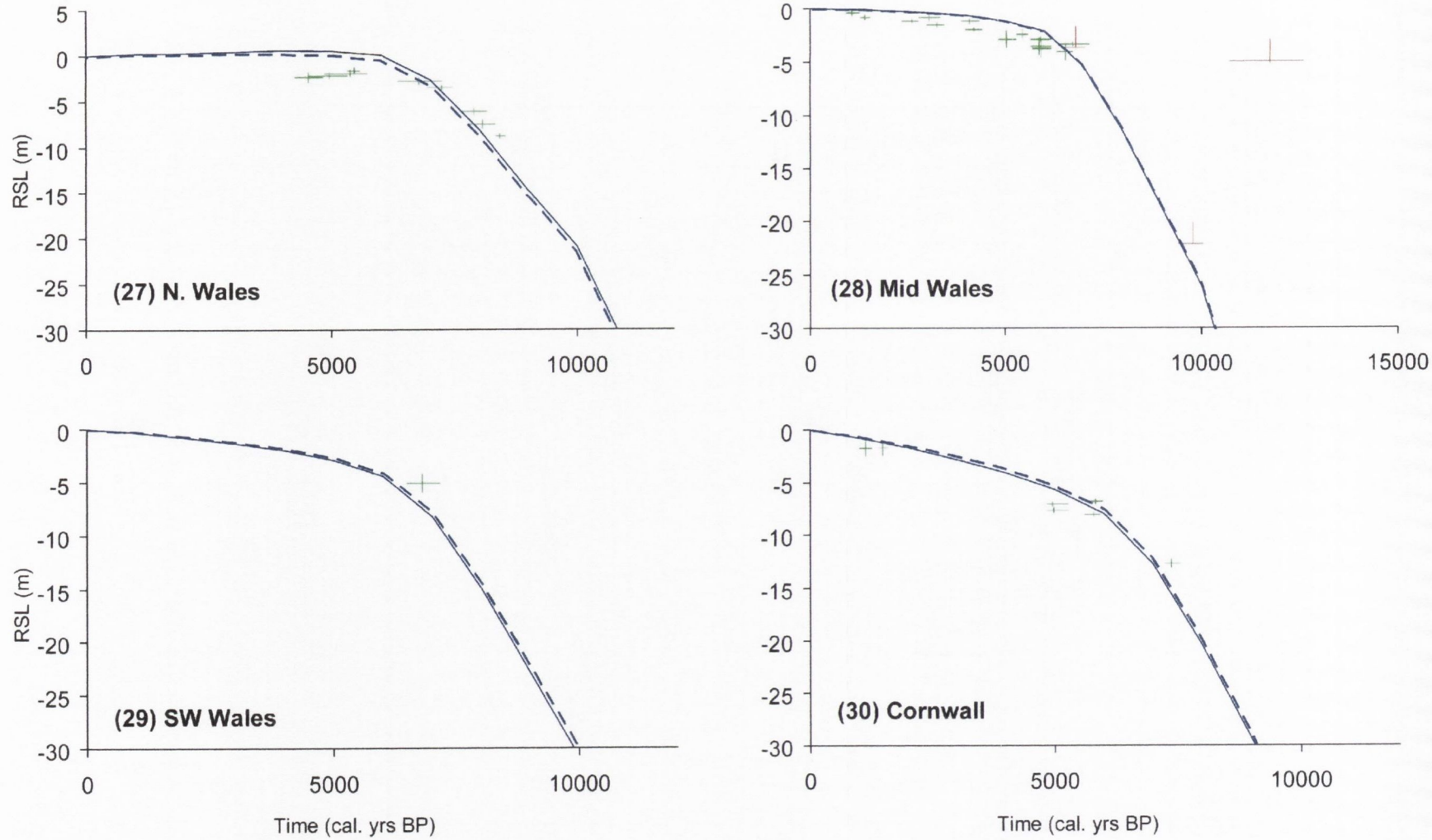


Figure 5.16(cont) Relative sea level predictions generated by running BIM- 1 with v_{UM} 5 x 10^{20} Pa s (*solid line*) and Model E with v_{UM} 4 x 10^{20} Pa s (*long dashed line*)

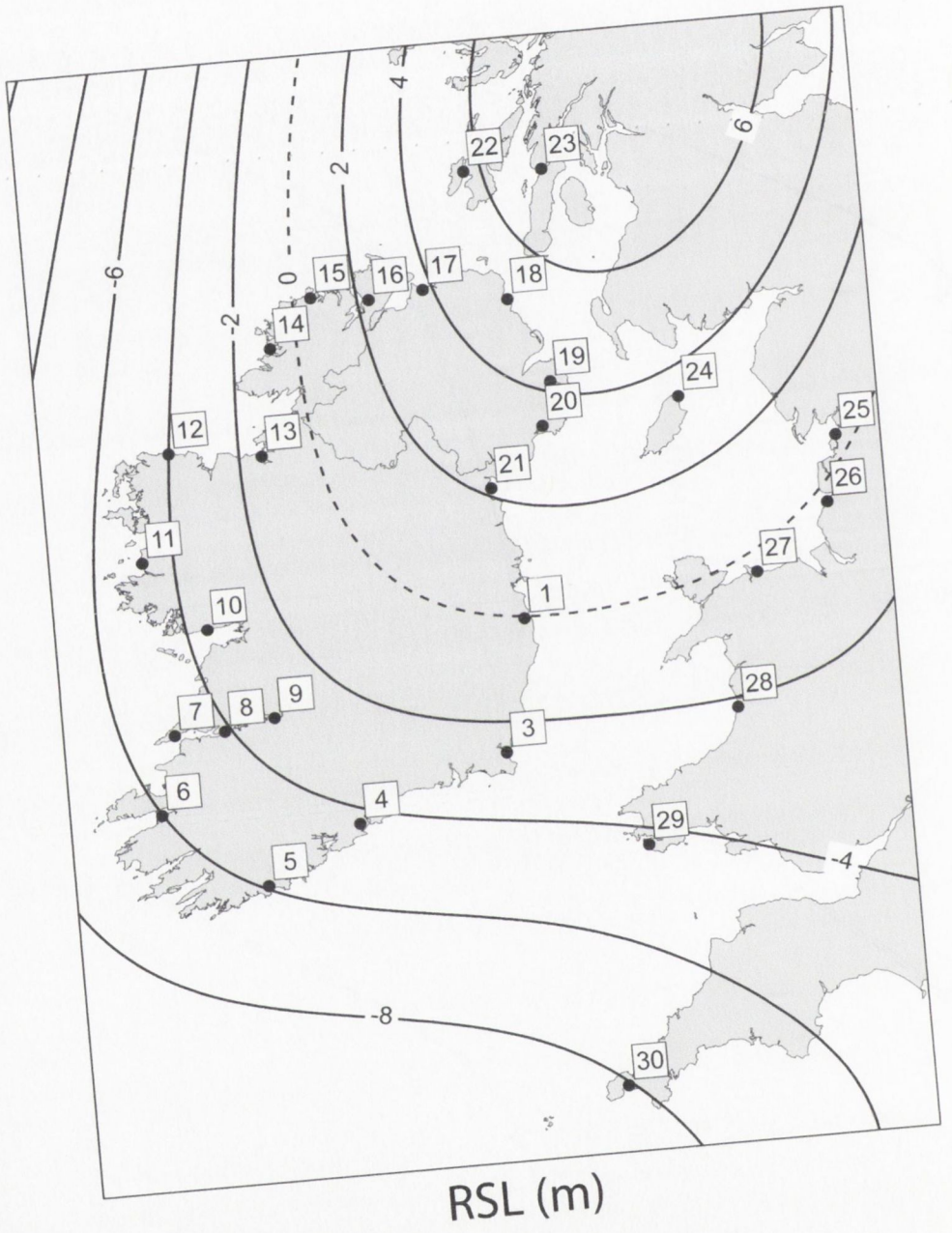


Figure 5.17 Model E RSL isobase map at 6000 BP run with v_{UM} 4 x 10^{20} Pa s

Finally, the effects of running the best fit model (Model E) with a lower upper mantle viscosity Earth model of 4 x 10^{22} Pa s are considered. Reference to Figures 5.16, 5.17 and Table 5.3 shows that this combination of a thin Irish ice sheet during the deglacial phase coupled with a reduced upper mantle Earth model does yield a close fit with the empirical RSL data. The Holocene highstand is removed from west Donegal (14) and model predictions are now in good agreement with the index point data from this locality. In north Antrim (18) (where the isostatic component of the RSL signal is at its greatest) the Holocene highstand is c. +5 m. This figure is significantly lower than the c. +9m highstand predicted by running BIM- 1 with an Earth model of v_{UM} 5 x 10^{20} Pa s (such as it was in the original Shennan *et al.*, 2006a analyses). Elsewhere around the Ulster coastline, model predictions satisfy all the constraints placed by the primary limiting date evidence as well as a large number of the secondary limiting dates.

5.6 Discussion

The analyses undertaken in section 5.5 outlines the extent to which model predictions of RSL respond to adjustments in both ice and Earth model input parameters. Significant improvements in the fit between BIM- 1 predictions and the Irish RSL data can be achieved through altering the Irish component of this regional ice model yet changes in the ice model alone have not completely reconciled discrepancies with observations of RSL. In particular, the Holocene highstand remains in west Donegal and this is unsubstantiated by field evidence. Interestingly, a similar problem relating to the position of the 0m Holocene isobase on the west coast of Britain has been encountered in recent modelling studies of the British data (Shennan *et al.*, 2002; Shennan *et al.*, 2006a) and, as in the case of this investigation, a solution could not be found through adjustments in the regional ice model. Model predictions of the geographical location of the 0m isobase can be marginally improved through increasing the eustatic meltwater contribution since 7000 BP to a maximum of 3 m (e.g. Lambeck, 2002). However, Holocene highstands persist in both west Donegal and north Wales which are at odds with available field evidence.

Both Milne *et al.* (2006) and Shennan *et al.* (2006a) have shown that predictions of RSL in the postglacial are sensitive to ice sheet evolution in the pre-LGM period. For this reason, changes in the ice sheet thickness were considered during the temporary retreat phase (25 000 – 26 000 BP). However, removing this retreat stage only imparted a very minor GIA signal after the LGM (Figure

5.7). Because no further changes can be made to the Irish ice model without contradicting the empirical glaciological evidence, it would appear that the Earth model employed in the Shennan *et al.* (2006a) analyses is untenable with the Irish RSL evidence.

Significant alterations to BIM- 1 RSL predictions can also be achieved through lowering the upper mantle viscosity of the Earth model. A greatly improved fit between BIM- 1 predictions and the Irish RSL dataset is achieved by adopting an upper mantle viscosity of 4 x 10^{20} Pa s which is a similar value to that used by Lambeck and co workers in their earlier analyses from the region. However, the problematic Holocene highstand remains in west Donegal. Whilst this may be removed by lowering the upper mantle viscosity still further to a value of 3 x 10^{20} Pa s, the overall fit with available RSL data from the north of Ireland becomes extremely poor: The greatest Holocene highstand around the Ulster coastline is no more than 2.5 m above present and the RSL predictions become significantly at odds with the Holocene index point data from south Down (20) and south west Scotland. In order to reconcile the differences between the BIM-1 (v_{UM} 3 x 10^{20} Pa s) and the Holocene RSL data, a substantial increase in crustal loading would be required during a large part of the Late Devensian. However, the resultant rise in RSL in the Late Devensian time interval would almost certainly be at odds with available geomorphological evidence of RSL from Ireland. Similarly, it is hard to see how such a low viscosity upper mantle Earth model would be compatible with the long time-series records of RSL change from the west coast of Scotland, particularly since trimline evidence provides a strict upper bound for the maximum thickness of ice at the time of the LGM. Because of the above, such a low viscosity upper mantle is extremely unlikely to provide an acceptable solution.

The above demonstrates that a resolution for model misfits cannot be achieved through changes to one or the other of the ice/Earth model. Instead, it is evident that combined revisions to both the existing BIM- 1 ice model and the original Earth model are desirable in order to achieve a solution that satisfies the constraints provided by both the RSL and glaciological evidence from Ireland. Indeed, running the best fit ice model (Model E) with a slightly reduced upper mantle viscosity of 4 x 10^{20} Pa s would appear to provide a very good overall solution. Using this ice-Earth model combination satisfies nearly all the Late Devensian RSL evidence from around the Ulster coastline and in Donegal, the predicted 0m isobase now occupies the same geographical location as that

reconstructed through field evidence (Figure 5.17). On the Isle of Man and in North Wales, the fit with the Holocene data is considerably improved. Importantly, the well constrained 0m Holocene isobase on the east coast of Britain is largely unaffected by the revision to the Earth model.

An important, additional test of this revised Earth-ice model solution is the quality of the fit to the long time-series RSL records from Scotland (see Figure 5.16). On first inspection, it appears as though the new reduced viscosity model is incompatible with the index point data from both Islay and Knapdale. However, this need not hold true: BIM- 1 is constrained by glaciological evidence and is also inferred on the basis of the Earth model employed in the Shennan *et al.* (2002) investigation. Significantly, the analyses of Shennan *et al.* (2006a) has shown that the 96 km lithosphere used in the Shennan *et al.* (2002) paper is untenable with observational RSL data from Britain. This means that there now exists considerable scope to make local scale adjustments in the British Ice model, specifically in areas which are poorly constrained by trimline data. At this point, it is worth pointing out that if the British component of BIM- 1 is tuned to the RSL data using a reduced upper mantle Earth model, the associated changes in ice loading will alter predictions of RSL in Ireland. For the most part, these changes are likely to be negligible: However, they probably will have a noticeable impact on predictions in the north east of Ireland with an overall increase in the height of postglacial RSL predictions a likely outcome. This addendum is an important caveat to the accuracy of the current RSL predictions and the associated palaeogeographic reconstructions.

Very recently, Simms (unpublished data) and Kelley *et al.* (in press) have uncovered new evidence of postglacial RSL in north Antrim (18) and north Down (19). The former has used AMS radiocarbon dates from fossil limpets to infer a Holocene highstand of c. 8-10 m for this locality. The latter have used dates obtained from molluscan fauna entrained within inter-bedded sand and gravel units in Belfast Lough to evidence a sea level lowstand of c.-30 m msl at 13 400 BP. Both of these findings would appear to be at odds with the best fit ice model: run with an Earth model of V_{UM} 5 x 10^{20} Pa s, model E delivers predictions of the RSL lowstand depth of c.-16 m in north Down and in north Antrim, the highstand is suggested to have reached a magnitude of c.6.5 m (Figure 5.13). Similarly, run with an Earth model of V_{UM} 4 x 10^{20} Pa s, the best fit model delivers predictions of the RSL lowstand depth of c.-16 m in north Down whilst in north Antrim, the

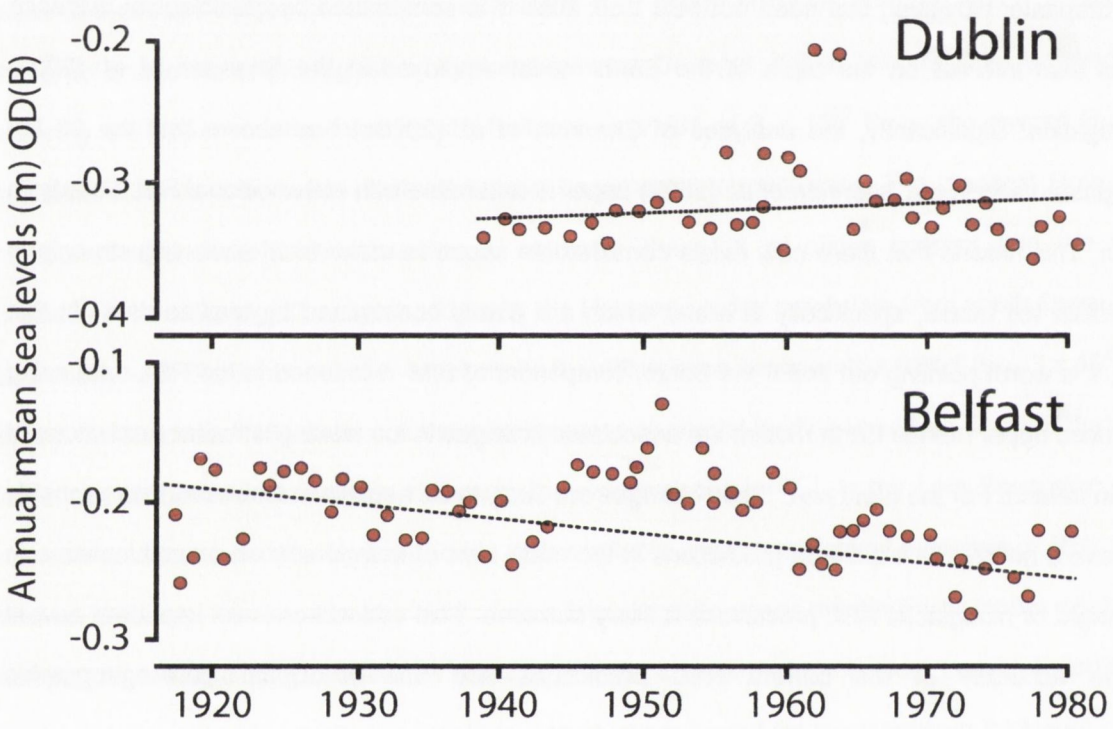


Figure 5.18 Annual mean sea-level variations for Dublin and Belfast (up to 1980). (Taken from Carter, 1982b)

highstand is suggested to have reached a magnitude of c. 5 m (Figures 5.16 and 5.17). It would be extremely hard to simultaneously reconcile both these lines of evidence through adjustments to either the ice or Earth model: To increase the highstand in north Antrim, either greater ice loading is required or an increase in upper mantle viscosity within the Earth model. Conversely, to increase the depth of the lowstand in north Down, either a reduction in ice loading or upper Earth mantle viscosity is needed. At this juncture, it is worth considering the methodologies employed within both these investigations. Aside from the potential errors associated with dating marine carbonate deposits, the indicative meanings of the adopted sea-level indicators are equivocal at best and therefore evidence of this type cannot be confidently used to test/constrain the model.

Significantly, all of the models considered in the analyses presented here deliver maximal predictions of Late Devensian RSL along Ireland's west coast that are well below present msl. No evidence is found in support of the suggestion by McCabe (2005); McCabe *et al.* (1986; 2005) and Thomas and Chiverrell (2006) of higher than present Late Devensian RSL along this stretch of coastline.

This study cannot offer unequivocal evidence for or against rejecting the South of Ireland End Moraine as the southerly limit of Late Devensian Irish Ice. However, in order to fit the RSL data, a significant thickness of ice (c.500 m) is needed over the Irish Midlands, which are only a short distance to the north of the South of Ireland End Moraine. If the moraine is taken as the true southern limit of LGM ice extent, it follows that the ice sheet would have had a very steep southern margin and this may be glaciologically implausible.

Station Name	Record Span	Number of complete yrs	Trend and S.D (mm yr^{-1})	Model E GIA Trend (mm yr^{-1})	
				V_{UM} 4 x 10^{20} Pa s	V_{UM} 5 x 10^{20} Pa s
Belfast	1918-1963	45	-0.25±0.34	-0.552	-0.844
Dublin	1938-1996	58	0.23±0.3	-0.149	-0.299

Table 5.4 Twentieth century mean sea-level trends for tidal stations at Belfast and Dublin - (see Woodworth *et al.*, 1999 for details of data quality and analysis) - and predicted rates of GIA (generated by Model E) since 1000 BP

Finally, Table 5.4 and Figure 5.18 show 20th century mean sea-level trends derived from the tidal stations at Belfast and Dublin (taken from Woodworth *et al.* 1999 and Carter, 1982b). These values are given along side Model E estimates of rates of GIA operating over the past millennia. Both sites

show negative trends of modelled GIA, reflecting the ongoing crustal response to deglaciation. This rate of rebound is highest with an Earth model of V_{UM} 5 x 10^{20} Pa s although does not exceed 1 mm yr^{-1} at either site. At Dublin, the tidal gauge record reveals a positive trend in 20th century msl change of 0.23 mm yr^{-1}, which contrasts with the situation at Belfast where msl has been falling at a rate of 0.25 mm yr^{-1}. When run with the Earth model (V_{UM} 4 x 10^{20} Pa s), the 20th Century trend in msl (at both sites) exceeds crustal rebound by approximately 0.3 to 0.4 mm yr^{-1} and this figure rises to around 0.6 mm yr^{-1} when the Earth model (V_{UM} 5 x 10^{20} Pa s) is employed. None of these values are in agreement with the work of Church *et al.* (2001) (IPCC third Assessment Report 2001); Church and White (2006) whose findings suggest that 20th Century (eustatic) sea-level rise has taken place at a rate of 1-2 mm yr^{-1}. Neither do they agree with the analyses of Woodworth *et al.* (1999); Shennan and Horton (2002) which employ UK RSL data to infer a similar 20th Century rising trend of around 1 mm yr^{-1}. This could mean that model predictions of contemporary crustal rebound at both Belfast and Dublin are too low. As previously mentioned, this is a strong possibility for Belfast where re-calibration of the British Ice Sheet using an Earth model of (V_{UM} 4 x 10^{20} Pa s) will likely increase the rate of late Holocene crustal rebound in this region. Alternatively, the tidal records from Belfast and Dublin may not be representative of the global 20th Century msl trend, either because they are too short or because they are heavily conditioned by local/regional scale factors such as meteorological forcing and anthropogenic alterations to the coastline. This is certainly a possibility at Belfast where reclamation of the Lough after the Second World War is said to have introduced an artificial falling trend in the tidal record (Carter, 1982b).

5.7 Palaeogeographic Reconstruction

An important aspect of research into the position of palaeo sea level is understanding how vertical changes in RSL translate into lateral migration of the coastline. In this region, the insularity of Ireland and the (possible) emergence of an LGM/postglacial land bridge with Britain has been of especial interest. For some time discussion has been prevalent in Irish Quaternary literature (e.g. Devoy, 1983, 1985, 1995; Sleeman *et al.*, 1986) with the topic of plant colonization in particular instigating the suggestion of a land-bridge to Ireland (Mitchell, 1972; 1986). However, and despite the volume of interest, many existing reconstructions in palaeogeography for this region remain at the level of speculation and the need for better data remains relevant (Devoy, 1995).

The reconstruction of palaeogeography may be achieved through solving the equation 5.3: (where the locus of $h(t) = 0$ determines the location of the shoreline at time t):

$$h(t) = h(t_0) - \Delta\zeta_{rsl}(t) + \delta h$$

(Eq. 5.3)

Lambeck (2004)

h(t) *equals altitude/depth of a point at time* ***t****,* ***h(t$_0$)*** *represents contemporary altitude/depth and* $\Delta\zeta_{rsl}$ *is sea-level change since time* ***t*** *(****δh*** *is a correction for any erosion or sedimentation that may have occurred in the interval* ***t*** *-* ***t$_0$*** *although in reality, this is very hard to accurately define).*

Present day topographic and bathymetric data for Ireland and adjacent areas was sourced from the National Geophysics Data Centre and is defined on a 2 minute grid. (The seafloor data are provided by Smith and Sandwell (1997) and are derived from satellite altimetry observations combined with shipboard echo-sounding measurements).

In Figure 5.19, palaeographic reconstructions based upon the best fit ice model (Model E) run with an upper mantle viscosity of 4×10^{20} Pa s are shown. Interestingly, it is apparent that despite the occurrence of significant isostatic and eustatic change since the LGM, there has been surprisingly little lateral migration of the coastline around Ireland. Indeed, for the most part it would appear that the Irish coastline has not deviated by more than about 30 km from its contemporary location throughout the entire postglacial period. This contrasts with the situation in Britain where widespread change in palaeogeography has occurred over the past 20 000 years, particularly in the English Channel and southern North Sea (Figure 5.20). This may be explained by the contrasting bathymetry between the two regions: The coastline around southern Britain is flanked by very shallow seas whereas around much of Ireland, it is characterized by a relatively steeply sloping shelf.

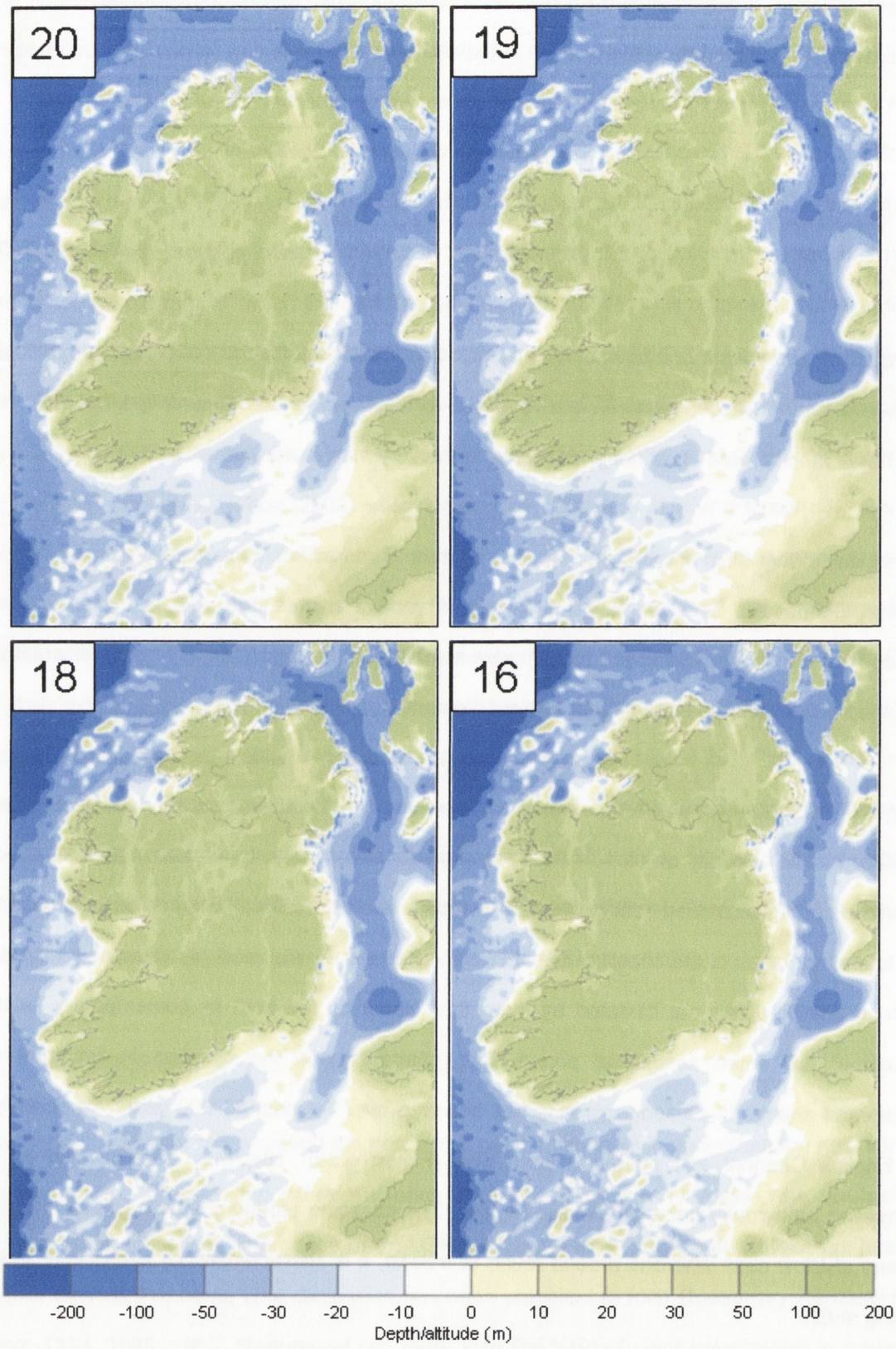


Figure 5.19 Reconstructions of palaeogeography (during the postglacial period) for Ireland and adjacent areas. (Time intervals in cal. kyr BP)

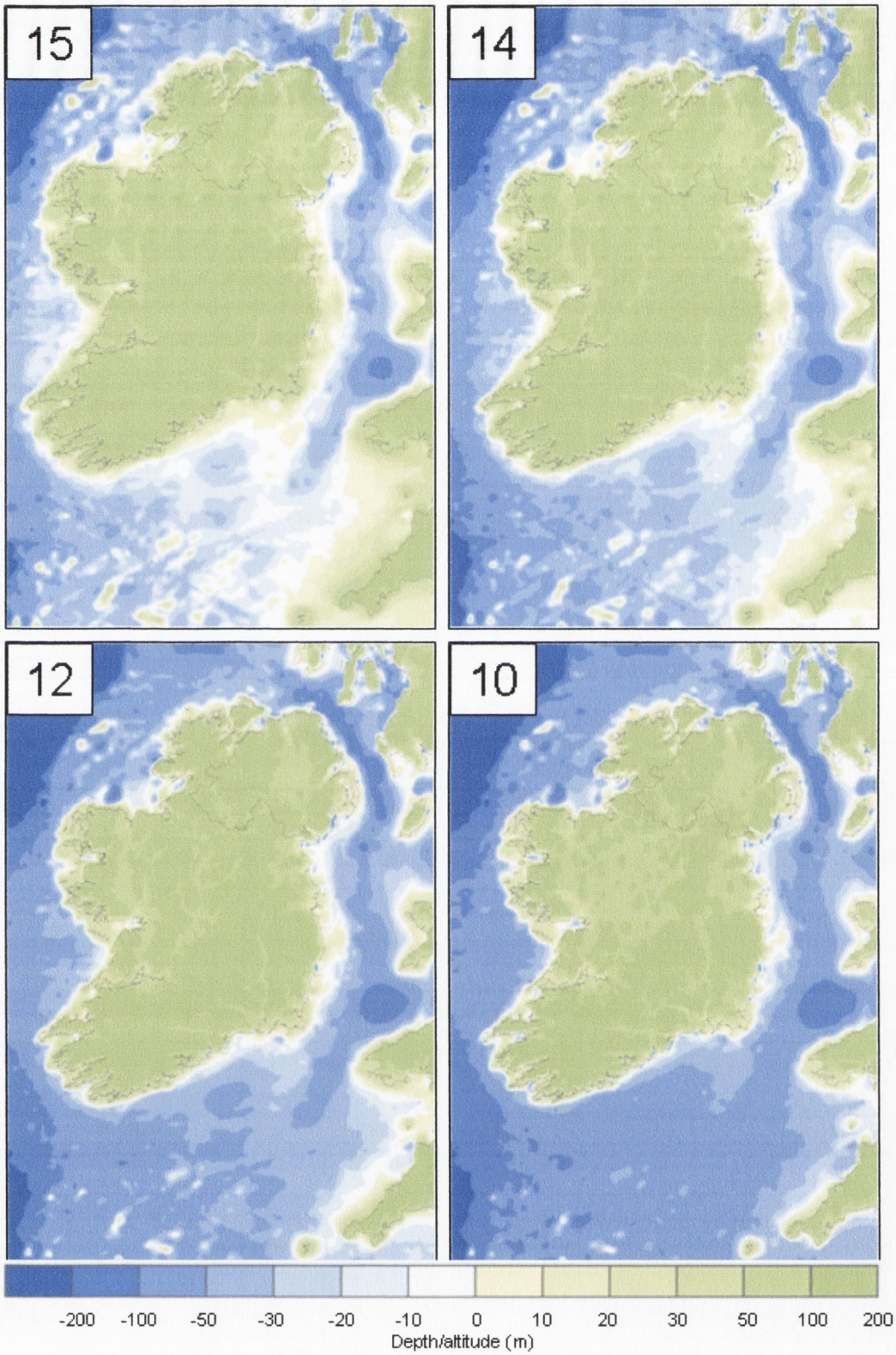


Figure 5.19 Reconstructions of palaeogeography (during the postglacial period) for Ireland and adjacent areas. (Time intervals in cal. kyr BP)

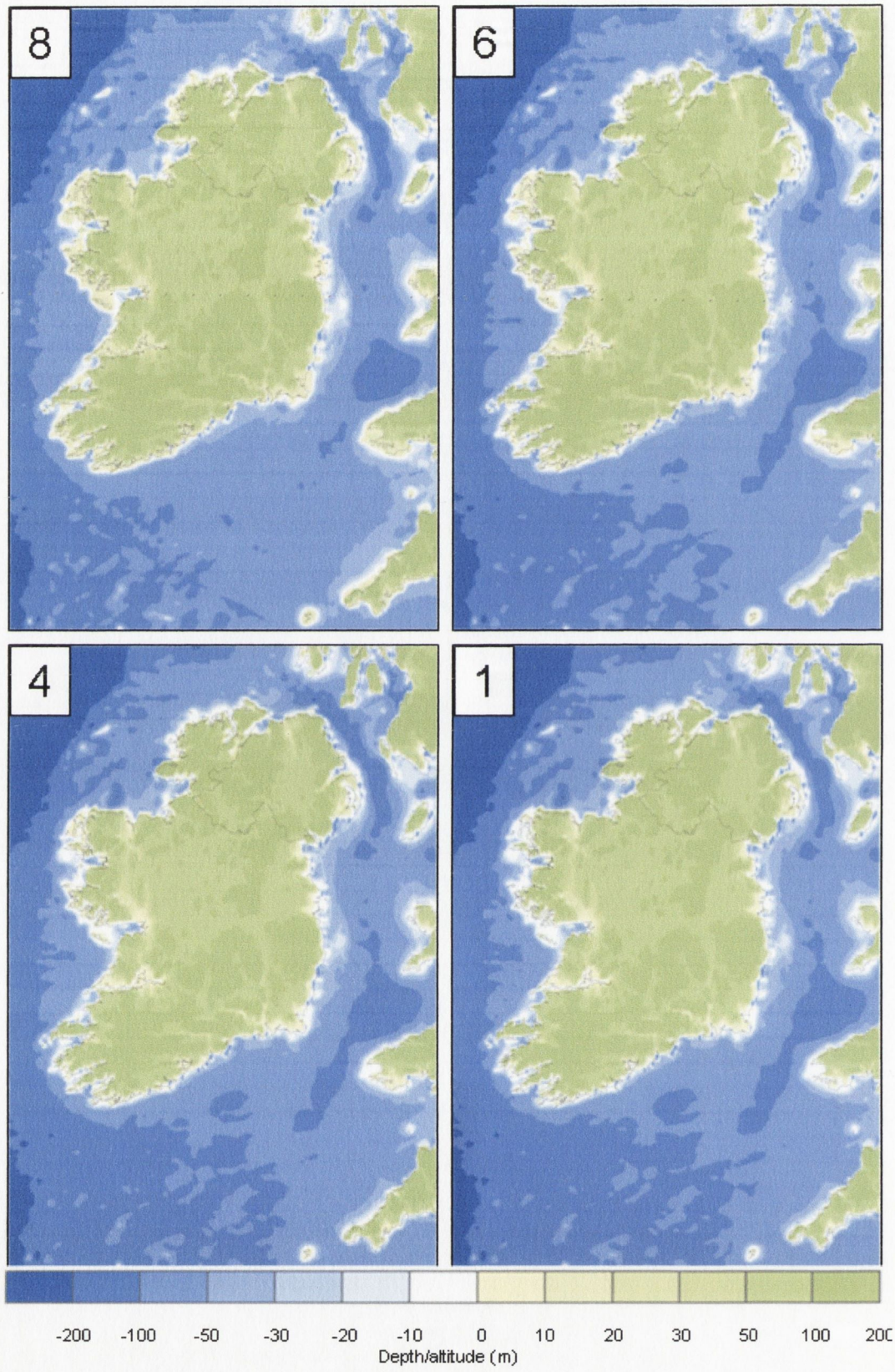


Figure 5.19 Reconstructions of palaeogeography (during the postglacial period) for Ireland and adjacent areas. (Time intervals in cal. kyr BP)

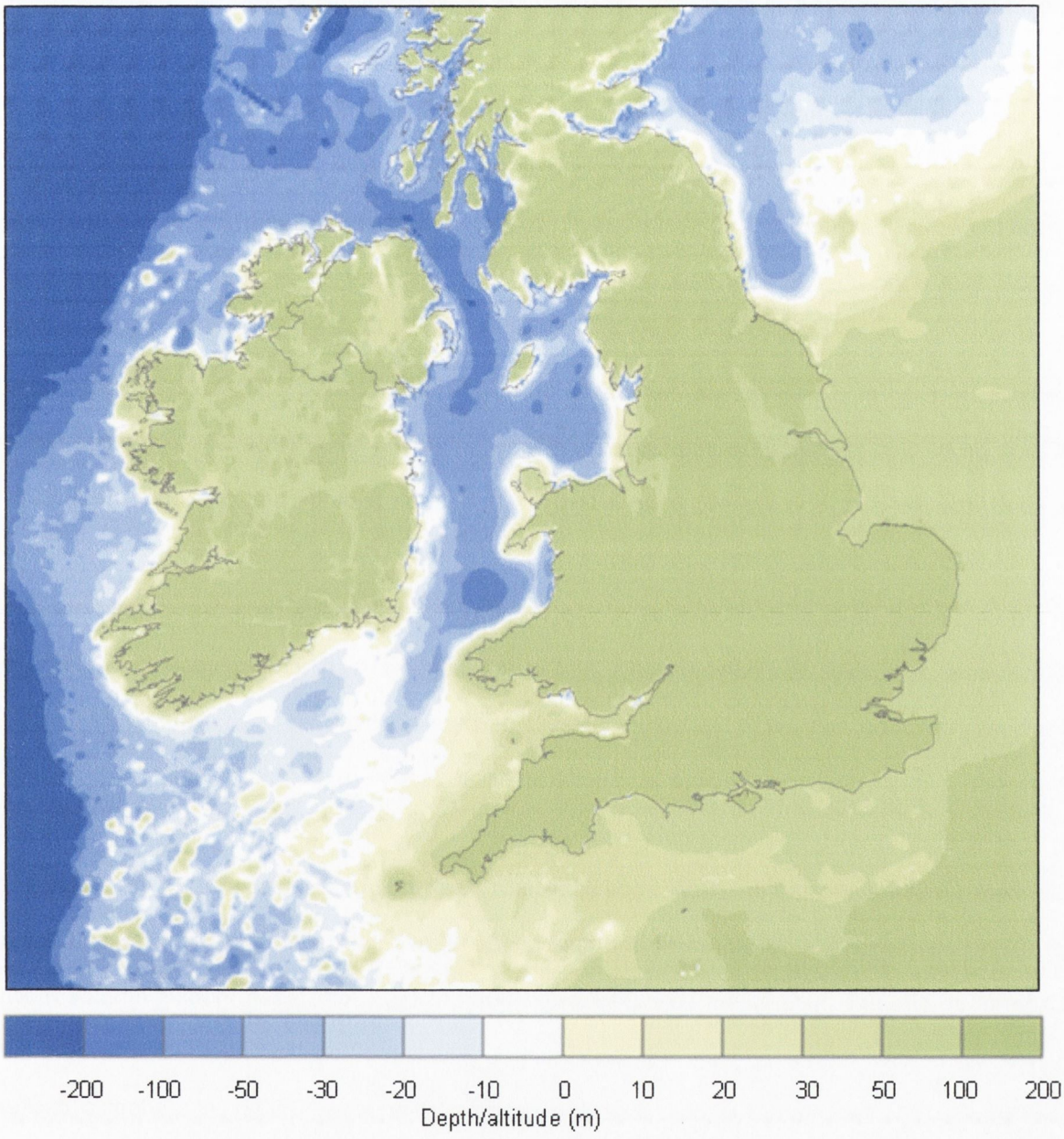


Figure 5.20 Reconstruction of British Isles palaeogeography (20 000 cal.yrs BP)

It is also apparent that at no point since the LGM does this Earth-ice model solution deliver a land bridge between Britain and Ireland. This represents a departure from the findings of a number of previous studies which have envisaged the operation of a land bridge at various times during the last 20 000 years (e.g. Mitchell, 1976, 1986; Devoy, 1985; Preece *et al.*, 1986; Wingfield, 1995; Lambeck, 1996; Peltier *et al.*, 2002). The Celtic Sea does however, become an archipelago and it is noticeable that water depths are commonly less than 20 m up until c. 15 000 years BP. Given uncertainties in the accuracy of model predictions for this locality during the Lateglacial, it is feasible that a land bridge may have existed at some point up until this date. Indeed, ice sheet thickness over most of southern Ireland remains poorly constrained and a reduction in ice thickness in the model will increase the chance of a land bridge occurring. However, whilst Mitchell (1986); pp 64 pictures *'organized oakwoods advancing across a land bridge, carrying those forest animals, the red deer and the wild boar, along with them'*, the findings presented here refute this possibility on two counts. Firstly, conditions for a land bridge between Ireland and Britain are only found before the Bølling interstadial, at a time when tundra conditions (and associated flora and fauna) were pervasive. Secondly, if such a land bridge did exist at all, it most probably would have been very low lying and within the altitudinal limits of marine activity. As such, it would be unlikely to provide a hospitable route for biota to migrate across to (southern) Ireland.

Lambeck (1996) suggested the occurrence of a Celtic Sea land bridge (section 2.3) whilst Peltier *et al.* (2002) envisaged the development of a land bridge across the Celtic Sea, Irish Sea and North Channel at different times in the Late Devensian (Figure 2.9). Both these models employ much thinner ice models than used here and the resultant reduction in isostatic depression favours the persistence of a land bridge in both time and space. Peltier *et al.* (2002) also adopt a thicker lithosphere than that employed here (90 km versus 71 km) and this retards crustal depression yet further, enabling the development of a land bridge across the North Channel. This latter finding is significantly at odds with the model reconstruction presented here since at no point during the postglacial period do water depths between the north of Ireland and southeast Scotland rise above -40 m msl.

5.8 SUMMARY

- Observations of postglacial RSL around Ireland are limited in both time and space and are generally less complete than records found in Britain. However, it has been shown that the newly compiled and quality assessed Irish RSL database does provide an important data set with which to test GIA models for Ireland and the UK.

- Assuming that the influence of lateral Earth structure beneath the British Isles produces a minor impact on predictions of RSL in this region, it has been demonstrated that the Irish ice sheet depicted in Shennan *et al.* (2006a) is largely incompatible with observations of Late Devensian and Holocene RSL in Ireland.

- Employing the same Earth model as Shennan *et al.* (2006a) the revised 'best fit' ice model for Ireland comprises a thick, spatially extensive Irish ice sheet of around 700m over much of north and central Ireland at the LGM. Relatively thick ice out to the continental shelf off the west coast of Scotland and Ireland coupled with thick Irish Sea Ice is required to produce a sufficient fit with observational RSL data from the Late Devensian.

- It is also evident that predictions of RSL are sensitive to the specific rate and timing of deglaciation prior to the LGM. The Irish RSL data strongly favour very rapid deglaciation after 21 000 BP: Early (pre 21 000 BP) deglaciation produces unacceptably low predictions of maximum Late Devensian RSL in the north of Ireland and SW Scotland whilst gradual deglaciation between 20 000 to 16 000 BP deteriorates the fit with observations of Holocene RSL.

- Only a limited fit with available records of RSL can be obtained by altering the Irish ice sheet component of BIM-1. Significant discrepancies with RSL data could not be resolved, particularly those in relation to the geographical position of the 0 m Holocene isobase. However, it was possible to solve this problem by making a minor reduction to the value of upper mantle viscosity employed in the Earth model. Despite the significant changes in RSL predictions brought about through making this adjustment to the Earth model, the results still

conclusively demonstrate that the Irish component of BIM-1 is not compatible with the RSL data.

- The primary conclusion from the ice modelling analyses – the need for a thinner Irish ice Sheet soon after the LGM and throughout the deglacial phase - remains a robust requirement of the RSL data regardless of the adopted viscosity model.

- No evidence is found for the operation of a land bridge between Ireland and Britain at any point since the LGM. The Celtic Sea does, however, become an archipelago characterized by very shallow (<20 m) water depths. Relatively minor reductions in model ice thickness for southern Ireland may well result in a land bridge forming across the Celtic Sea although such a feature is likely to have been ephemeral and very low lying.

CHAPTER SIX

Field Data

6:1 THE SHANNON ESTUARY

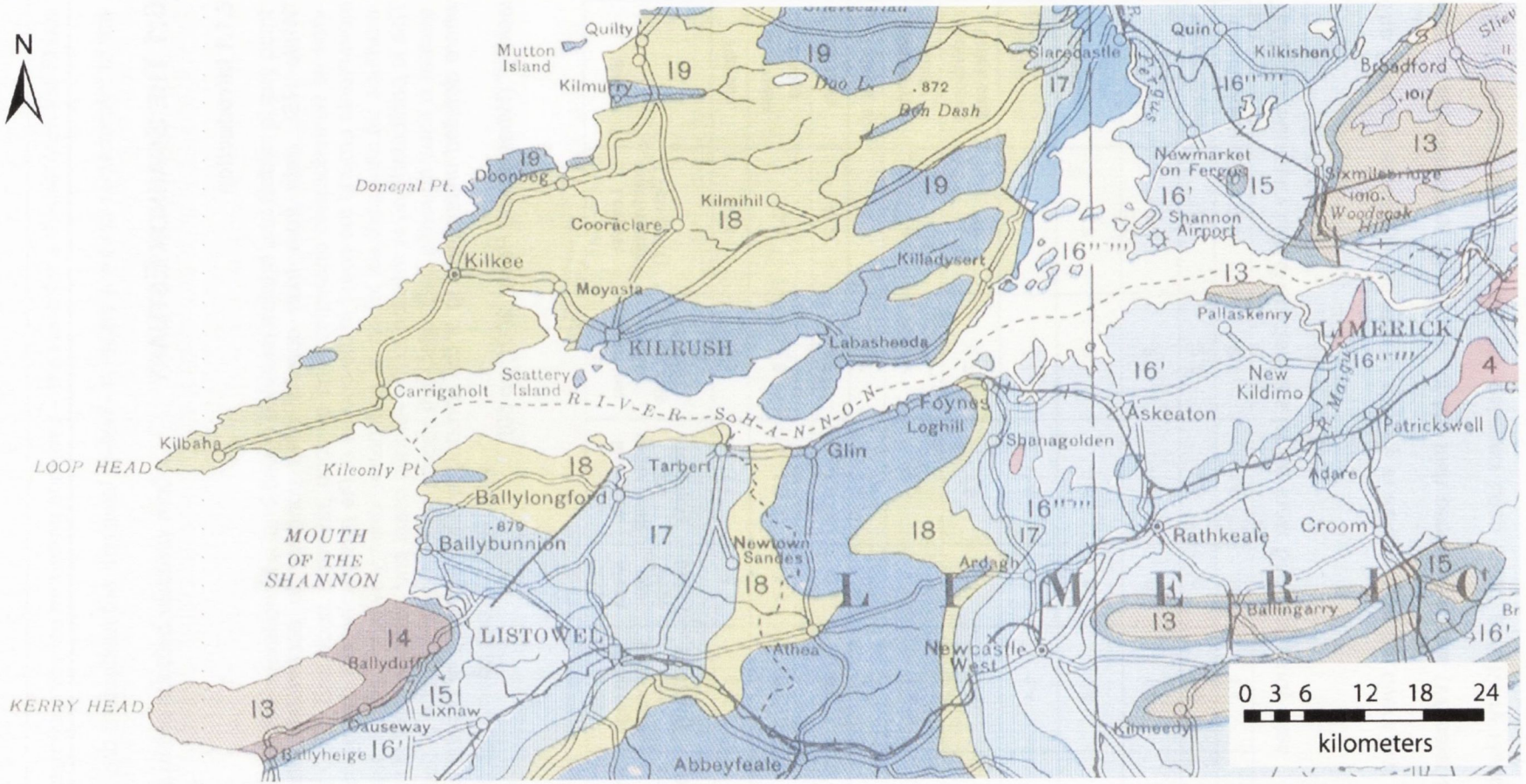
6.1.1 Introduction

"While RSL data from Irish coastal sites continue to accumulate, the history of sea-level movements around the coast in the course of the postglacial period remains patchy (Carter, 1990). This is particularly so in relation to the Atlantic coast and perhaps most significant in the Shannon, where detailed understanding of the chronology and sequence of coastal and sea-level change is lacking" (Wheeler and Healy, 2001; pp 50).

Site	Lowest astronomical Tide (LAT)	Mean low water spring tide (MLWST)	Mean low water neap tide (MLWNT)	Mean sea level (MSL)	Mean high water neap tide (MHWNT)	Mean high water spring tide (MHWST)	Highest astronomical Tide (HAT)
Rahone	-3.1	-2.3	-1.1	-0.3	0.7	1.9	2.5
Blackweir Bridge	-3.2	-2.5	-1.3	-0.4	0.7	2	2.6
Moyasta	-3.2	-2.5	-1.3	-0.4	0.7	2	2.6
Islandmagrath	-3.1	-2.7	-1.2	-0.2	0.9	2.2	2.8
Ringmoylan Quay	-3.3	-2.7	-1.2	0.1	1.4	2.8	3.4
Ballinacurra Creek east	-3.3	-2.7	-1.6	0	1.4	2.9	3.5
Ballinacurra Creek west	-3.3	-2.7	-1.6	0	1.4	2.9	3.5
Coonagh South	-3.3	-2.7	-1.6	0	1.4	2.9	3.5
Coonagh West	-3.3	-2.7	-1.6	0	1.4	2.9	3.5
Meelick Creek	-3.3	-2.7	-1.6	0	1.4	2.9	3.5
Clonmacken	-3.3	-2.7	-1.6	0	1.4	2.9	3.5

Table 6.1 Summary tidal characteristics for the inner Shannon study sites

The Shannon estuary is the largest estuary in Ireland, with the rivers that flow into it draining an area of approximately 15,700 km^2 of the Irish central lowlands (Wheeler and Healy, 2001). The Atlantic Ocean enters the estuary between Loop Head and Kerry Head at the mouth of the



4 Basalt

11 Quartzite

13 Old Red Sandstone

14 Kiltorcan Old Red Sandstone

15 Lower Avonian Shales and sandstones

16 Carboniferous Limestone

17 Upper Avonian Shales and Sandstones

18 Millstone Grit and Flagstone

19 Coal Measures

Figure 6.1 Geology of the Shannon region. (Williams, 1979)



Figure 6.2 The inner Shannon estuary in Henry Pelham's map of Clare (1787). (Pheonix maps)

Shannon although tidal influence extends almost 100 km to the east, upstream of Limerick City. The Shannon itself is macrotidal with tidal amplitudes increasing as one travels east (Table 6.1). The tidal range in the inner estuary at Limerick exceeds 5 m, representing one of the largest tidal ranges found round the Irish coast (Admiralty Charts, 1997). However, despite the rivers flowing into the Shannon estuary draining almost 1/3 of Ireland's land area, freshwater input into the estuary is actually low in relation to the size of the estuary. Tidal waters account for 70% of the volume of upper estuary (Wheeler and Healy, 2001), enabling the development of salt-marshes far up the estuary (Curtis and Sheehy-Skeffington, 1998). Depths within the estuary range from c.37 m at the mouth to less than 5 m in the channel at Limerick City, although dredging has artificially deepened the channel in places.

Reference to Figure 6.1 shows that the estuary cuts through a diversity of rock types. In the lower Shannon region, Millstone Grit and Flagstone dominates whilst to the east, the older Visean and Tournaisian limestones are the most common rock outcrop (Sleeman and Pracht, 1999). Carboniferous limestone is widespread in the upper part of the estuary but its susceptibility to both chemical and physical weathering has directly led to widening of the estuary and the surrounding low relief (O'Sullivan, 2001). Estuary morphology has also been influenced by erosion and deposition associated with the Devensian ice mass (e.g. McCabe, 1985; Warren. 1985). Glacial till is extensive throughout the region, particularly in the upper Shannon where it is frequently overlain by estuarine alluvium and wetland sediments. Much of the upper estuary is also flanked by drumlins.

The earliest settlers in the Shannon estuary region were Mesolithic (c.9000 - 6000 BP) hunter gathers and a diversity of archaeological finds testifies to a strong human-environment interaction ever since this period (O'Sullivan, 2001). Indeed, anthropogenic influences appear to have played a significant role in the evolution of the Shannon estuary lowlands and it is almost certain that local wetland succession patterns were affected by early rudimentary agricultural activities producing consequent morphological responses within the estuary system (Healy and Hickey, 2002). The key change in the historic period has been the embanking and draining of most of the estuary marshes over the past 3-4 centuries, with reclamation and enclosure of at least 6,500 ha of the Shannon estuary lowlands for agriculture and other purposes having taken place (Healy and Hickey, 2002). The present estuary is therefore considerably narrower than it was in the past (Figure 6.2).

Index Point Tier	Region	Site	Site Location (Latitude)	Site Location (Longitude)	Labcode	C-14 Age BP	% C-13	Calibrated Age BP (2σ)	Midpoint	±	Altitude m OD(Belfast)	Material	Microfossil Evidence	Reference Water Level	Indicative Range (m)	Tidal port 1	HAT	MHWST	MHWNT	MTL/MSL	MLWNT	MLWST	LAT	Indicative Meaning	RSL (Relative to msl)	Error Term (m)	Reference
2	7	Clonderalaw Bay	52.632	-9.303	GrN-20144	250±30		0-428	214	214	-1.3	Fishtrap		MSL	50	Tarbert Island		2.0	0.8	-0.2	-1.3	-2.5		0	-1.3	?	O'Sullivan (2001)
2	8	Deel Estuary	52.624	-8.993	GrN-21931	640±20		558-663	610.5	53	-1.2	Fishtrap		MSL	50	Mellon Point		2.8	1.4	0.1	-1.2	-2.7		0	-1.2	?	O'Sullivan (2001)
2	8	Deel Estuary	52.625	-8.994	GrN-20975	740±15		667-689	678	11	-1.2	Fishtrap		MSL	50	Mellon Point		2.8	1.4	0.1	-1.2	-2.7		0	-1.2	?	O'Sullivan (2001)
2	8	Fergus Estuary	52.781	-8.945	GrN-20139	1495±35		1306-1512	1409	103	-1.2	Fishtrap		MSL	50	Mellon Point		2.8	1.4	0.1	-1.2	-2.7		0	-1.2	?	O'Sullivan (2001)

Limiting Date Tier	Region	Site	Site Location (Latitude)	Site Location (Longitude)	Labcode	C-14 Age BP	% C-13	Calibrated Age BP (2σ)	Midpoint	±	Altitude m OD(Belfast)	Material	Microfossil Evidence	Extent of Marine Influence	Tidal port 1	Tidal port 2	HAT	MHWST	MHWNT	MTL/MSL	MLWNT	MLWST	LAT	Indicative Meaning	Max/Min Height of msl	Error Term (m)	Reference
1	7	Bran Lough, (Claire)	52.80	-9.57	Beta-62536	360±60		305-507	406	101	0.95	Peat		MHWST	Galway	Carrigholt		1.9	0.7	-0.3	-1.1	-2.3		2.17	-1.22	0.3	Devoy et al, (1996)
1	7	Bran Lough, (Claire)	52.80	-9.57	Beta-62537	530±60		497-654	575.5	79	0.77	Peat		MHWST	Galway	Carrigholt		1.9	0.7	-0.3	-1.1	-2.3		2.17	-1.4	0.3	Devoy et al, (1996)
1	7	Bran Lough, (Claire)	52.80	-9.57	Beta-62538	570±60	-25	514-657	585.5	72	0.8	Sandy Peat		MHWST	Galway	Carrigholt		1.9	0.7	-0.3	-1.1	-2.3		2.17	-1.37	0.3	Devoy et al, (1996)
1	7	Bran Lough, (Claire)	52.80	-9.57	Beta-62539	2260±80	-25	2009-2468	2238.5	230	-0.15	Sandy Peat		MHWST	Galway	Carrigholt		1.9	0.7	-0.3	-1.1	-2.3		2.17	-2.32	0.3	Devoy et al, (1996)
1	7	Rinvella Bay (Clare)	52.584	-9.734	UB-2094	3935±80		4097-4781	4439	342	2	Wood Peat		MHWST	Kilrush			2	0.7	-0.4	-1.3	-2.5		2.40	0	1.0	Pearson, (1979)
1	7	Rinvella Bay (Clare)	52.584	-9.734	UB-2093	4185±50		4573-4845	4709	136	2	Wood		MHWST	Kilrush			2	0.7	-0.4	-1.3	-2.5		2.40	0	1.0	Pearson, (1979)
1	9	Coonagh West	52.669	-8.700	Beta - 205301	3820±50	-28	4089-4408	4248.5	160	0.45	Wood		MHWST	Limerick Dock	Mellon Point		2.9	1.4	0.0	-1.6	-2.7		2.87	-2.42	0.2	TVAS Report, (2005)
2	9	Coonagh West	52.669	-8.700	Beta - 205299	1970±40	-26.7	1825-1997	1911	86	1.447	Charred material		MHWST	Limerick Dock	Mellon Point		2.9	1.4	0.0	-1.6	-2.7		2.88	-1.4	?	TVAS Report, (2005)
2	9	Coonagh West	52.669	-8.700	Beta - 205300	3260±40	-29	3390-3574	3482	92	0.947	Wood		MHWST	Limerick Dock	Mellon Point		2.9	1.4	0.0	-1.6	-2.7		2.88	-1.9	?	TVAS Report, (2005)
2	9	Carrigdirty Rock	52.667	-8.779	GrN-20976	3330±25		3479-3633	3556	77	-0.465	House site		HAT	Mellon Point		3.4	2.8	1.4	0.1	-1.2	-2.7		3.33	-3.8	?	O'Sullivan (2001)
2	9	Meelick Rocks	52.67	-8.71	GrN-21930	4160 ± 20		4588-4825	4706.5	119	-1	Tree trunk		MHWST	Mellon Point			2.8	1.4	0.1	-1.2	-2.7		2.73	-3.7	?	O'Sullivan (2001)
2	9	Carrigdirty Rock	52.667	-8.779	Beta-102086	4710 ± 60		5319-5583	5451	132	-1.59	skull fragment		MHWST	Mellon Point			2.8	1.4	0.1	-1.2	-2.7		2.73	-4.3	?	O'Sullivan (2001)
2	9	Carrigdirty Rock	52.667	-8.779	Beta-102087	4820±50		5333-5654	5493.5	161	-1.6	Alder basket		MHWST	Mellon Point			2.8	1.4	0.1	-1.2	-2.7		2.73	-4.3	?	O'Sullivan (2001)
2	9	Carrigdirty Rock	52.667	-8.779	GrN-21936	5820 ± 40		6503-6730	6616.5	114	-2.38	Wooden plank		MHWST	Mellon Point			2.8	1.4	0.1	-1.2	-2.7		2.73	-5.1	?	O'Sullivan (2001)
2	9	Meelick Rocks	52.67	-8.71	GrN-21929	6240 ± 25		7027-7253	7140	113	-2.76	Tree trunk		MHWST	Mellon Point			2.8	1.4	0.1	-1.2	-2.7		2.73	-5.5	?	O'Sullivan (2001)

Table 6.2 Existing sea level data from the Shannon estuary (database regions 7, 8 and 9)

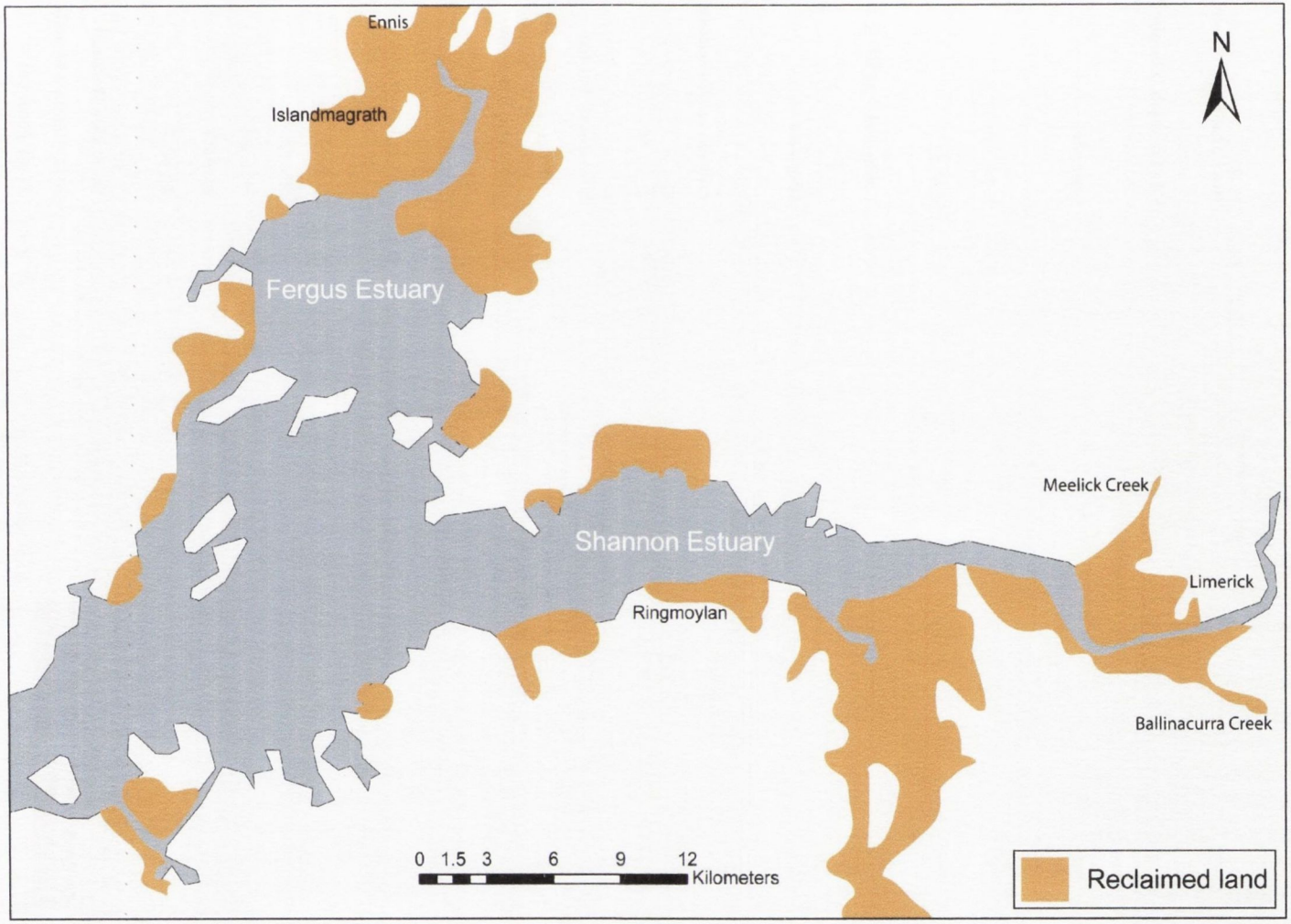


Figure 6.3 Reclamation and anthropogenic influence on the Fergus and inner Shannon Estuary. (Adapted from Healey and Hickey, 2002)

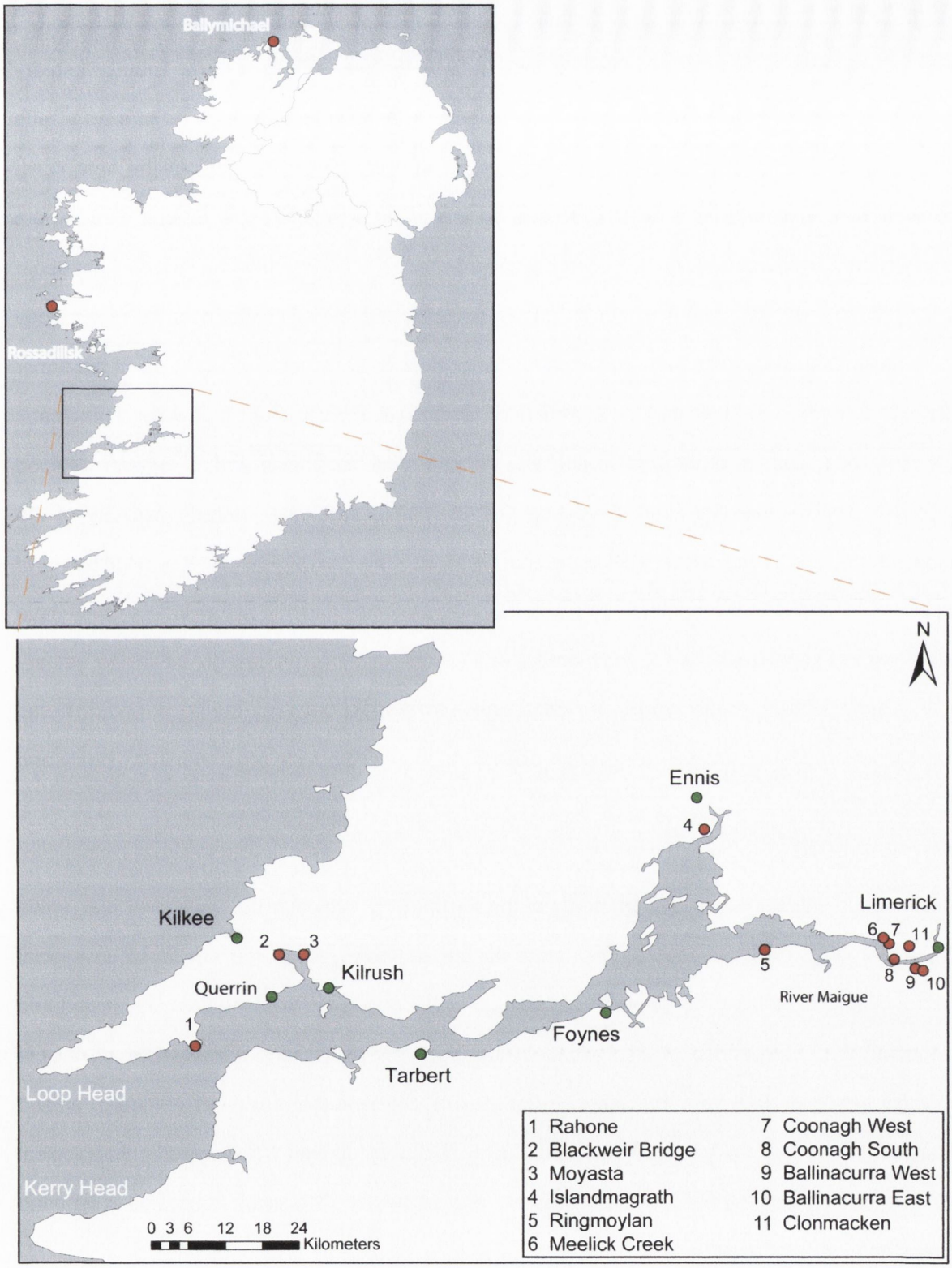


Figure 6.4 The Shannon estuary study sites along with the location of other sites mentioned in the text

It is likely that most of the reclamation along the Shannon within the inner estuary was achieved in the 19th and 20th century, although reclamation in some parts may have begun as early as the 17th century. However, in contrast to other regions (such as the Severn estuary, the Thames estuary and Romney marsh), there is only limited evidence for reclamations dating backing to the 10th and 11th centuries AD (Healy, 2002). Figure 6.3 reveals the geographical extent of the land claim activities with reclamation particularly focused on the areas adjacent to the Fergus estuary, the River Maigue and Limerick City. The other significant change to the estuary has been bought about by the development of Limerick as a port, leading to increasing human control over the morphology of the navigable estuary channel. The construction of river embankments along much of the upper Shannon estuary has maintained the current river course with the channelling of water during peak flow promoting scouring of the bed. In addition, the use of embankments and revetments as flood protection devices has significantly reduced the dissipative and water storage capacity of the adjoining low-lying areas, which were once tidal wetlands. The corollary of this is a modification of the estuarine physical environment along with its hydrodynamics and this will almost certainly have had important consequences for sedimentary and morphodynamic processes within the estuary itself. These include modifications in water and sediment circulation, transport patterns and changes in the tidal prism, with associated alterations in ecology, hydrology and RSL (Healy and Hickey, 2002).

The Shannon estuary lies on the periphery of the former BIIS. Although ice cover over this region will have imparted a glacio-isostatic signature on the sea-level record, the effects of postglacial eustatic sea-level rise are thought to have been dominant during much of the Late Devensian and throughout the entire period of the Holocene (Lambeck, 1996). Despite the absence of reliable empirical sea level data from the Shannon (chapter 4), limits on the maximum altitude of palaeo mean sea level may be inferred. This information is mainly derived from dated archaeological evidence collected during the Shannon estuary inter-tidal survey (O'Sullivan, 2001) and is included in the sea-level database (Table 6.2).

Eight study sites from the Shannon estuary that were believed to hold information on past sea-level change were identified and are presented in the following section. In addition to those sites shown in Figure 6.4, a number of other sites were also visited. However, these did not contain sequences

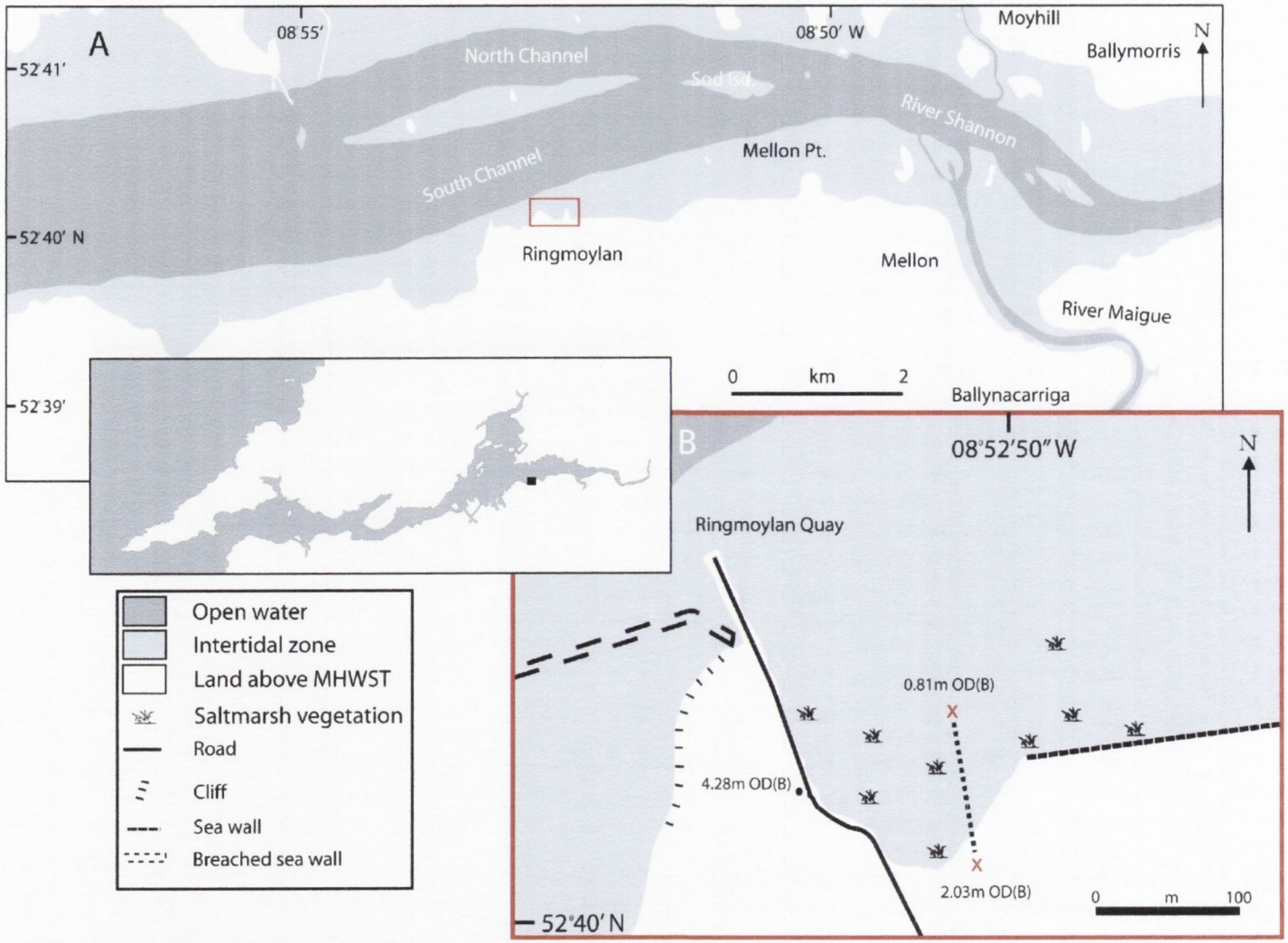


Figure 6.5 Site map showing **A** the location of Ringmoylan Quay in relation to other sites in the inner Shannon; **B** the location of the surface foraminiferal transect



Figure 6.6 Unconstrained cluster analysis based on unweighted Euclidean distance of foraminiferal death assemblages from the salt-marsh at Ringmoylan Quay. Only samples with counts greater than 40 individuals and species which reach 5% of the total sum are included



Plate 6.1 The contemporary salt-marsh found at Ringmoylan Quay (looking southeast)

Photo taken June 2006

that were suitable for the study of (Holocene) RSL change and thus have not been discussed here in detail. Instead, lithostratigraphic evidence from these sites has been included in Appendix IV.

6.1.2 Contemporary Foraminifera in the Inner Shannon

Ringmoylan Quay

Contemporary foraminifera have been collected from the salt-marsh at Ringmoylan Quay on the south shore of the inner Shannon estuary (Figure 6.5a, b). As previously mentioned, a large proportion of the inner Shannon coastline is characterized by sea walls and embankments and the reclaimed land contained within them is often extensively grazed. The stretch of coastline on which Ringmoylan Quay is located is no different, and the coastline to both the east and west of the marsh is artificially maintained (Figure 6.3 and Figure 6.5b). Significantly however, the upper salt-marsh at Ringmoylan Quay appears undisturbed and the marsh transect is not backed by a sea wall. This meant continuous sampling could be achieved across the high marsh-upland transition. Erosion of the salt-marsh is taking place and small-scale cliffs between the vegetated marsh and un-vegetated mudflat have emerged in places. Vertical zonation of marsh vascular flora is evident although is not strongly defined (Plate 6.1).

Figure 6.6 displays the surface distribution of foraminifera across the salt-marsh at Ringmoylan Quay. All samples yielded (dead) foraminifera counts in excess of 40 with abundance generally inversely related to marsh surface elevation. Species diversity was relatively high, with 17 different species encountered across the marsh and upper mudflat. Cluster analysis separates the foraminiferal data into two broad zones. In zone I (2.2 – 1.9 m OD(B)), samples are dominated by the agglutinated species *Haplophragmoides sp.*, *Jadammina macrescens* and *Miliammina fusca* and together, these three species comprise over 80% of sample population found on the upper marsh surface. Foraminifera abundance decreases with elevation throughout this zone and no foraminifera are found higher than 2.2 m OD(B) (60 cm below the height of MHWST). In zone II (1.9 m – 0.8 m OD(B)), calcareous taxa are more abundant and commonly comprise greater than 60% of the foraminifera population. The most common calcareous species are *Brizalina sp.* and *Fursenkoina sp.* and between them, they usually constitute more than 30% of sample population. The agglutinated species *Trochammina ochracea* is also well represented in this lower zone, with

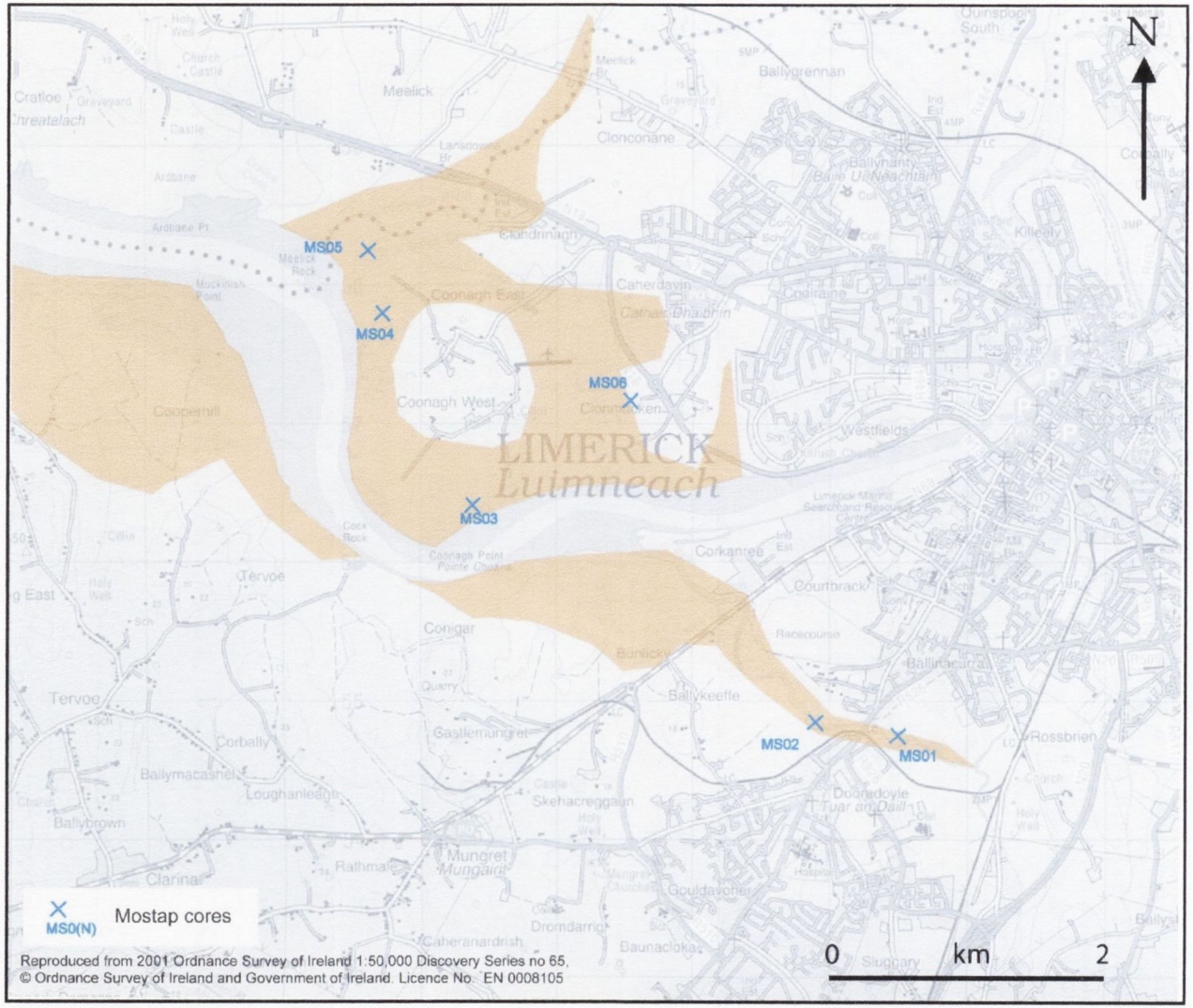


Figure 6.7 Reclamation and anthropogenic influence on the inner Shannon estuary (after Healey and Hickey, 2002). The positions of boreholes MS01 to MS06 are also shown

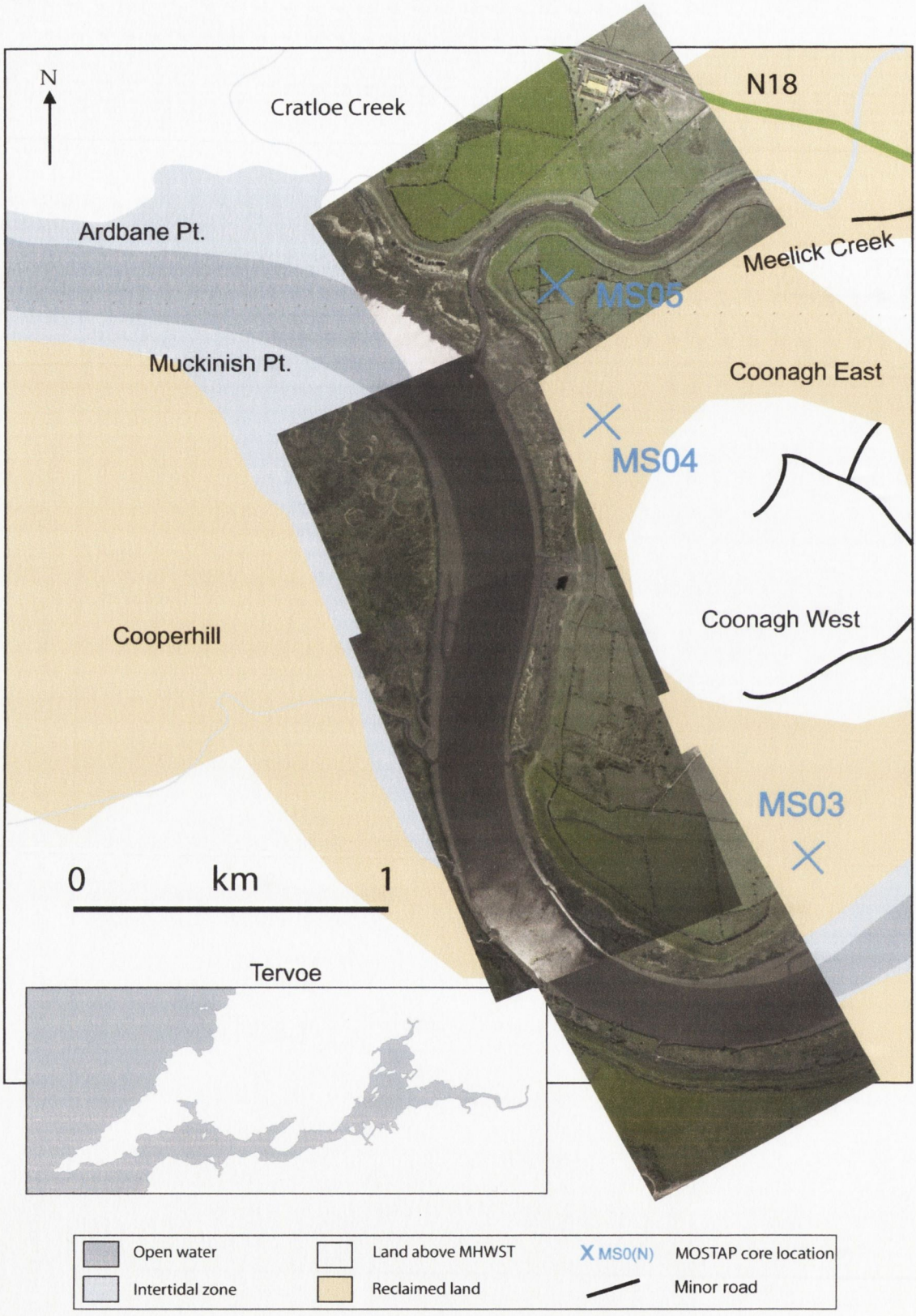


Figure 6.8 Location map of core sites MS02, MS03 and MS04 in the inner Shannon estuary



Plate 6.2 Location of core MS03 extracted from reclaimed land at Coonagh south (looking south)

Photo taken April 2005

frequencies commonly exceeding 20% of the sample population. As with the upper foraminifera zone, foraminifera abundance generally increases as marsh surface elevation decreases.

6.1.3 Study Sites From the Inner Shannon

All six MOSTAP cores collected form sites in the Inner Shannon Estuary (Coonagh, Ballinacurra, Meelick and Clonmacken) were commissioned and extracted by the National Roads Authority (although the litho-, bio- and chronostratigraphical analyses presented here were undertaken by the author). Detailed stratigraphic information of the local area was required by the NRA to aid identification of a suitable route for the proposed south Limerick bypass which requires the construction of a road tunnel under the estuary. The line of the south Limerick bypass crosses the Shannon estuary a short distance to the west of Limerick city and for the most part, the route lies on the flood plain of the Shannon Estuary. The surface of the flood plain is commonly less than +2 m OD(B) and until the recent embankment of the river in the 19th century, was subject to regular tidal inundation. Figures 6.7 and Figure 6.8 show clear evidence of this anthropogenic modification of the coastline with sea walls and embankments flanking much of the estuary. All of the cores used to provide material for both dating and biostratigraphical analysis are located on areas of reclaimed land which are now primarily used for pastoral farming (Plate 6.2). Although this part of the inner Shannon is still subjected to macro tidal conditions, no salt-marshes are reported this far up the estuary by Curtis and Sheehy-Skeffington (1998). Instead, the banks of the estuary are characterized by extensive reed bed communities.

From the above site description, it is clear that the inner Shannon has a number of characteristics which are known to complicate detailed study of past changes in sea level (section 3:4:1): The Shannon estuary is macrotidal and has also been subject to a substantial degree of anthropogenic alteration. This will have affected the tidal characteristics of the estuary to some degree. However, several factors combine to provide a strong argument for the NRA cores from the inner Shannon to be considered in detail.

Firstly, preliminary coring commissioned by the NRA (section 6.1.3(i)) shows that peat horizons exist at depth in the inner estuary. Many of these peat units are evident well below -10 m OD(B) and as such, are likely to hold a record of environmental change spanning most if not all of the

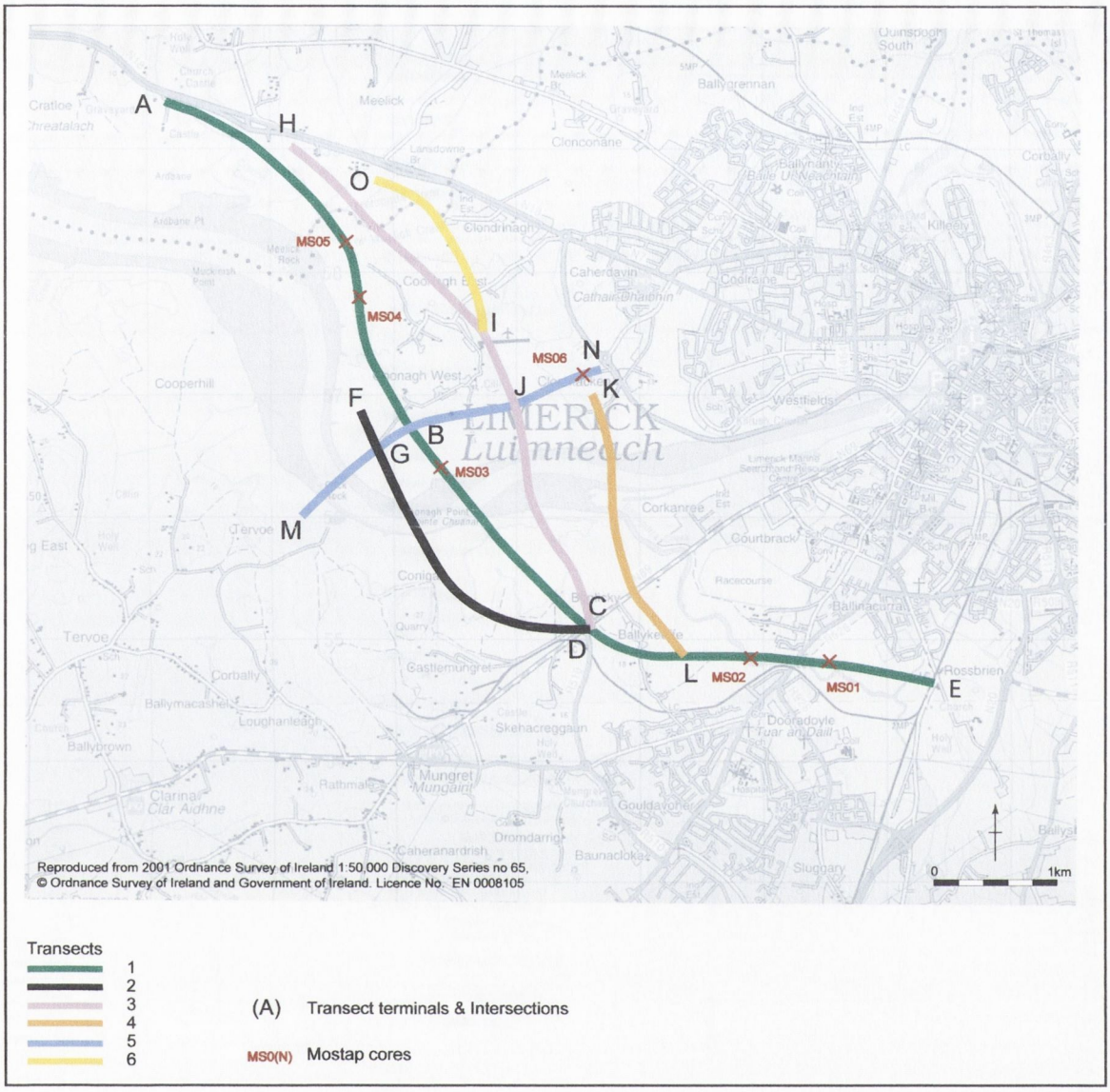


Figure 6.9 Location of the pre-existing NRA borehole transects. MOSTAP core locations are also shown. (Adapted from Carter and Barton, 2005)

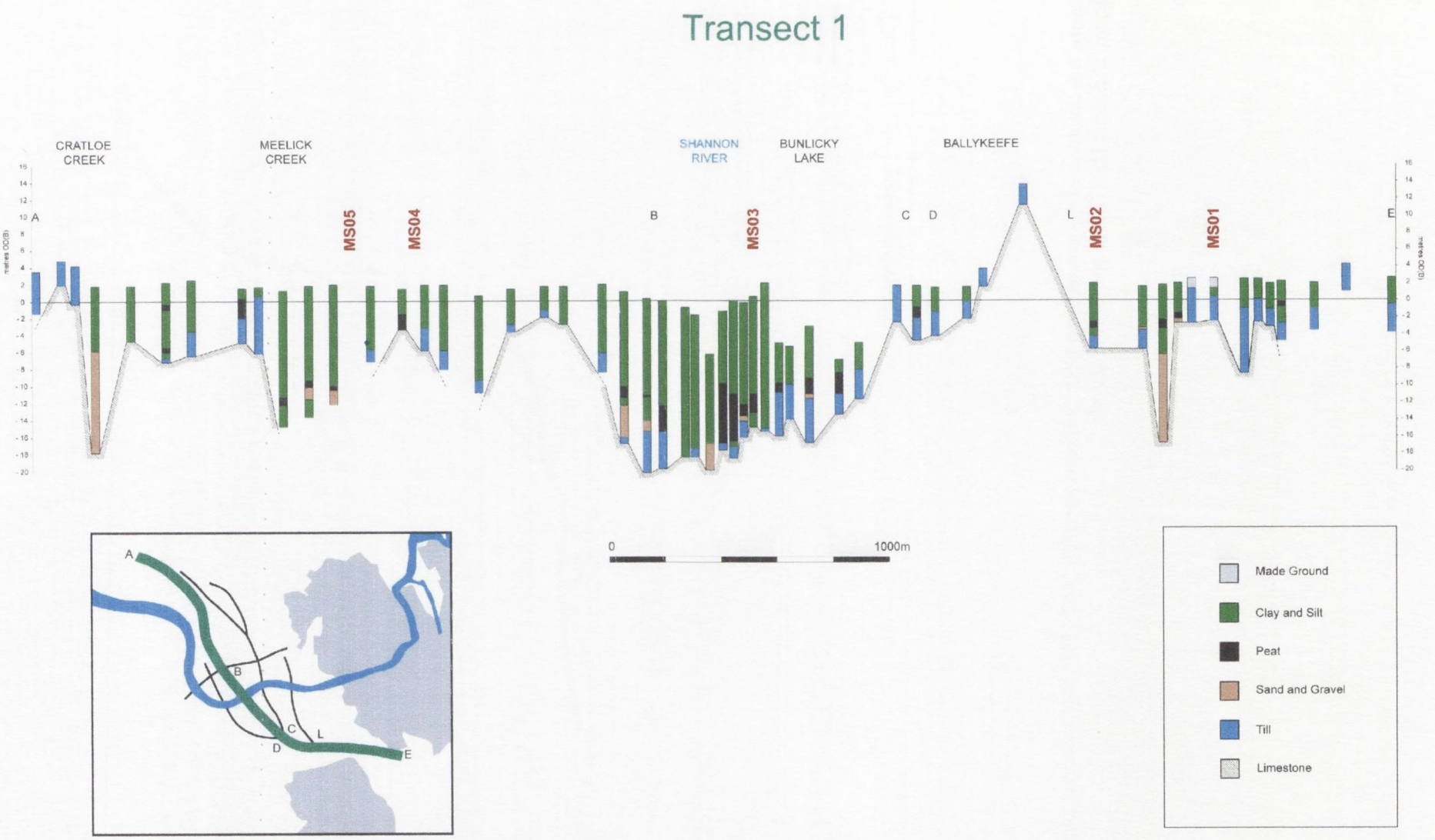


Figure 6.10a Simplified stratigraphy of NRA boreholes taken along transect 1. (Adapted from Carter and Barton, 2005)

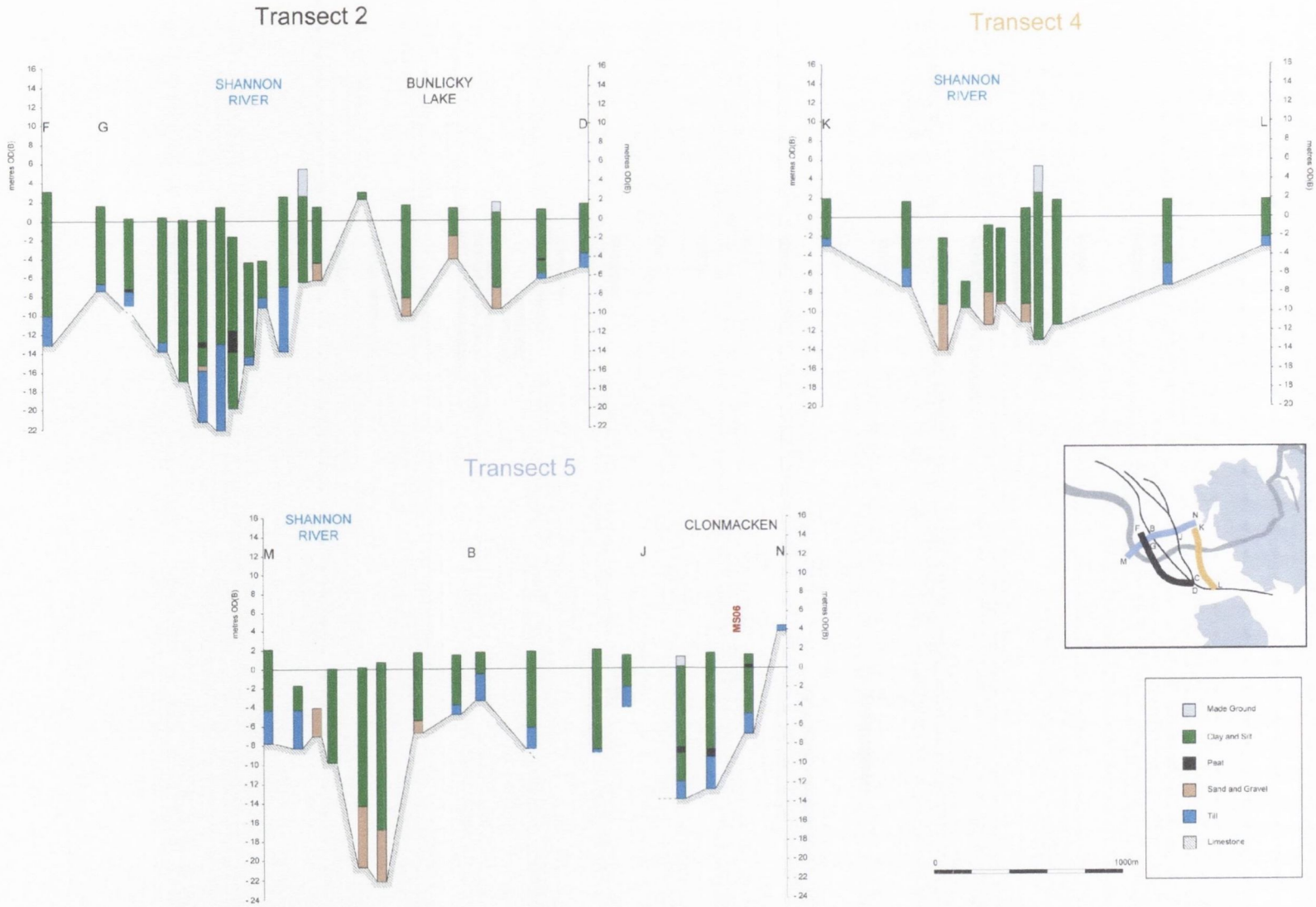


Figure 6.10b Simplified stratigraphy of NRA boreholes taken along transects 2,4,5. (Adapted from Carter and Barton, 2005)

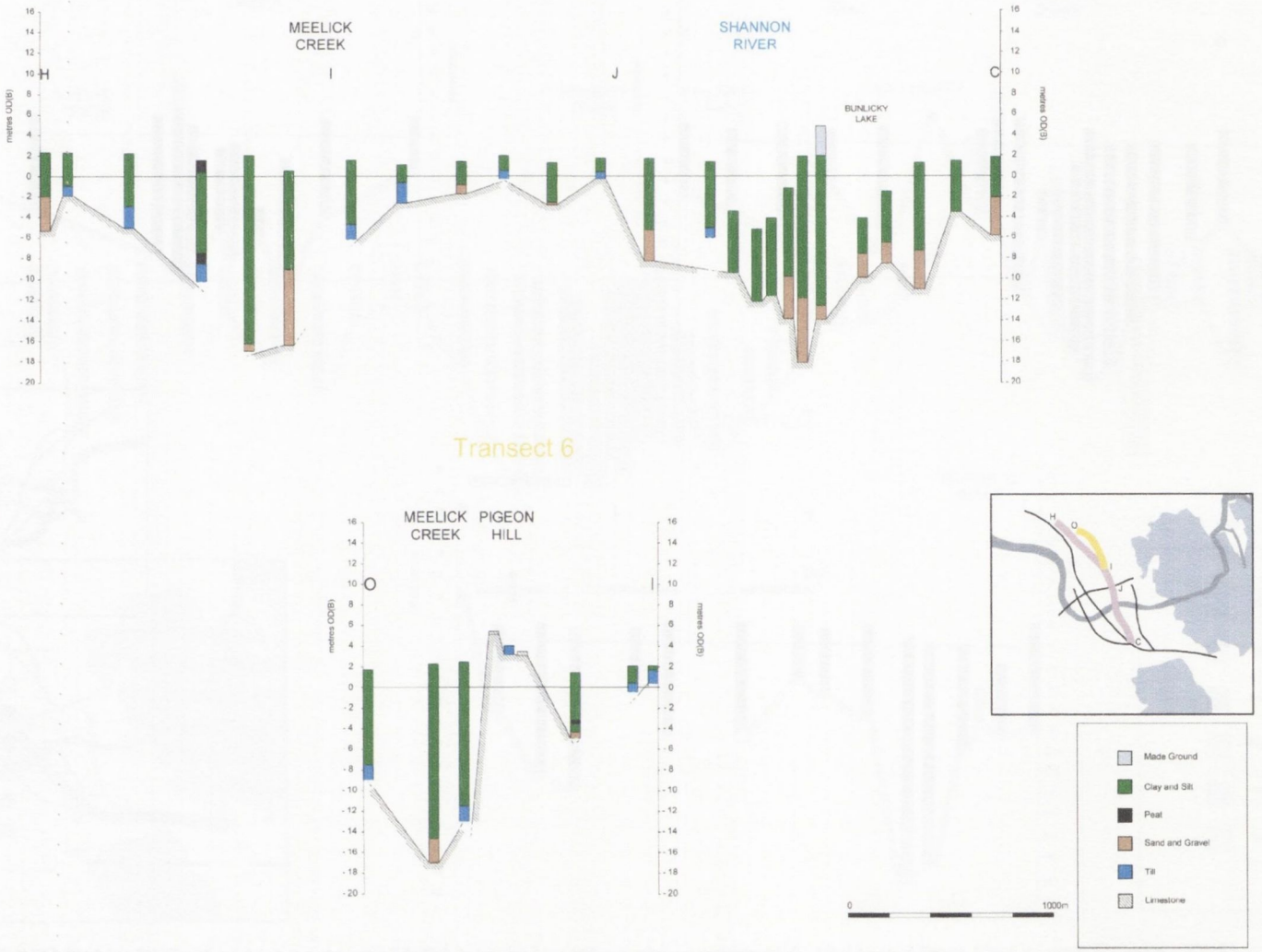


Figure 6.10c Simplified stratigraphy of NRA boreholes taken along transects 3 and 6. (Adapted from Carter and Barton, 2005)

Holocene. Stratigraphic units at this depth are far below the reach of the coring equipment available to the author. Moreover, it is generally the case that older dates are more powerful constraints on glacial rebound models (Chapter 5). In fact in many instances, late Holocene SLIP data is probably of less use as a model constraint than early Holocene limiting data: Even if the deep organic units highlighted by the preliminary NRA coring proved to be wholly terrestrial in origin or erosive in nature, they could still provide important information if dated.

Secondly, other studies have shown that a coherent signal of past RSL change may be extracted from macrotidal estuaries that have undergone extensive land reclamation (e.g. Haslett *et al.*, 2001). This has been achieved through the adoption of the sea level index point methodology coupled with detailed analysis of contemporary micro fauna found within the local environment.

Finally, many of the salt-marsh sites visited in both the inner and outer Shannon are underlain by wholly monogenic sequences which are largely unsuitable for detailed analysis of changes in (Holocene) RSL (Appendix IV). In a regional context, the west coast of Ireland is characterized by a paucity of sea level information (section 4.4.1) and previous studies show this is probably related to the scarcity of suitable sedimentary sequences along this coastline. The decision to carry out detailed stratigraphic analysis of the inner Shannon cores must be viewed in light of these circumstances.

6.1.3(i) Existing National Roads Authority borehole logs

Over the past few years a large number of borehole transects have been commissioned and logged by the NRA. These are sourced from a wide geographical area and cover several square kilometres of the inner Shannon floodplain. This additional borehole data has been made available and enables the results of the detailed MOSTAP (site specific) core logs (section 6:1:3) to be placed into landscape context.

The location of these transects are shown in Figure 6.9 and simplified core stratigraphies are displayed in Figure 6.10abc. (The locations of the MOSTAP cores analyzed in sections 6.1.3(ii-iv) are also shown). From these figures, it is evident that a significant thickness of unconsolidated organic and minerogenic material exists in the inner Shannon estuary. Most of the cores terminate

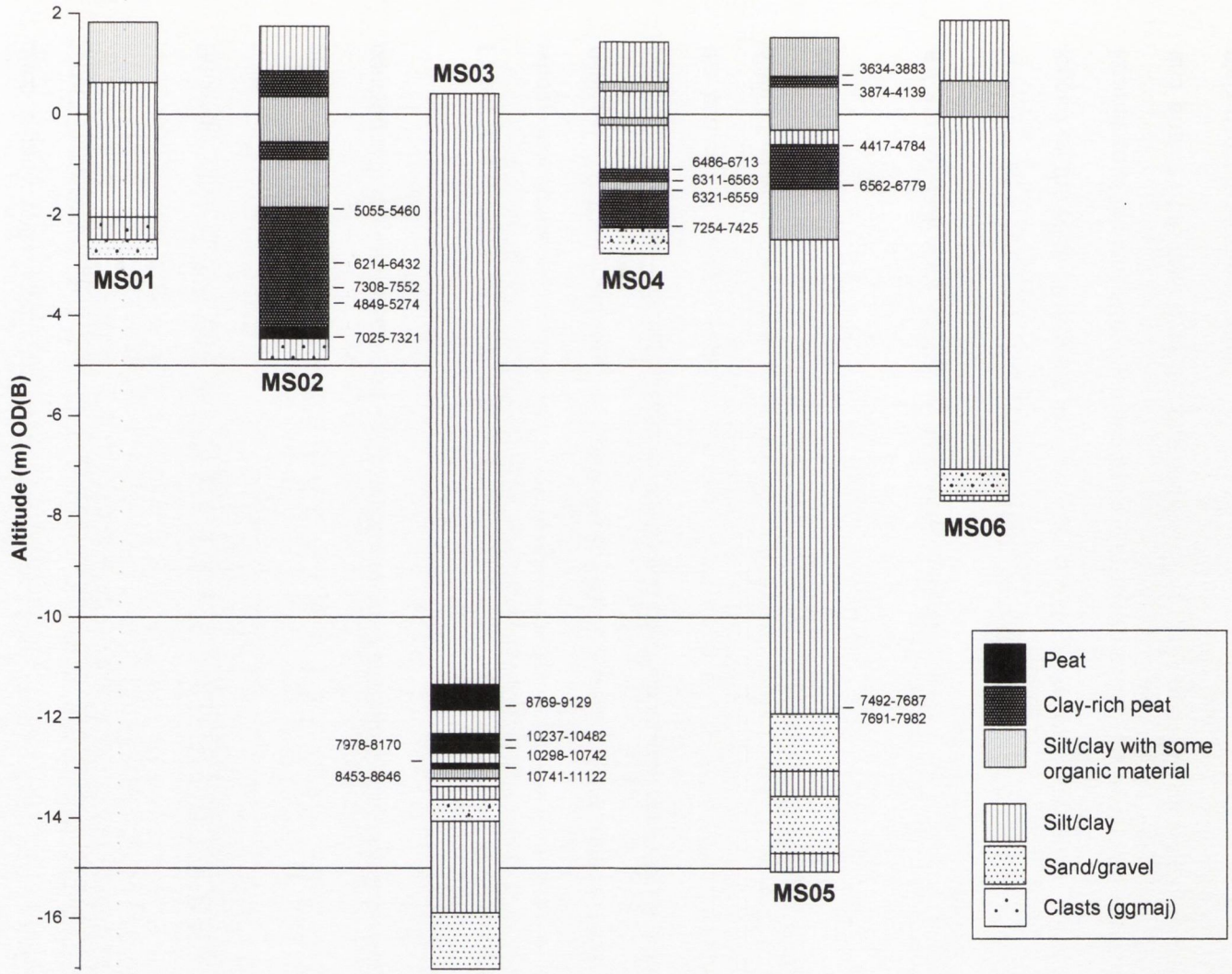


Figure 6.11 Sedimentology and relative altitudes of cores MS01-MS06 referred to in the text. (Dates expressed as calibrated ranges)

on limestone bedrock and are up to 20 m in depth. At the base of a number of the cores is a variable thickness of till which is found resting upon bedrock. This till is largely absent in many of the topographic hollows and is often replaced by a coarse sand/gravel unit. The majority of the cores however, are dominated by fine grained clastic deposits which often rest directly upon the glacial till and coarse sand/gravel unit. The planar upper surface of these fine textured clastic units at c.2.5 m OD(B) betrays the current level of high water spring tides and contrasts strongly with the irregular bedrock, gravel and till surface that it covers. Within these lengthy minerogenic units, there do however, exist a number of inter-leaved organic horizons of varying thickness. These peat horizons are encountered across a range of altitudes between c. +1.0 m OD(B) and -17 m OD(B) although appear laterally discontinuous. Rarely are they found directly overlying bedrock.

6.1.3(ii) MOSTAP core lithostratigraphy

Six MOSTAP cores were extracted from adjacent sites within the inner Shannon estuary. Four of these cores (MS03, MS04, MS05, MS06) were located within 2km of each other whilst two separate cores (MS01 and MS02) were taken in close proximity of one another on the south side of the Shannon estuary. Simplified core lithostratigraphy can be seen in Figure 6.11 whilst detailed logs of each core are included in Tables 6.3a-f. (A full description of the Troels-smith (1955) sediment scheme can be found in Appendix III).

Altitude m (ODB)	Depth (cm)	Physical Properties					Constituent Parts	Description
		nig	strf	sicc	elas	lims		
+1.70 to +0.70	20 - 120	2	0	3	0	-	As3, Th1, ptm +, lf+	Dark clay with organics, iron staining and occasional shell fragments
+0.70 to -0.30	120 - 220	2	0	3	0	0	As4, ptm+	Grey clay with occasional shell fragments
-0.30 to -1.20	220 - 310	2	0	3	0	0	As4, Th+	Grey clay with very occasional organics
-1.20 to -1.82	310 - 372	2	0	2	0	0	As4, Th+, ptm+	Grey clay with occasional shells and organics
-1.82 to -1.96	372 - 386	1+	0	2+	0	0	As4	Light grey clay
-1.96 to -2.41	386 - 431	1	0	3	0	0	As4, Ggmaj+	Creamy brown, firm, 'plastic' clay with occasional stones
-2.41 to -2.80	431 - 470	1+	0	3	0	0	Gmin 2, Ag 1, Ggmaj 1, As+	Sand/clay with frequent stones

Table 6.3a Detailed stratigraphy of MS01 (Ballinacurra East)

Ballinacurra East (MS01)

Core MS01 reveals approximately 4.5 m of minerogenic sediment which varies in size from >5 mm (ggmaj) to <0.002 mm (As) (Table 6.3a). The core terminates in a coarse clastic unit bearing large clasts at a depth of c. -2.80 m OD(B). However, the majority of the core is comprised of a finer silt-clay unit which spans from the top of the core to c. -2.0 m OD(B). No organic horizons are present.

Altitude m (ODB)	Depth (cm)	Physical Properties					Constituent Parts	Description
		nig	strf	sicc	elas	lims		
+1.63 to +1.43	20 - 40	2+	0	3+	0	-	As2, Ag2	clayey silt
+1.43 to +0.94	40 - 89	2	0	3	0	0	As4, Th+, lf++	iron stained clay with occasional organics
+0.94 to +0.43	89 - 140	2+	0	3	0	0	Ag2, th+, sh2	Organic silt
+0.43 to -0.47	140 - 230	2	0	2+	0	0	As3, th1, dl+	Organic clay with occasional wood fragments
-0.47 to -0.82	230 - 265	3	0	2	0	0	As2, th1, sh+, dl1	Very organic clay with large wood fragments
-0.82 to -1.57	265 - 340	2+	0	2	0	0	As3, th1, dl+	Organic clay with frequent wood fragments
-1.57 to -1.97	340 - 380	3	0	2+	0	0	As1, sh3, th+	Clay-rich peat
-1.97 to -2.77	380 - 460	3	2	2+	0	0	As1, th1, sh2, dl+	Organic clay with laminations of clastics and organics
-2.77 to -3.37	460 - 520	3	2	3	0	0	As1, tl+, sh3, dh++	Clay-rich peat
-3.37 to -3.70	520 - 553	3+	2	3	0	0	Sh3, As1, th+	Silty peat
-3.70 to -4.13	553 - 596	3+	0	2+	0	0	As2, sh2, th+	Very dark grey organic clay
-4.13 to -4.37	596 - 620	4	0	3	0	0	Sh4 th+	Black humified peat
-4.37 to -4.65	620 - 648	1	0	2+	0	4	As4, th+	Pale coloured clay with occasional organics
-4.65 to - -4.77	648 - 660	1+	0	3	0	0	As4, Ag+, ggmaj++	Brown/grey very stiff clay with frequent large clasts

Table 6.3b Detailed stratigraphy of MS02 (Ballinacurra West)

Ballinacurra West (MS02)

Core MS02 evidences over 6.5 m of intercalated organic and minerogenic sediment which rests upon bedrock (Table 6.3b). At the base of the core, an over-consolidated clastic unit with large clasts is encountered which fines upwards to a (pale) clay unit at c. -4.65 m OD(B). Resting unconformably over this clay unit is a thin black humified peat horizon of approximately 25 cm thickness which grades into a dark organic clay at c. -4.1 m OD(B). Between c. -4.1 m OD(B) and

+1.0 m OD(B), a series of clay/silt-rich peat units inter-leave with minerogenic sediment horizons. All of the peat units are poorly humified and fragments of wood are clearly visible. Above c. +1.0 m OD(B), the core is comprised of fine grained clastic material which is partially iron stained.

Altitude m (ODB)	Depth (cm)	Physical Properties					Constituent Parts	Description
		nig	strf	sicc	elas	lims		
+0.28 to -0.22	20-70	2+	0	3+	0	-	As 2, Ag 2, Th +	Dark brown soil
-0.22 to -1.40	70-188	2	0	3	0	0	As 3, Ag 1, Fe +, Th +	Silty clay with occasional organics and iron staining
-1.40 to -1.72	188-220	4	0	3+	0	0	As 4	Jet black clay
-1.72 to -3.87	220-435	2+	0	3	0	0	As 4, Th +	Grey clay with occasional organics. (Well preserved plant fragments)
-3.87 to -6.02	435-650	2+	0	2+	0	0	As 4, (occ. Ptm+, Th+)	Darkish grey clay with (very) occasional shell fragments and organic flecks)
-6.02 to -8.22	650-870	2+	0	2+	0	0	As 4, (occ. Ptm+, Th+)	Darkish grey clay with (very) occasional shell fragments and organic flecks)
-8.22 to -8.72	870-920	2	0	2	0	0	As 4, ptm ++, Th ++	Green/grey clay with occasional shell and organic fragments
-8.72 to -9.22	920–970	2	2	2	0	0	As 4, ptm +, Th +	Laminated clay deposit with shell and organic fragments
-9.22 to -10.62	970-1110	2	0	2	0	0	As 4, ptm ++, Th ++	Green/grey clay with occasional shell and organic fragments
-10.62 to -11.34	1110-1182	2	2	3	0	0	As 4, ptm + (occ. Layers of ptm ++	Clay with frequent layes of shells. (Well preserved plant fragments)
-11.34 to -11.84	1182-1232	3+	0	4	0	3	Sh 4	Very well humified peat (but no visible plant fragments.) Abrupt contact with overlying layer clay
-11.84 to -12.32	1232-1280	2	0	2	0	0	As 3, Ag 1	Light grey clay
-12.32 to – 12.47	1280-1295	3+	0	3	0	3	Sh 4, Gmin +, As +	Well humified peat. Abrupt contact with overlying layer clay
-12.47 to -12.50	1295-1298	3	0	2	0	1	Gmin 3, As 1	Thin band of sand
-12.50 to -12.62	1298-1310	3+	0	3	0	3	Sh 4, Gmin +, As +	Well humified peat. Abrupt contact with underlying clay unit
-12.62 to -12.91	1310-1339	2	0	2+	0	2	As 4, Gmin +, th++, Ptm +	Greenish clay with occasional sand, organics and shell fragments
-12.91 to -13.02	1339-1350	3	0	3	0	2	(Sh 4), (Sh 1, Ptm1, Gmin 2)	Peat and sand lenses

-13.02 to -13.22	13.50 – 13.70	2	0	2+	0	0	As 2, Ag 1, Th 1	Green/grey organic clay
-13.22 to -13.37	13.70 – 13.85	2	0	2+	0	0	Gmaj 1, Gmin 2, Ag 1, Th +	Sandy silt with intercalations of gravel and occasional organics
-13.37 to -13.64	13.85 – 14.12	2	0	2+	0	0	As 2, Ag 1, Gmin +, Th +	Silty clay with a trace of sand and organics. Occasional intercalations of gravel
-13.64 to -14.07	14.12 – 14.55	2	0	2	0	0	Gmin 2, Gmaj 1, Ag 1, Th+, GG maj+	Grey sand with occasional organics and stones
-14.07 to -14.82	14.55 – 15.30	1	0	2	0	0	As 4	Fine clay
-14.82 to -15.89	15.50 – 16.37	2	0	1+	0	2	As 4	Fine clay with brownish tint.
-15.89 to -16.94	16.37 – 17.42	2	0	2	0	2	Gmin 2, Gmaj 1, Ag 1	Light grey silty sand/gravel. (Sharp contact with overlying clay unit)

Table 6.3c Detailed stratigraphy of MS03 (Coonagh South)

Coonagh South (MS03)

The core taken from Coonagh South is the longest of the six MOSTAP cores and reveals over 17 m of predominantly clastic sediment (Table 6.3c). The core terminates in a light grey silt/sand unit which rests upon bedrock at c. -17 m OD(B). This clastic sediment unit persists to an altitude of c. -13.0 m OD(B), although variations in particle size are apparent. Between c. -13.0 and -11.0 m OD(B), the stratigraphy becomes more complex: A number of well humified and compacted peat units are present and these are separated from thin (<5 cm) bands of sand/shells by sharp, erosive contacts. An exception to this pattern however, is the conformable (regressive) contact between the well humified peat unit at -11.34 to -11.84 m OD(B) and the underlying light grey clay unit. Above these intercalated peat horizons, clastic sediments once again dominate and the top 11.5 m of the core is almost entirely comprised of clay although both shells and occasional organics are present within the sediment matrix.

Coonagh West (MS04)

Over 4 m of sediment is recovered in core MS04 from Coonagh West and details of this are given in Table 6.3d. The core terminates at a depth of c.-2.7 m OD(B) in a saturated black silt/sand unit with large clasts. The unit gradually fines upwards to c.-2.2 m OD(B) where it is overlain unconformably by an organic clay horizon. Between c. -1.0 m OD(B) and -2.0 m OD(B), organic material becomes more abundant and represents between 25-75% of the sediment matrix. The top 2.5 m of this core is dominated by fine-grained sediment with silt/clay commonly accounting for

over 75% of the sediment in this unit. Plant fragments are often visible and the core becomes heavily iron stained above +0.5 m OD(B).

Altitude m (ODB)	Depth (cm)	Physical Properties					Constituent Parts	Description
		nig	strf	sicc	elas	lims		
+1.31 to +0.91	20-60	2+	0	2+	0	-	Ag2, As2, th+, lf++	Iron stained organic clay/silt
+0.91 to +0.71	60-80	2+	0	2+	0	0	Lf1, Ag2, As1, th++	Heavily ironed stained clayey silt
+0.71 to +0.53	80-98	2+	0	2+	0	0	Lf++, th1, As2, Ag1	Iron stained organic silty clay
+0.53 to +0.01	98-150	2	0	2+	0	0	As4, th++	Light grey clay with organics
+0.01 to -0.14	150-165	2+	0	2+	0	0	As3, Dh1	Soft clay with visible plant remains
-0.14 to -1.03	165-254	2	0	2+	0	0	As4, th++	Light grey clay with organics
-1.03 to -1.28	254-279	3	1	2+	0	0	As2. sh2, th+	Very organic clay
-1.28 to -1.44	279-295	2+	0	2+	0	0	As3, sh1, dh+	Organic clay
-1.44 to -2.01	295-352	3+	0	2+	0	0	Sh3, As1, th+	Clay-rich peat
-2.01 to -2.18	352-369	3	0	2+	0	0	Th1, sh1, As2	Organic clay
-2.18 to -2.39	369-390	1	0	2	0	1	Th+, gmin3, ggmaj++, Ag1	pale (heavily consolidated) silty sand with large clasts
-2.39 to -2.69	390-420	4	0	1	0	0	Ggmaj1, gmin2, Ag1, th++	Saturated black silty/sand unit with large clasts and occasional organics

Table 6.3d Detailed stratigraphy of MS04 (Coonagh West)

Meelick Creek (MS05)

Core MS05 is taken from Meelick Creek which is located approximately 500 m to the north of Coonagh. Here, just under 17 m of sediment has been recovered and the details of this core are presented in Table 6.3e. The core terminates in a very stiff pink clay which overlies bedrock at c. -15 m OD(B). Above this unit between c.-12 m OD(B) and -14.5 m OD(B), the core is dominated by clastic sediment with sand accounting for between 25-75% of the sediment matrix. Overlying this section, a lengthy, (largely homogenous) silt-clay unit is present to an altitude of c. -2.4 m OD(B). Variations in the ratio of silt to clay do exist however, and sand is also present in the sediment column. In addition, a number of small (<5 mm) shells (*Hydrobia ulvae*) are also evident. The top 4 m of the core is characterized by intercalations of organic and minerogenic sediment which rest

conformably over one another. All of the clay-rich peat units are poorly humified and both wood fragments and reeds are frequently encountered.

Altitude m (ODB)	Depth (cm)	Physical Properties					Constituent Parts	Description
		nig	strf	sicc	elas	lims		
+1.40 to +0.85	20-75	2+	0	2+	0	-	As3, Dl1, lf++, Sh+	Heavily iron stained clay with occasional woody fragments
+0.85 to +0.60	75-100	3	0	2+	0	0	Th1, As2, dl+, dh1	Very organic clay
+0.6 to -0.25	100-185	3	0	2+	0	0	As3, th1, dl+, ptm+	Organic clay with occasional shell fragments and twigs
-0.25 to -0.54	185-214	3	0	2+	0	0	As4, th+	Dark grey clay with organic fragments
-0.54 to -0.70	214-230	3	0	2	0	0	Th1, As1, sh2 dl+	Very organic silt/clay
-0.70 to -1.07	230-267	3+	0	2	0	0	Dl+, th1, sh1, As2	Very organic clay with woody fragments
-1.07 to -1.40	267-300	3+	0	2	0	0	Th1, As1, Sh2	Clay-rich peat
-1.40 to -2.40	300-400	3	0	2	0	0	As3, Sh1, Dh++	Darkish grey clay with frequent (reeds)
-2.40 to -3.60	400-520	2	0	2	0	0	As3, Ag1	Grey clay with a trace of silt
-3.60 to -5.40	520-700	2	0	2	0	0	As2, Ag2, gmin+	Silt/clay with trace of sand
-5.40 to -5.60	700-720	2	0	2	0	0	As1, Ag2, gmin1	Sandy silt with some clay
-5.60 to -7.00	720-860	2	0	2	0	0	As1, Ag3, gmin+	Clayey silt with some sand. (occasional bands of As3, Ag1)
-7.00 to -8.70	860-1030	2+	0	2	0	0	As2, Ag2	Grey silt/clay
-8.70 to -11.85	1030-1345	2+	0	2	0	0	As3, Ag1, ptm +	Silty clay
-11.85 to -13.00	1345-1460	2+	0	2+	0	0	Gmin2, Ag2, ptm+	Sandy silt with occasional small shells
-13.00 to -13.50	1460-1510	2+	0	2+	0	0	Ag3, gmin1	Sandy silt
-13.50 to -14.00	1510-1560	2+	0	2+	0	0	Ag2, gmin2,	Silt/sand
-14.00 to -14.64	1560-1624	2+	0	2	0	0	Gmin3, Ag1	Silty sand
-14.64 to -14.99	1624-1659	1+	0	3	0	4	As4	Brown/pink very stiff clay (occasional As2, Ag2 intrusions)

Table 6.3e Detailed stratigraphy of MS05 (Meelick Creek)

Clonmacken (MS06)

Core MS06 has been extracted from the site at Clonmacken and contains just under 10 m of sediment. The stratigraphy from this core is detailed in Table 6.3f and is comprised entirely of clastic sediment. At the bottom of the core, a saturated soft black clay unit is found resting on bedrock at c. -7.6 m OD(B). This unit is very thin and is overlain by a coarse sediment unit comprised predominantly of sand and clasts at c. -7.5 m OD(B). Generally, the core fines with altitude and above -6.5 m OD(B), the sediment matrix comprises >75% clay although both shells and wood fragments are encountered.

Altitude m (ODB)	Depth (cm)	Physical Properties					Constituent Parts	Description
		nig	strf	sicc	elas	lims		
+1.74 to +1.35	20-59	2	0	2+	0	-	As3, lf+, Ag1	Silty clay with iron staining
+1.35 to +0.74	59-120	1+	0	3	0	0	As4, lf++	Stiff iron stained clay
+0.74 to +0.02	120-192	2+	0	3	0	0	As3, Dl1, sh+	Organic clay with wood fragments
+0.02 to -1.41	192-335	2	0	3	0	0	As4	Light grey clay
-1.41 to -2.41	335-435	2	0	3	0	0	As4, dh+	Light grey clay with organic fragments
-2.41 to -2.62	435-456	2	0	3	0	0	As4	Light grey clay
-2.62 to -2.70	456-464	4	0	2+	0	4	Stirp?	Tree stump or large branch?
-2.70 to -5.36	464-730	2+	0	2	0	0	As4	Grey clay. Very occasional shells and organics
-5.36 to -5.66	730-760	2+	0	2	0	0	As4, ptm+	Grey clay. occasional shells
-5.66 to -5.93	760-787	2+	0	2	0	0	As4	Grey clay. Very occasional shells and organics
-5.93 to -6.51	787-845	1+	1	3	0	4	As3, ptm+	Light coloured laminated shelly clay
-6.51 to -6.98	845-892	1+	0	2+	0	4	As2, gmin1, Ag1	Pinkish clay with occasional intrusions of black sand/silt
-6.98 to -7.51	892-945	1+	0	2+	0	0	Ag1, As+, gmin2, ggmaj1	Very coarse sandy silt with many clasts
-7.51 to -7.61	945-955	3	0	1+	0	4	As4	Saturated soft black clay

Table 6.3f Detailed stratigraphy of MS06 (Clonmacken)

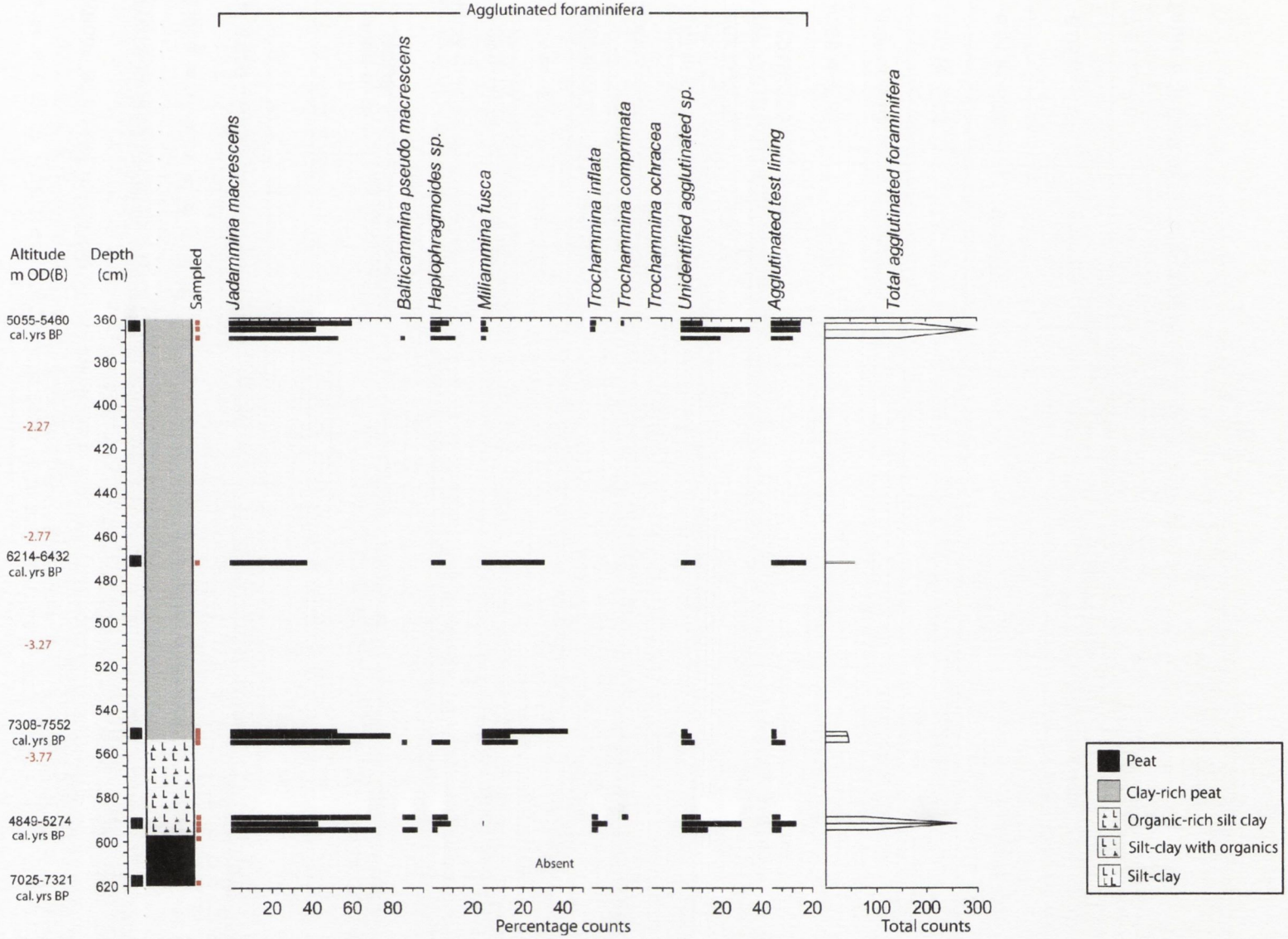


Figure 6.12a Fossil foraminifera from core MS02 with associated lithostratigraphy and dated horizons

Figure 6.12b Fossil foraminifera from core MS03 with associated lithostratigraphy and dated horizons

6.1.3(iii) MOSTAP core biostratigraphy

Ballinacurra West (MS02)

Core MS02 revealed an abundance of foraminifera contained within many of the clastic sedimentary units and clay-rich peat sections described in Table 6.3b. Counts of 40 or higher were permitted in all of the samples and these are presented in Figure 6.12a along with associated lithostratigraphy and dated horizons. Between 360 and 595 cm, all of the samples are composed entirely of agglutinated foraminiferal species and no calcareous species are found. However, in both the peat unit and coarse clastic (basal) unit found below 595 cm, no foraminifera are present. Each of the samples encountered between 360 and 595 cm are dominated by *Jadammina macrescens* which frequently comprise over 50% of the assemblage. *Jadammina macrescens* are most abundant at a depth of 550 cm, at the base of the silty-peat unit. Here, they make up over 80% of the foraminiferal assemblage. *Miliammina fusca* are also found in reasonably high concentrations in the silt-rich peat unit although rarely make up over 40% of the sample population. Test linings are discovered in all of the samples although these seldom total more than 15% of the sample population.

Coonagh South (MS03)

In core MS03, foraminiferal sampling was concentrated around the transitions between the organic and clastic units encountered between -11 m OD(B) and -13 m OD(B). With the exception of the clay unit found at a depth of 13.10 to 13.39 m, no foraminifera were present in this section of the core. However, in the clay unit, counts well in excess of 40 were permitted in all of the samples and these are presented in Figure 6.12b along with associated lithostratigraphy and dated horizons. The three foraminifera samples from the minerogenic unit are markedly similar to one another, with each assemblage comprised of over 60% *Ammonia sp.* The agglutinated component is negligible in all three samples whilst all other species make up less than 10% of the total sample population. No foraminifera are found within the peat units above and below the clay unit.

Although no foraminifera were found across the (conformable) regressive contact at -11.84 m OD(B), extremely low concentrations of both pollen and diatoms were present. Two frustules of the estuarine diatom species *Nitzschia navicularis* were present along with very low counts of both *Pinus* and *Betula* pollen. An individual pollen grain of *Ruppia sp.* was also present although it

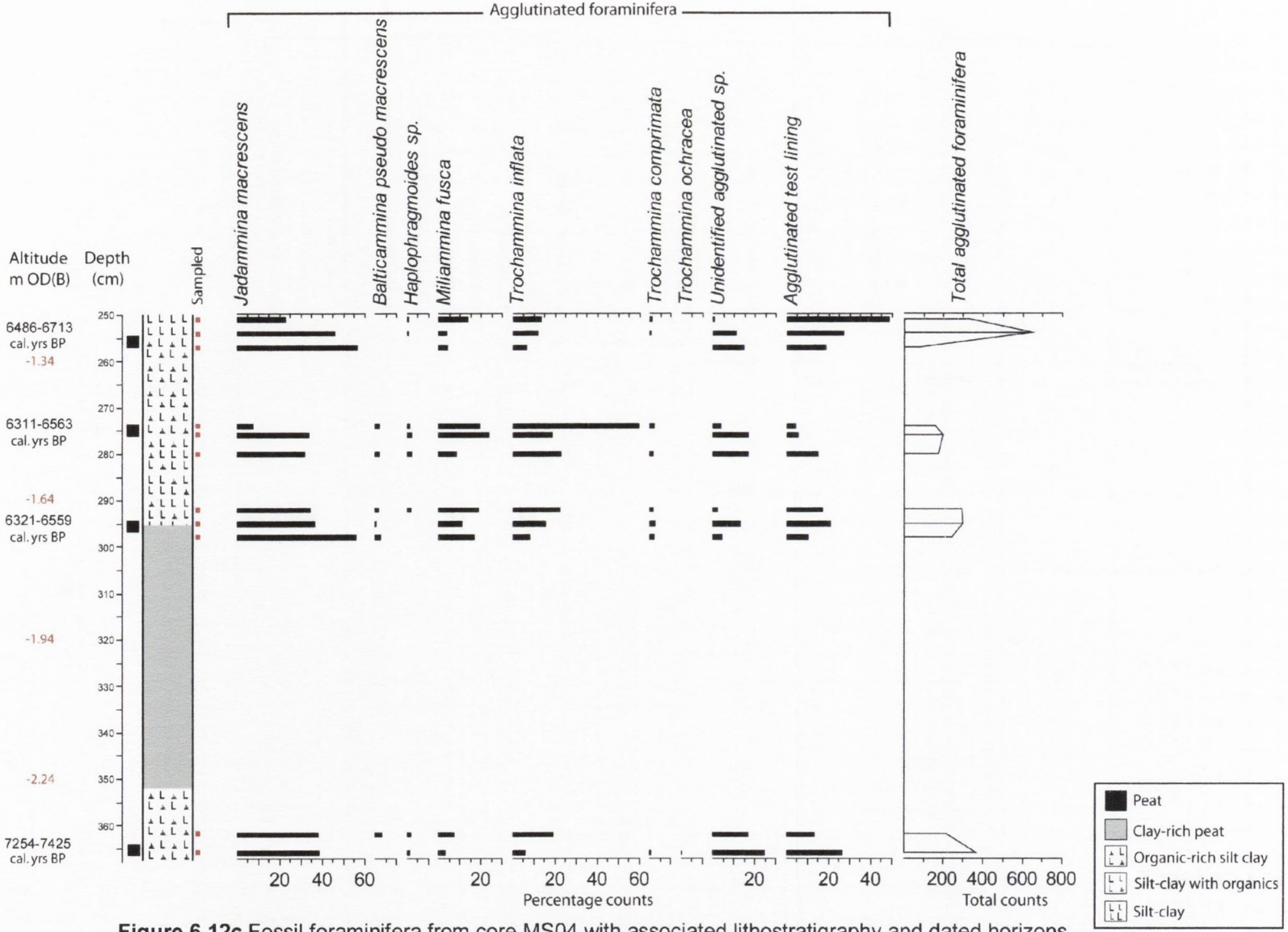


Figure 6.12c Fossil foraminifera from core MS04 with associated lithostratigraphy and dated horizons

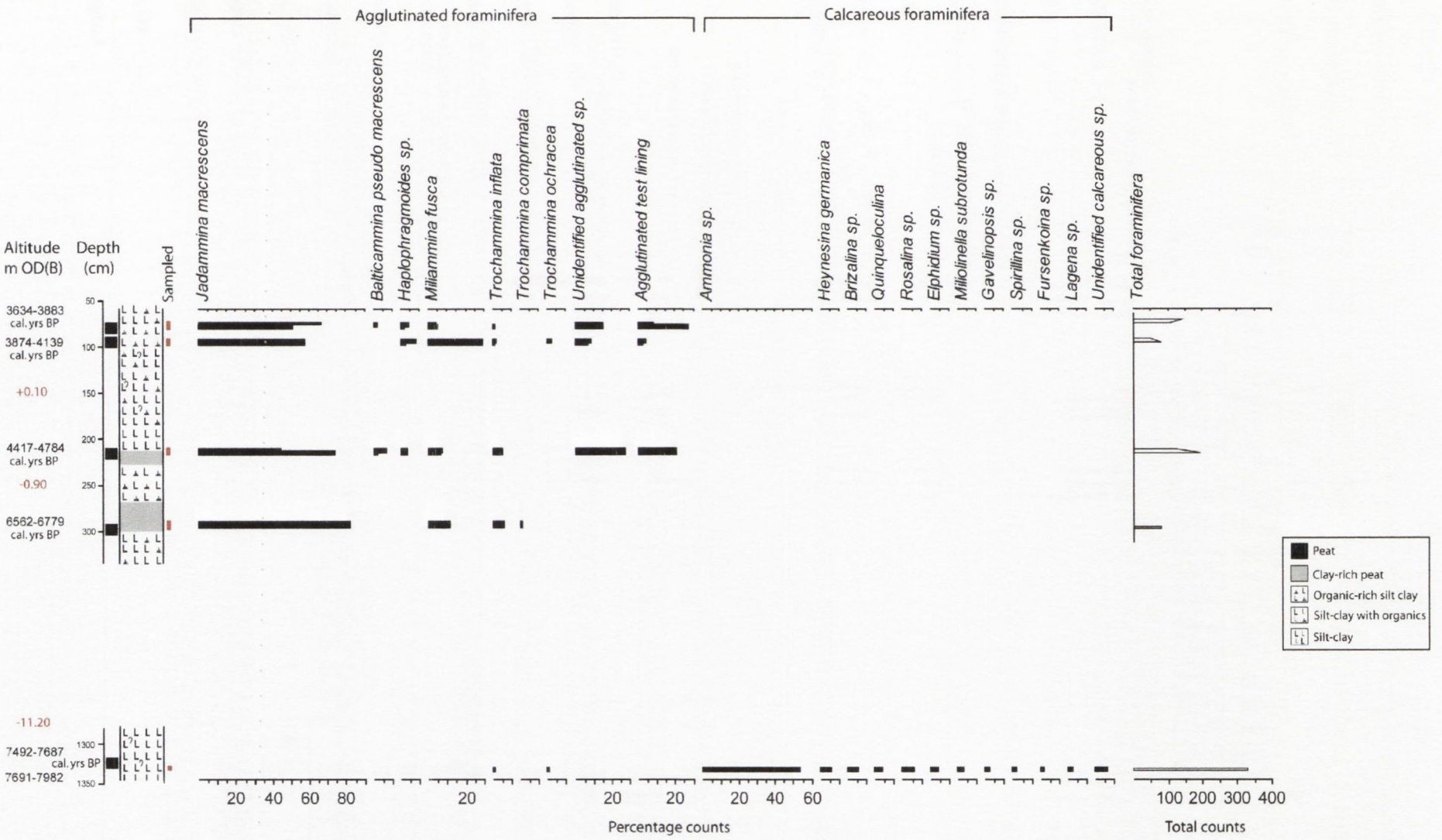


Figure 6.12d Fossil foraminifera from core MS05 with associated lithostratigraphy and dated horizons

remains unclear whether this was *Ruppia maritima*, (a coastal species found in brackish water and tolerant of tidal exposure).

Coonagh West (MS04)

In core MS04, sampling for foraminifera was focused on the organic clay units which are present below c. -1.0 m OD(B). All samples contained relatively abundant foraminifera enabling counts of 40 or more. The results of these counts are presented in Figure 6.12c. All of the samples are made up of an entirely agglutinated assemblage and no calcareous forms are found. *Jadammina macrescens* is the most dominant species and commonly accounts for 40% or more of the sample population. This species is most dominant (c.60% population) in the clay-rich peat unit found at a depth of 295 cm and least dominant (c.10% population) in the organic clay unit found at 275 cm. Both *Miliammina fusca* and *Trochammina inflata* are also found in relatively high proportions although with the exception of the sample taken from a depth of 275 cm, both of these species are subordinate to *Jadammina macrescens.* Agglutinated test linings are also prevalent throughout this section of the core and comprise c.50% of the sample population at a depth of 252 cm.

Meelick Creek (MS05)

Foraminiferal sampling from core MS05 was concentrated in the top 3 m of the core where minerogenic and organic layers inter-leaved. However, a solitary sample was also counted from a depth of 1331 cm since this horizon provided material for a paired foraminifera-shell date. The results of these counts are presented in Figure 6.12d. In the uppermost 3 m of the core, agglutinated foraminifera are abundant whilst calcareous species are absent. *Jadammina macrescens* are the most common species in the assemblage and are particularly dominant in the clay-rich peat horizons found at a depth of 220 cm and 270 cm. A number of other agglutinated species are also present although with the exception of *Miliammina fusca*, none of these species comprise more than 10% of any sample population.

The sample taken from a depth of 1331 cm is dominated by calcareous foraminifera and the agglutinated faction is negligible. *Ammonia sp.* dominate the assemblage and make up over 50% of the sample population. The remainder of the assemblage is evenly split between other

calcareous forms although rarely do these species constitute more than 5% of the sample population.

Site	Laboritory Code	Core	Material	Depth (cm)	Altitude (m ODB)	Weight (g)	^{14}C Age	Calibrated Age BP
Meelick Creek	Beta 211398	MS05	Organic Sediments	75	0.85	11.5	3480±50	3634-3883
Meelick Creek	Beta 203549	MS05	Peat	94	0.69		3660±40	3874-4139
Meelick Creek	Beta 203547	MS05	Peat	214	- 0.50		4030±40	4417-4784
Ballinacurra Creek (west)	Beta 211394	MS02	Peat	590	- 4.07	22.5	4390±50	4849-5274
Ballinacurra Creek (west)	Beta 203540	MS02	Peat	363	- 1.76		4590±40	5055-5460
Ballinacurra Creek (west)	Beta 203539	MS02	Peat	470	- 2.83		5530±50	6214-6432
Coonagh (west)	Beta 203544	MS04	Peat	276	- 1.21		5660±50	6311-6563
Coonagh (west)	Beta 203546	MS04	Organic Sediments	295	- 1.40		5670±40	6321-6559
Coonagh (west)	Beta 211397	MS04	Organic Sediments	254	- 1.03	25.0	5790±40	6486-6713
Meelick Creek	Beta 203548	MS05	Peat	293	- 1.29		5860±40	6562-6779
Ballinacurra Creek (west)	Beta 203536	MS02	Peat	618	- 4.31		6290±50	7025-7321
Coonagh (west)	Beta 203545	MS04	Organic Sediments	366	- 2.11		6400±50	7254-7425
Ballinacurra Creek (west)	Beta 203538	MS02	Peat	550	- 3.63		6500±50	7308-7552
Meelick Creek	Beta 211399	MS05	Shell *Hydrobia ulvae*	1329	- 11.69	0.01	7120±50	7492-7687
Meelick Creek	Beta 216647	MS05	Foraminifera	1331	- 11.71	0.006	7390±70	7691-7982
Coonagh (South)	Beta 211396	MS03	Shell *Hydrobia ulvae*	1335	- 12.87	0.01	7620±40	7978-8170
Coonagh (South)	Beta 211395	MS03	Organic Sediments	1335	- 12.87	24.0	7800±40	8453-8646
Coonagh (South)	Beta 203537	MS03	Peat	1224	- 11.72		8070±50	8769-9129
Coonagh (South)	Beta 203541	MS03	Peat	1292	- 12.40		9170±40	10237-10482
Coonagh (South)	Beta 203542	MS03	Peat	1308	- 12.56		9350±70	10298-10742
Coonagh (South)	Beta 203543	MS03	Peat	1348	- 12.96		9580±40	10741-11122

Table 6.4 Radiocarbon dates from the Inner Shannon estuary

6.1.3(iv) MOSTAP core chronostratigraphy

Of the six cores taken from the sites in the inner Shannon, four cores (MS02, MS03, MS04, MS05) provided suitable material for AMS radiocarbon dating. Details of all these dates are contained within Table 6.4 and Appendix V. Five dates were established from the (brackish) peat unit in

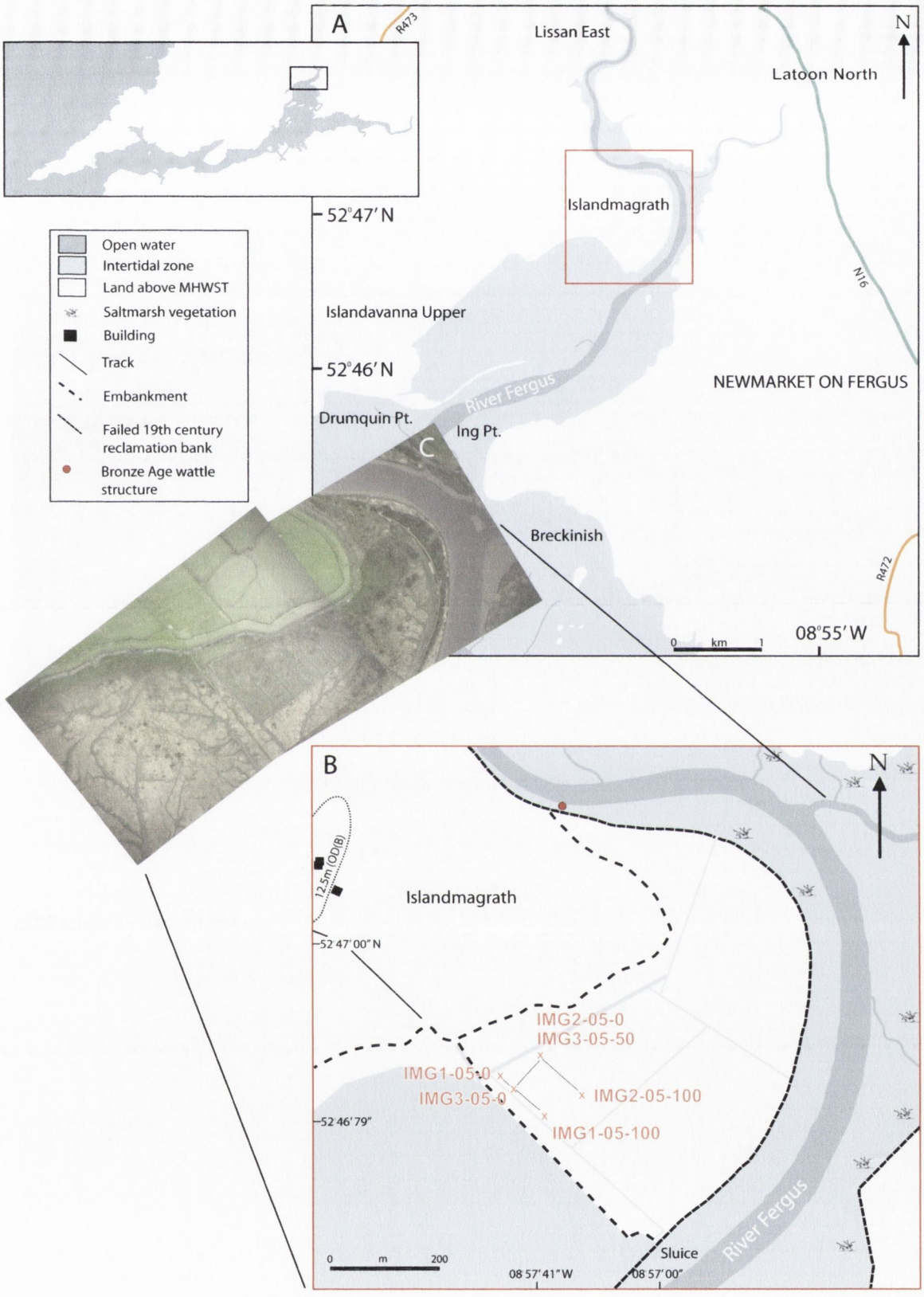


Figure 6.13 Site map showing **A** the location of the Islandmagrath study site in relation to the Fergus estuary; **B** the location of the transects at Islandmagrath; **C** aerial photograph(s) of the study site



Plate 6.3 The study site at Islandmagrath (looking southeast)

Photo taken August 2005

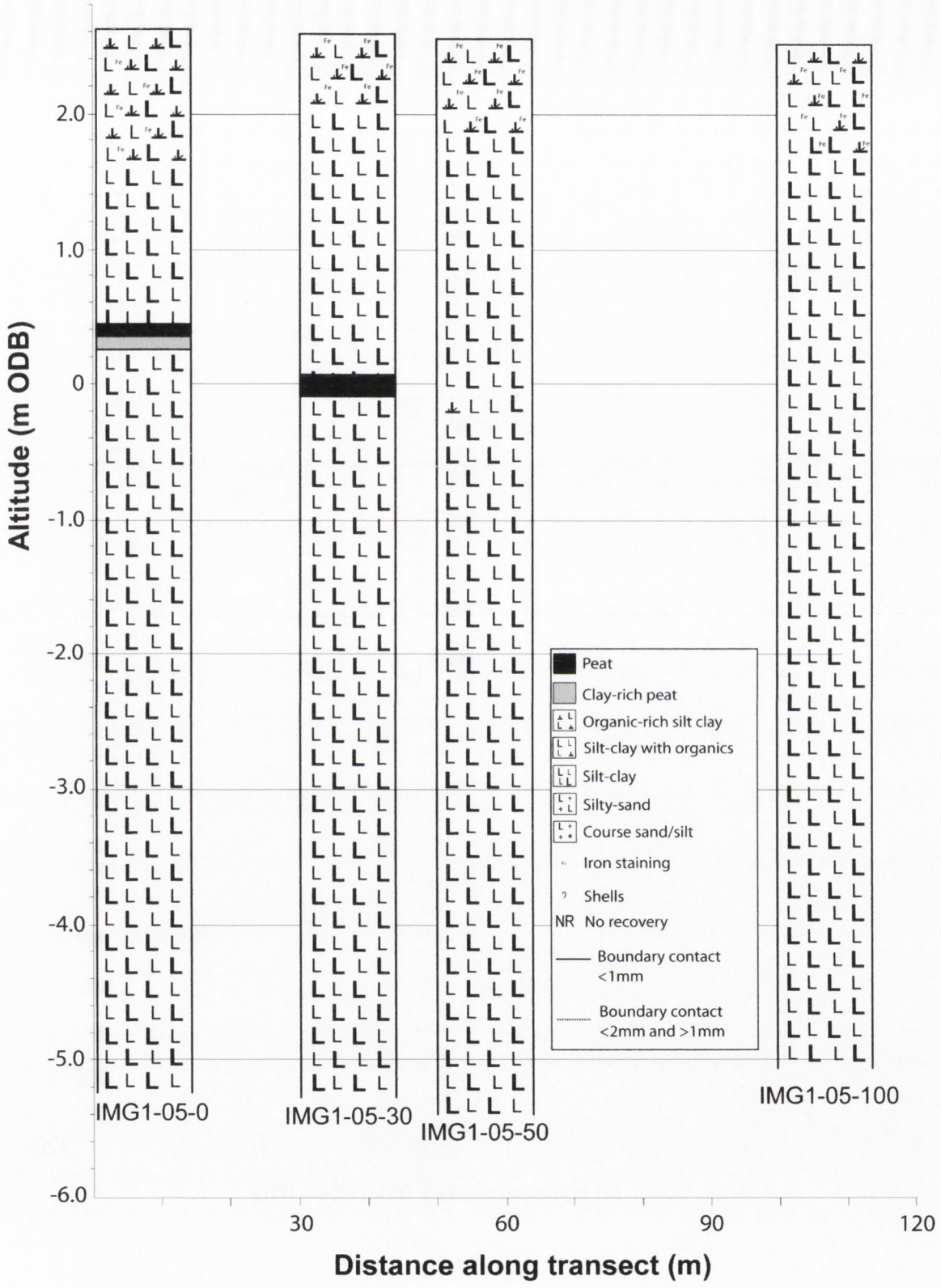


Figure 6.14a Summary lithostratigraphic diagram for transect IMG1 (Islandmagrath). See figure 6.13B for transect locations

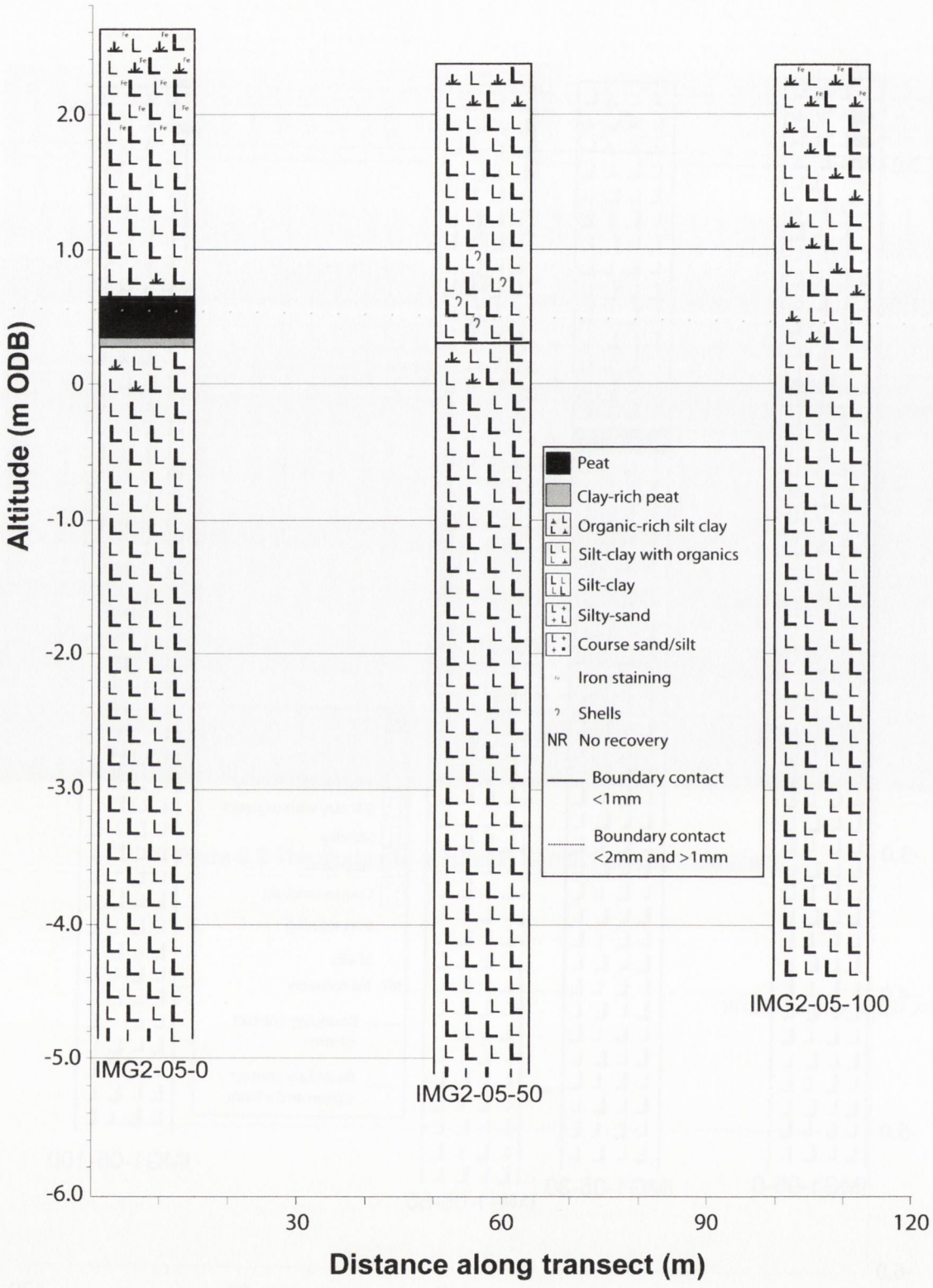


Figure 6.14b Summary lithostratigraphic diagram for transect IMG 2 (Islandmagrath). See figure 6.13B for transect locations

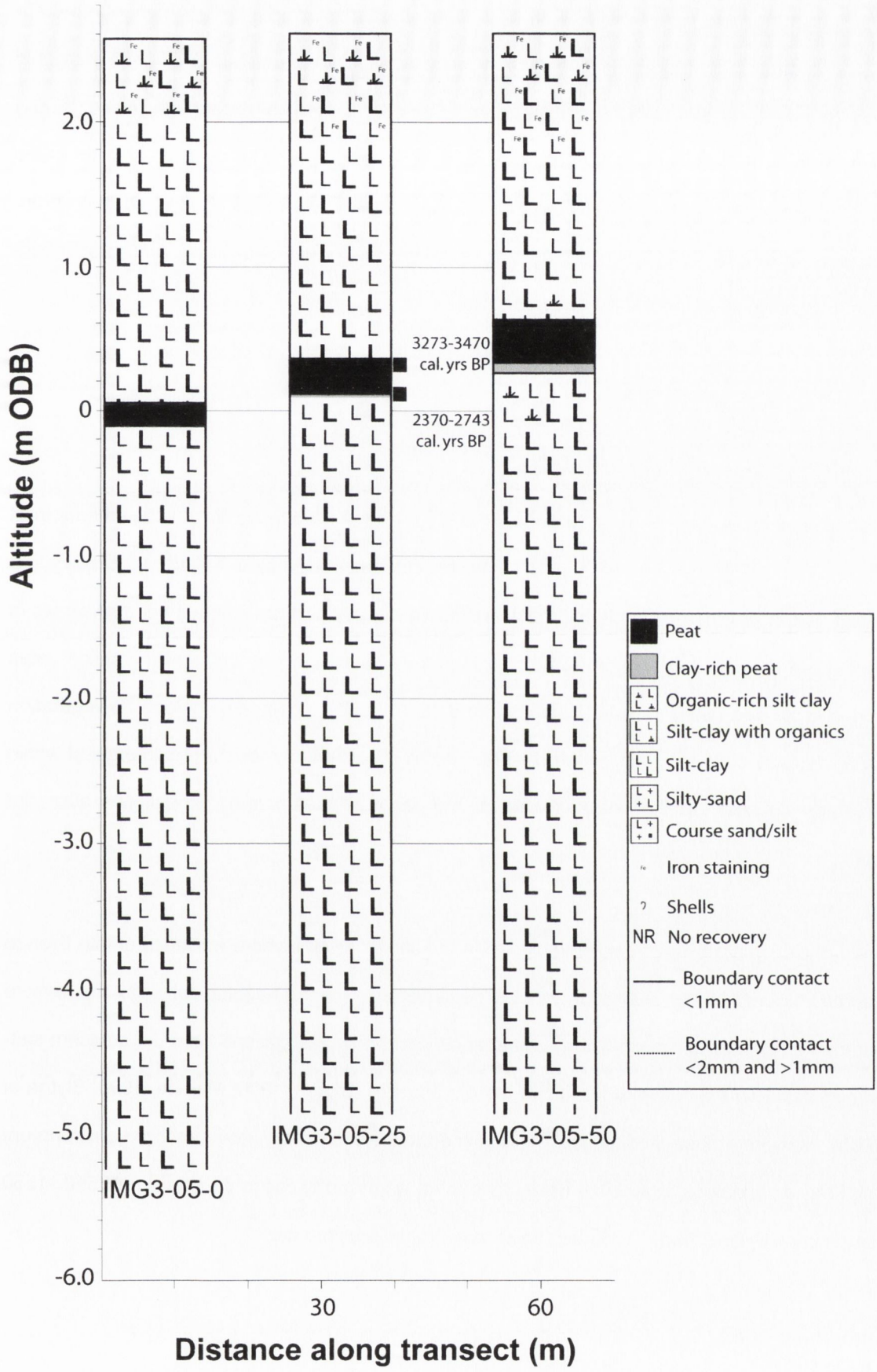


Figure 6.14c Summary lithostratigraphic diagram for transect IMG3 (Islandmagrath). See figure 6.13B for transect locations

MS02 and a further four dates were taken from the deep humified peat horizons encountered in MS03. These dates from MS03 were accompanied by two additional dates taken from the organic clay unit found at -12.87 m OD(B) in this core. The core taken from Coonagh West (MS04) provided 4 dates from the brackish peat horizons in the lower section of the core and the same number of dates were also provided by the brackish peat units found above -2.0 m OD(B) in MS05. Two further dates were taken from MS05 with a paired shell (*Hydrobia ulvae*) and foraminifera (*Ammonia sp.*) date taken from a depth of 13.31 m.

6.1.4 Study Sites From the Middle Shannon

6.1.4(i) Islandmagrath (Fergus estuary)

As with the sites located in the Inner Shannon, the study site at Islandmagrath is situated on a tract of reclaimed land (Figure 6.3 and Figure 6.13b). The site is bounded by a (failed) 19th century sea wall along its eastern margin and to the north and south west by earthen embankments. A series of deeply incised (c.2 m) drainage ditches bisect the site and were once controlled by sluice gates positioned in the sea wall (Figure 6.13c and Plate 6.3). The site is covered by halophytic vegetation (which is currently used for pastoral farming) and isolated pockets of reeds are also present within the drainage ditches. On the seaward side of the sea wall, salt-marsh communities align the estuary.

The site at Islandmagrath has been referred to in O'Sullivan (2001) where evidence of late Bronze Age human activity in the inter-tidal zone is documented. The archaeological finds made here encouraged a number of (unpublished) palaeoenvironmental investigations from the adjacent salt-marsh cliff found on the seaward margin of the old sea wall (Blythe, 1996; Malone, 1996; Blythe *et al.*, 1996). However, these studies choose to concentrate on the immediate foreshore environment in which the archaeological artifacts were discovered and tend to focus on Iron age (2550 -1550 BP) and early historic (1550 – 1150 BP) environmental developments.

a) Lithostratigraphy

Three transects made up of eight cores were collected from the site at Islandmagrath. Core lithostratigraphy is shown in Figure 6.14a-c. The cores range between 7-8 m in depth and provide material from a maximum depth of just over -5 m OD(B). All of the cores reveal a considerable

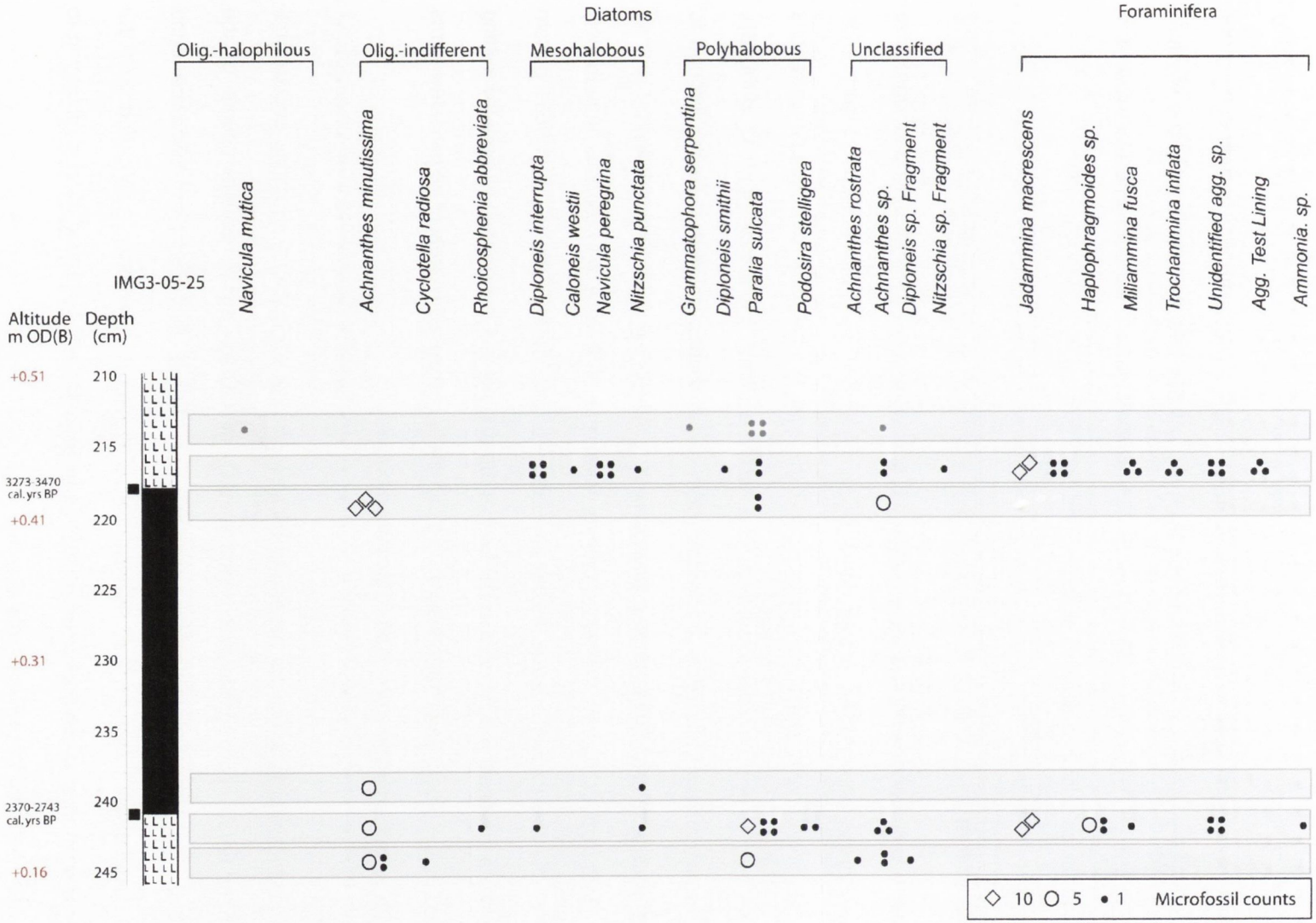


Figure 6.15 Islandmagrath fossil foraminifera and diatoms from core IMG3-05-25 with associated lithostratigraphy and dated horizons. (Diatom analysis by Caroline Hillier)

thickness of silt and clay and none of the cores are grounded on bedrock. (Sediment below 8 m in depth proved to be irretrievable by auger core). Below c.+1.5 m OD(B) all of the lengthy clastic units are largely homogenous although subtle down-core changes do exist in the silt:clay ratios. Although each of the cores is dominated by silt and clay sediment, 4 of the cores contained thin (<50 cm) peat horizons. These vary in altitude, ranging from c.+0.6 m OD(B) in core IMG3-05-50 to c.0 m OD(B) in core IMG3-05-0. In the peat unit encountered in both core IMG1-05-30 and in core IMG2-05-0, the transgressive contact with the overlying silt clay unit is abrupt, evidencing erosion. Above c. +1.5 m OB(B), an iron-stained organic rich silt/clay unit is present in all cores apart from IMG2-05-50 where no iron staining is observed and organic material is less abundant.

b) Biostratigraphy

Foraminiferal and diatom analyses were carried out across both the transgressive and regressive peat–clay contacts encountered in core IMG3-05-25. This evidence is put forward in Figure 6.15. Foraminifera and/or diatoms are present in all six of the analyzed samples although concentrations of both are low. Across the transgressive contact (which occurs at an altitude of +0.43 m OD(B)), the diatom assemblages evidence a transition from oligohalobous-indifferent species through to polyhalobous specimens. *Achnanthes minutissima* dominate in the peat horizon but are replaced by the mesohalobous diatom species *Diploneis interrupta* and *Navicula peregrina* immediately above the lithostratigraphic contact. In the silt-clay unit, low counts of the polyhalobous diatom *Paralia sulcata* are observed. Further evidence that the silt-clay unit is indicative of a marine setting is provided by the presence of foraminifera. The sample is comprised entirely of agglutinated forms with *Jadammina macrescens* dominating the assemblage.

Both the diatom and foraminiferal assemblages also provide evidence for a marine depositional environment below the regressive contact at +0.2 m OD(B). Again, several counts of the polyhalobous diatom species *Paralia sulcata* are present in the silt-clay unit and these are found alongside agglutinated foraminifera. As with the foraminiferal assemblage associated with the transgressive contact, *Jadammina macrescens* dominate although a number of tests belonging to *Haplophragmoides sp.* are also present.

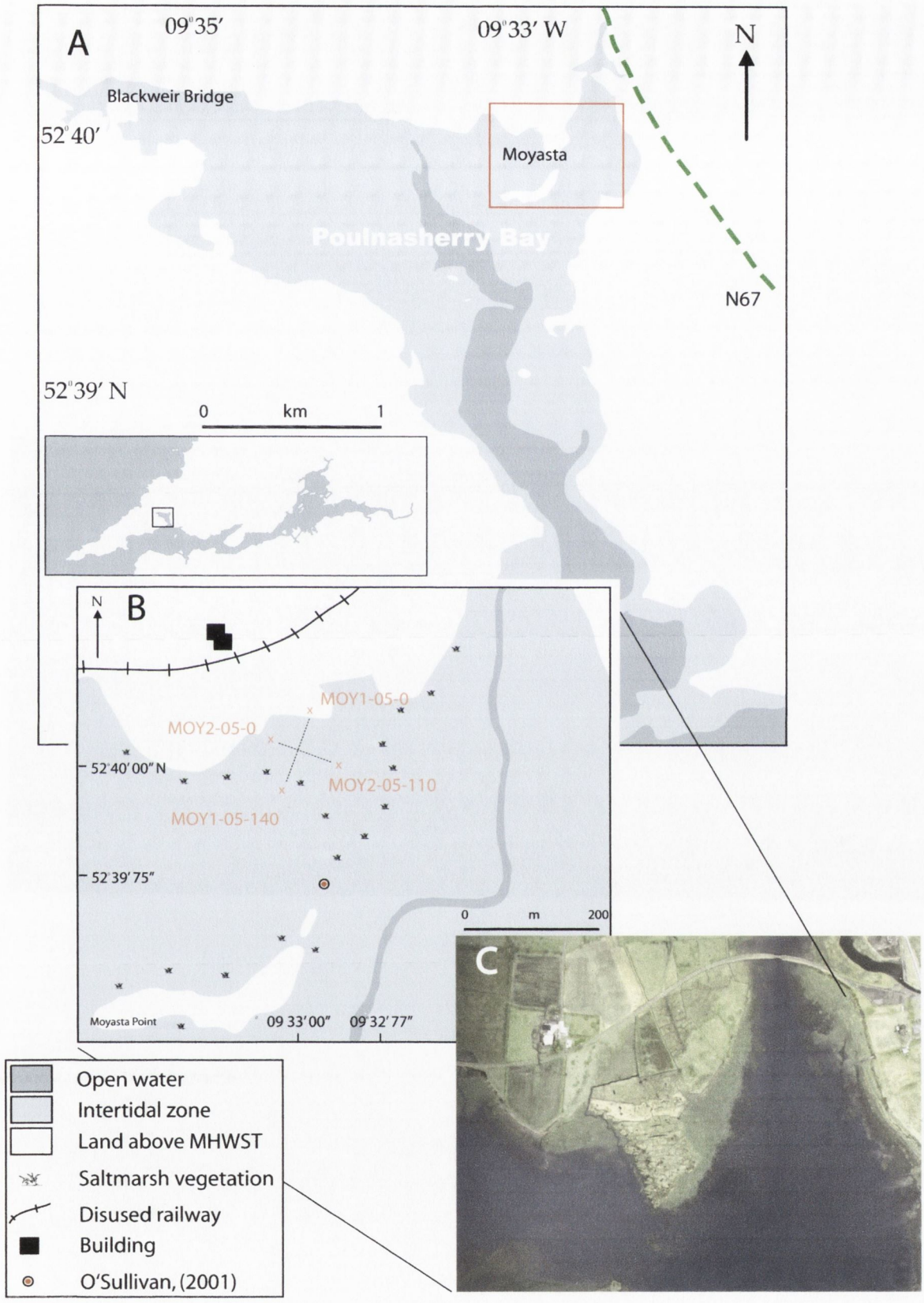


Figure 6.16 Site map showing **A** the location of the Moyasta study site in relation to Poulnasherry Bay; **B** the location of the transects at Moyasta; **C** aerial photograph of the study site



Plate 6.4 The study site at Moyasta (looking southwest)

Photo taken August 2005

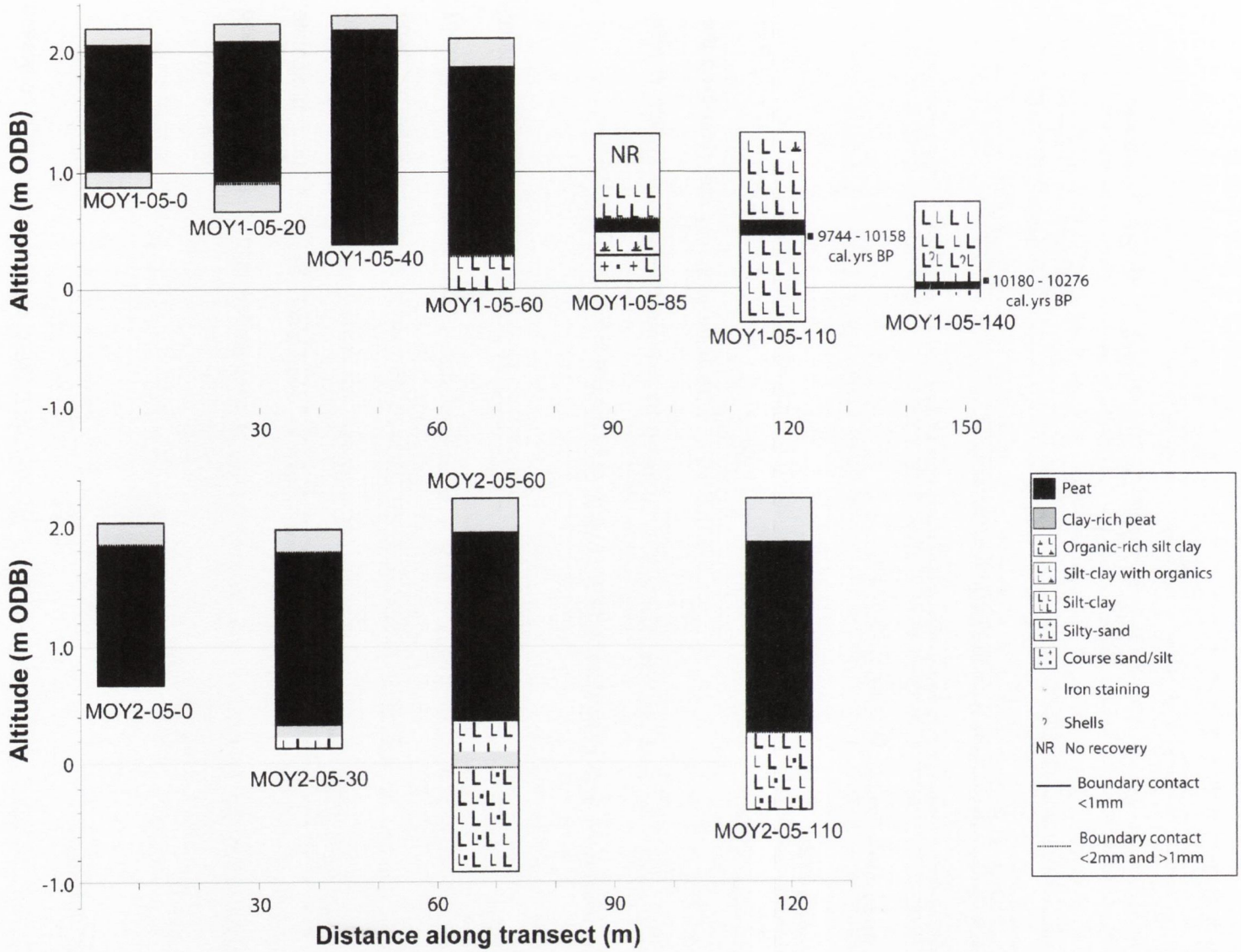


Figure 6.17 Summary lithostratigraphic diagram for transects MOY1 and MOY2 at Moyasta. (See figure 6.16B for transect locations)

c) Chronostratigraphy

Sediment samples taken from the top and bottom of the peat unit found in core IMG3-05-25 were submitted for radiocarbon dating. These organic sediment samples returned ages of 3273-3470 (Beta-211392) and 2370-2743 (Beta-211393) cal. years BP respectively (CALIB 5.0.1; 95% confidence limits) (Table 6.5; Appendix V).

Site	Laboratory Code	Core	Material	Depth (cm)	Altitude (m ODB)	Weight (g)	^{14}C Age	Calibrated Age BP
Islandmagrath	Beta 211393	IMG3-05-25	Peat	240	0.21	3.5	2510±40	2370-2743
Islandmagrath	Beta 211393	IMG3-05-25	Peat	218	0.43	3.0	3170±40	3273-3470

Table 6.5 Radiocarbon dates from the Fergus estuary

6.1.5 Study Sites From the Outer Shannon

6.1.5(i) Moyasta

The site at Moyasta is located on an eroding peat headland in the northeast of Poulnasherry Bay (Figure 6.16a and c). This headland is overtopped during high-water of spring tide, promoting the growth of halophytic plant species. The site is bordered by salt-marsh vegetation which is also present around the (peat) island immediately to the south (Figure 6.16b and Plate 6.4).

Previously, O'Sullivan (2001) has documented the existence of inter-bedded clastic and organic sediment horizons at this site (Figure 6.16b): A radiocarbon date from a submerged (Scots pine) tree root returned an age of 5603-5842 cal. BP (GrN-20145) whilst a date from the overlying *Sphagnum* peat unit revealed an age of 1413-1687 cal. BP (GrN-21928). However, the dated samples have not been related to ordnance datum or even been assigned an approximate elevation within the tidal frame and so they have not be included in the sea-level database (Chapter 4).

a) Lithostratigraphy

Eleven cores were taken from the study site and together, these delivered a spatially consistent and explainable pattern of change (Figure 6.17). Core depth ranges from over 2 m along transect MOY2 to less than a metre at the seaward end of transect MOY1. The altitude of the underlying bedrock is generally lower towards the seaward end of the transects with core MOY1-05-110

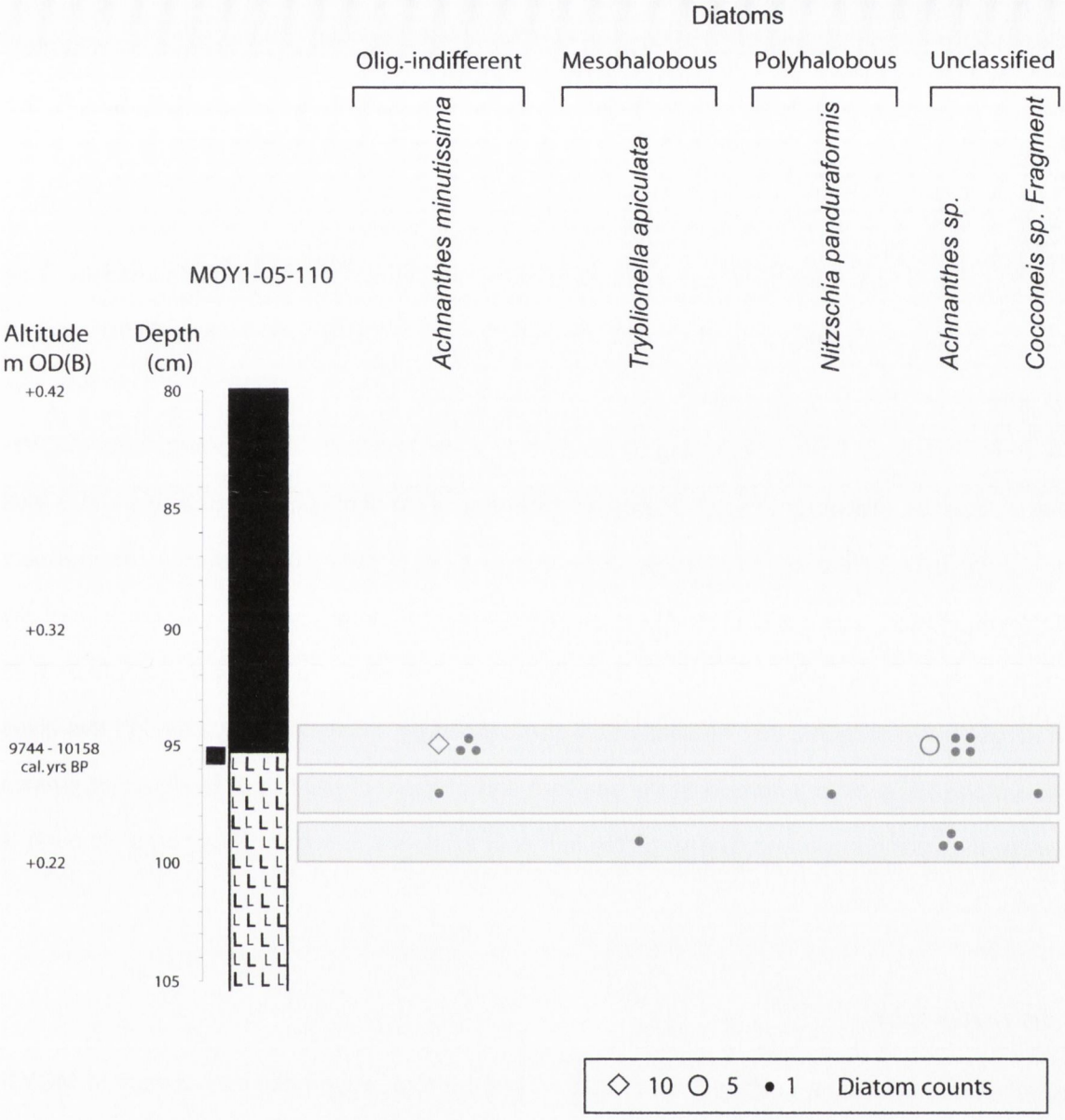


Figure 6.18 Moyasta fossil diatoms from core MOY1-05-110 with associated lithostratigraphy and dated horizons. (Analyst: Caroline Hillier)

terminating on bedrock at an altitude of c.-0.4 m. The substrate encountered at the base of the cores is variable along both of the transects: At the landward end of transect MOY1-05 and in core MOY2-05-0, the peat (or clay-rich peat) rests directly upon bedrock (or impenetrable substrate). However, in cores MOY1-05-60 to MOY1-05-140 and in cores MOY2-05-30 to MOY2-05-110 the peat lies upon a clastic substrate which is often heavily consolidated.

In each core a single, dark, humified peat unit is present although thickness varies spatially. This peat unit is commonly encountered at altitudes between +2 m OD(B) and +0.5 m OD(B) and no peat was found below 0 m OD(B). At the landward edge of the site (*MOY1-05-0*), this unit measures over 1m in depth, increasing to nearly 2 m a short distance along the transect (*MOY1-05-40*). However, seaward of MOY1-05-60, the thickness of the peat unit falls markedly and is less than 20 cm thick at the southern edge of the transect (MOY1-05-140). In the second transect (*MOY2-05*), which runs parallel to the coast, the thickness of this peat unit is more spatially consistent, measuring between 1-1.5 m in depth. At both the landward end of transect MOY1-05 and across transect MOY2-05, the peat unit is overcapped by a clay-rich peat although this gives way to a clay/silt sediment horizon at the seaward end of transect MOY1-05. Evidence of erosion between the peat horizon and overlying clay unit is in existence in some of the cores although is very localized in extent.

b) Biostratigraphy

Foraminiferal and diatom analyses were used across the regressive peat–clay contact in MOY1-05-110 and across the transgressive contact in MOY-05-140 to establish the environment of deposition of the sedimentary units, and to clarify the nature of the transition between them. In both of theses cores, no foraminifera are present and a similar absence of diatoms is encountered in samples from MOY1-05-140. In MOY1-05-110 however, very low counts of diatoms were obtained and these are displayed in Figure 6.18. The most common species is *Achnanthes minutissima*, which has been classified as oligohalobous-indifferent (de Wolf, 1982; Vos and de Wolf, 1993). This species is most abundant within the peat unit. A single count of the polyhalobous species *Nitzschia panduraformis* was retrieved at a depth of 97 cm from within the silt-clay horizon whilst a number of *Achnanthes sp.* are also found in both the peat and underlying silt-clay unit. However,

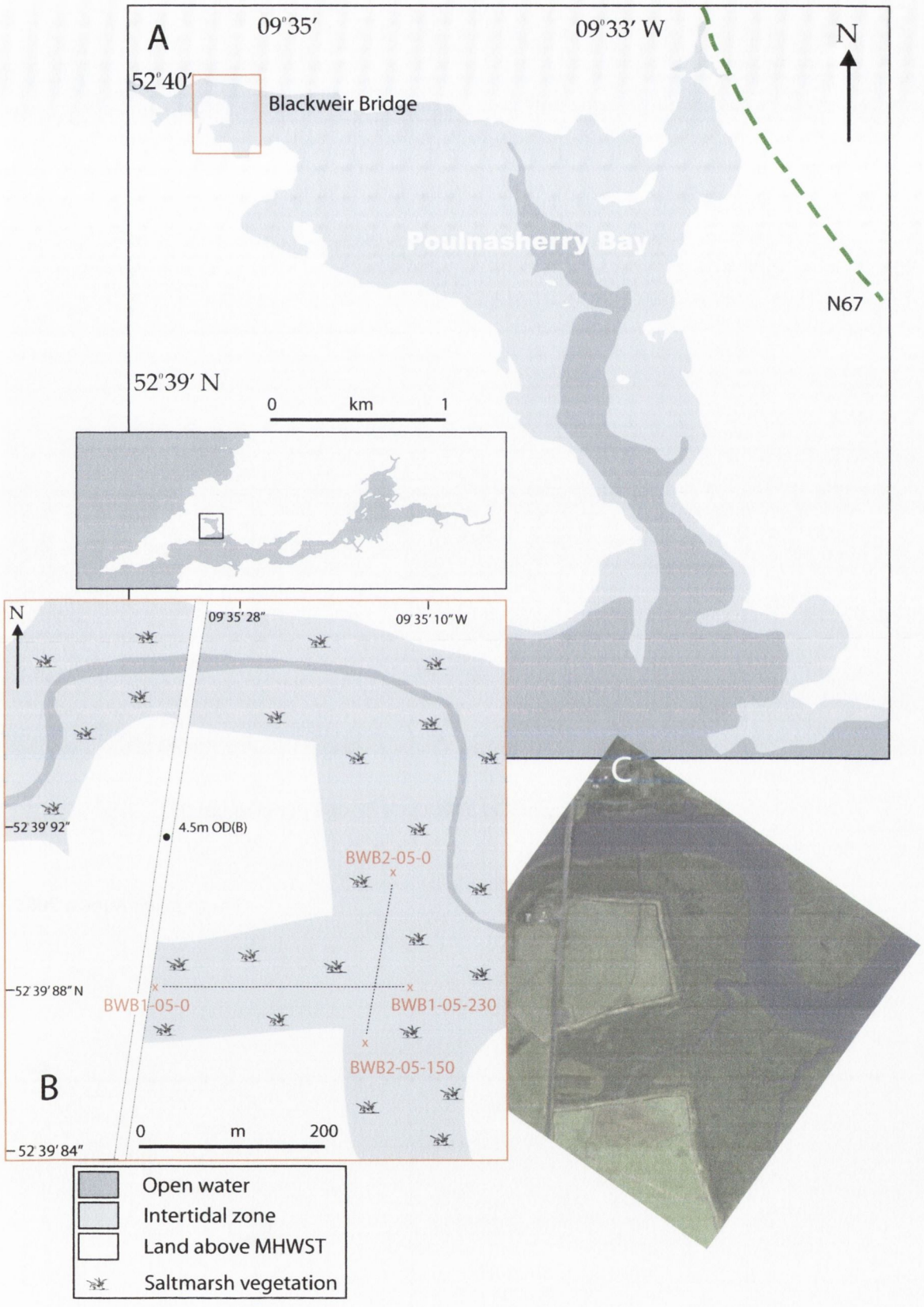


Figure 6.19 Site map showing **A** the location of the Blackweir Bridge study site in relation to Poulnasherry Bay; **B** transect location at Blackweir Bridge; **C** aerial photograph of the study site



Plate 6.5 The study site at Blackweir Bridge (looking east)

Photo taken August 2005

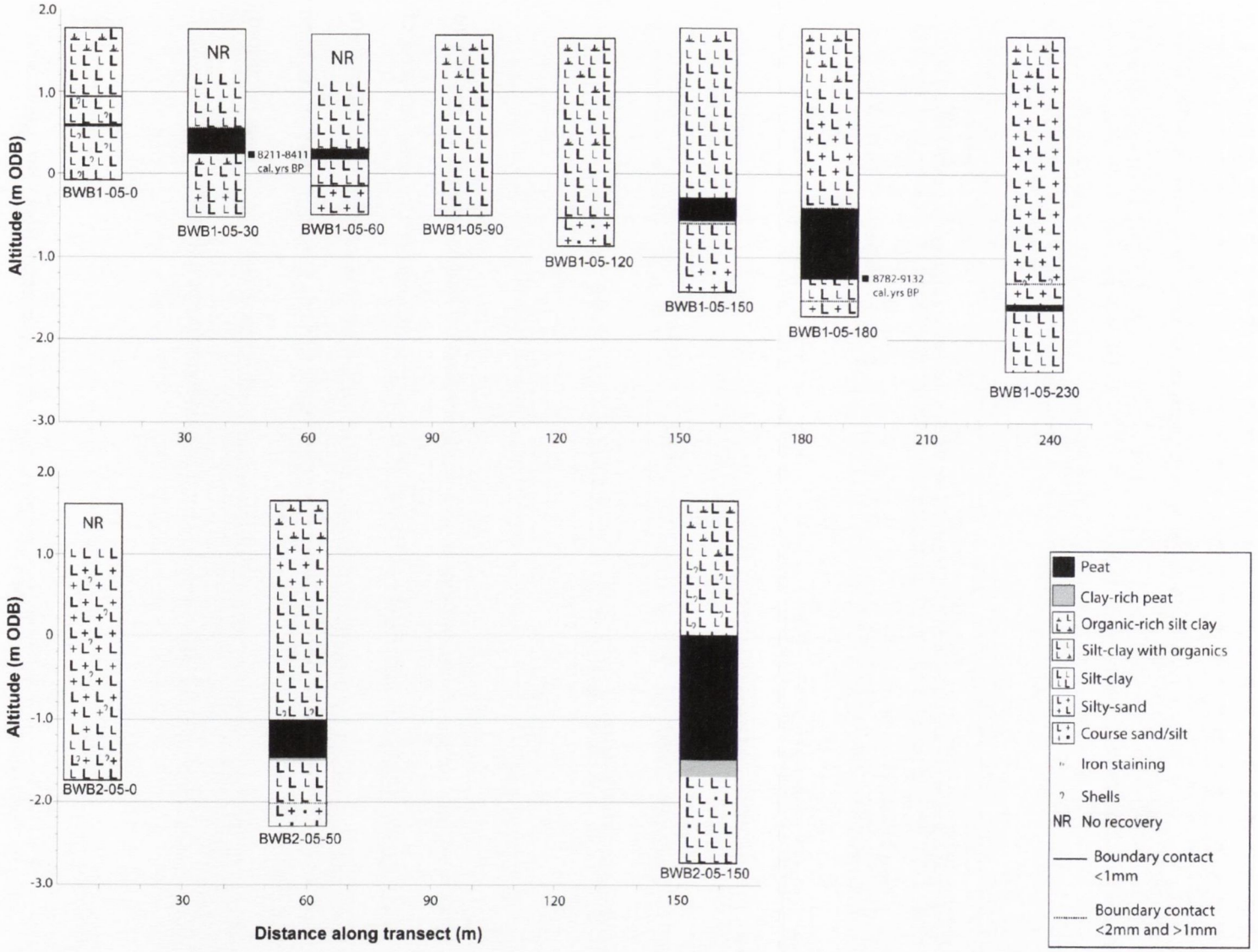


Figure 6.20 Summary lithostratigraphic diagram for transects BWB1 and BWB2 at Blackweir Bridge. (See figure 6.19B for transect locations)

these can not be identified to the species level and hence are unable to provide information on the salinity regime.

c) Chronostratigraphy

Sediment samples from the regressive contact in MOY1-05-110 and transgressive contact in MOY1-05-140 yielded abundant organic material for radiocarbon dating. These samples delivered ages of 9744-10158 (Beta-216643) and 10180–10276 (Beta-216644) cal. years BP respectively (CALIB 5.0.1; 95% confidence limits) (Table 6.6; Appendix V).

Site	Laboritory Code	Core	Material	Depth (cm)	Altitude (m ODB)	Weight (g)	^{14}C Age	Calibrated Age BP
Rahone	Beta 216640	RAH1-05-470	Peat	93	1.26	2.7	4050±40	4421-4798
Rahone	Beta 216641	RAH1-05-480	Peat	27	1.09	3.4	4060±40	4425-4801
Rahone	Beta 216642	RAH1-05-490	Peat	59	0.22	3.3	4880±40	5488-5712
Blackweir Bridge	Beta 216646	PBR1-05-30	Peat	155	0.20	3.5	7530±40	8211-8411
Blackweir Bridge	Beta 216645	PBR1-05-180	Peat	314	- 1.37	3.1	8090±40	8782-9132
Moyasta	Beta 216643	PBAY1-05-110	Peat	95	0.27	2.4	8850±40	9744-10158
Moyasta	Beta 216644	PBAY1-05-140	Peat	70	0.03	3.6	9070±40	10180 - 10276

Table 6.6 Radiocarbon dates from the outer Shannon estuary

6.1.5(ii) Blackweir Bridge

The site at Blackweir Bridge is situated less than 5 km to the west of Moyasta within an expansive salt-marsh system (Plate 6.5 and Figure 6.19a,c). This salt-marsh network occupies an area of over 1km^2 and is bisected by a number of large creeks. Adjacent to the site, thick peat beds are clearly visible although as with many surface outcrops of peat in this region, they have been extensively cut for fuel. Here, the removal of the upper peat surface has encouraged (localized) salt-marsh colonization and this may be clearly seen by aerial photograph (Figure 6.19b and c).

a) Lithostratigraphy

Two transects containing a total of eleven boreholes were recovered from the salt-marsh at Blackweir Bridge (Figure 6.20). These cores record up to c.5 m of organic and minerogenic sediment which variously overlies bedrock or impenetrable substrate. The deepest cores are found along the seaward margin of the site and these commonly extend below -2.0 m OD(B). At the base

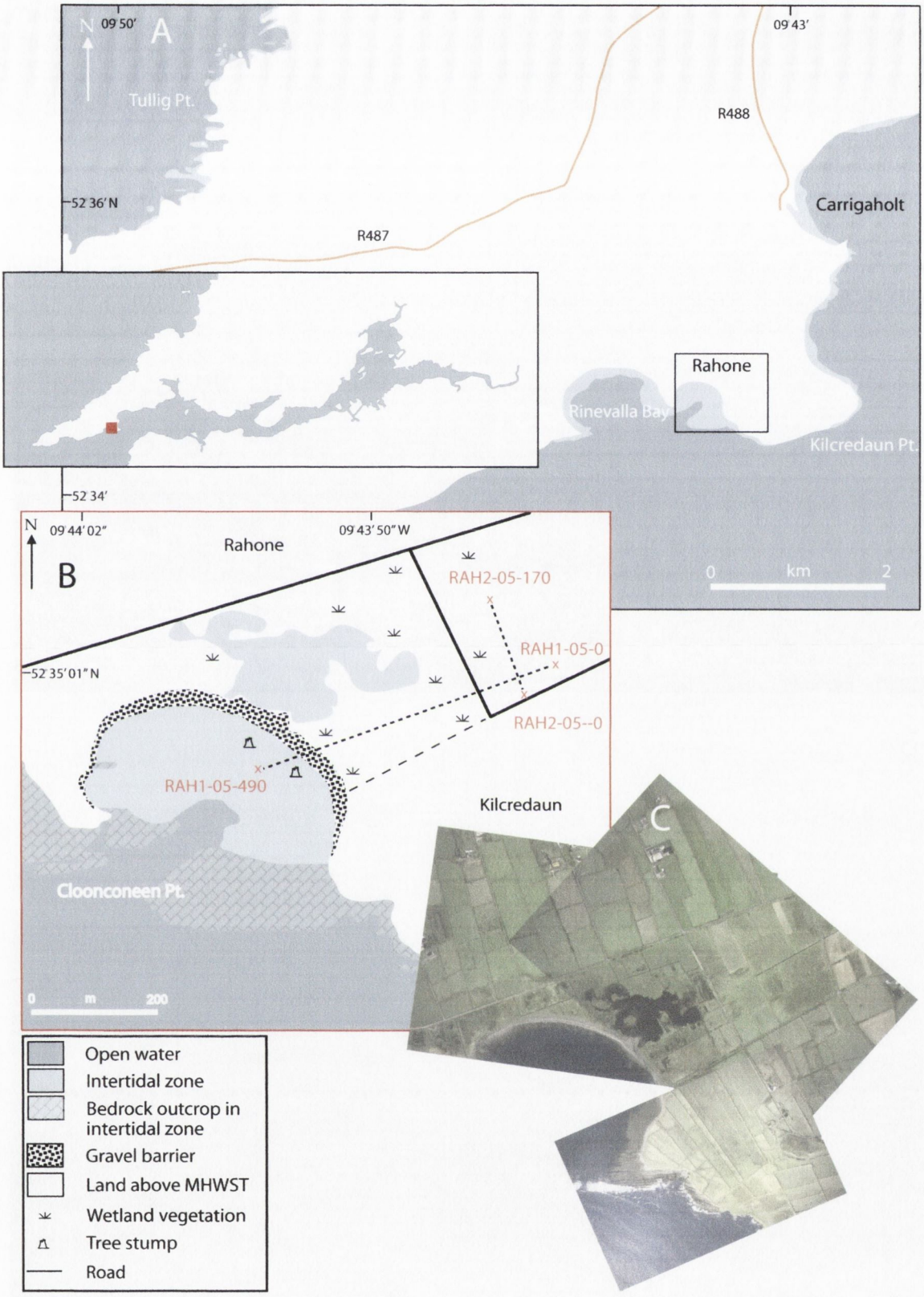


Figure 6.21 Site map showing **A** the location of the study site at Rahone in relation to the outer Shannon estuary; **B** the location of the transects at Rahone; **C** aerial photograph(s) of the study site

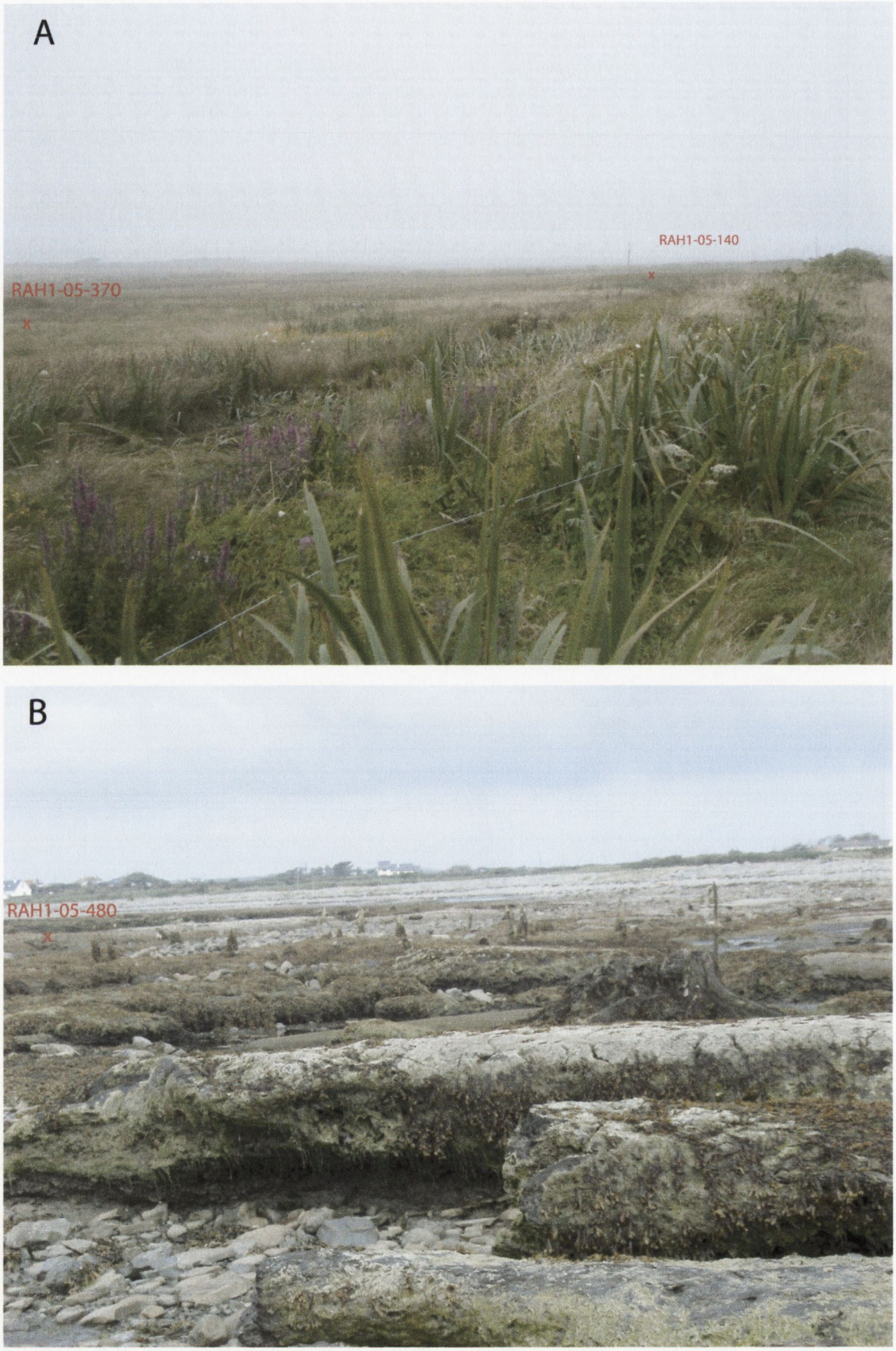


Plate 6.6 The study site at Rahone **A** looking northeast; **B** looking west (at the submerged peat bed and tree stumps)

Photos taken August 2005

of all of the cores is a heavily consolidated clastic sedimentary unit which frequently bears large clasts. In eight of the eleven cores, a compacted, well humified peat unit is found overlying this body of clastic sediment and the peat unit is often separated from the underlying clastic sediment horizon by a thin clay-rich peat section. In both transects, this peat unit is lateral discontinuous and varies in thickness between 5-150 cm. The peat unit also varies in height and is found at progressively lower altitudes as one moves seaward along transect BWB1: In core BWB1-05-0, this peat unit is found at c.+1.0 m OD(B) whilst in core BWB1-05-230 it is evident at -1.6 m OD(B). In all cores, this peat unit is overlain by a silt/clay unit which occasionally contains sand (BWB1-05-230) or shells (BWB1-05-150). Erosion is evident between a number of the sedimentary units and is an almost omnipresent feature of the upper surface of the of the peat unit.

b) Biostratigraphy

Foraminiferal and diatom analyses were undertaken across the regressive peat–clay contacts in BWB1-05-30 and BWB-05-180. However, sediment samples are barren of both microfossil groups.

c) Chronostratigraphy

Sediment samples from the above two regressive contacts were sent for radiocarbon dating analysis. The contact in core BWB1-05-30 returned an age of 8211-8411 (Beta 216646) cal. years BP whilst the contact at 314 cm in core BWB1-05-180 had an age of 8782-9132 (Beta 216645) cal. years BP (CALIB 5.0.1; 95% confidence limits) (Table 6.6; Appendix V).

6.1.5(iii) Rahone

The study site at Rahone is located a short distance to the west of Kilcredaun Point in the outer Shannon estuary (Figure 6.21). The site is fronted by an arcuate shingle barrier and outcrops of bedrock within the inter-tidal zone (Figure 6.21c). This shingle barrier has been breached, causing localized flooding on its landward margin. Halophytic vegetation has developed immediately behind the barrier (adjacent to the lagoon) although for the most part, the study site is covered by grass and used for grazing (Plate 6.6a). The submerged forest bed found in the adjacent Rinevalla Bay had previously attracted the attention of Pearson (1979) and this forest bed is also clearly evident at Rahone (Plate 6.6b).

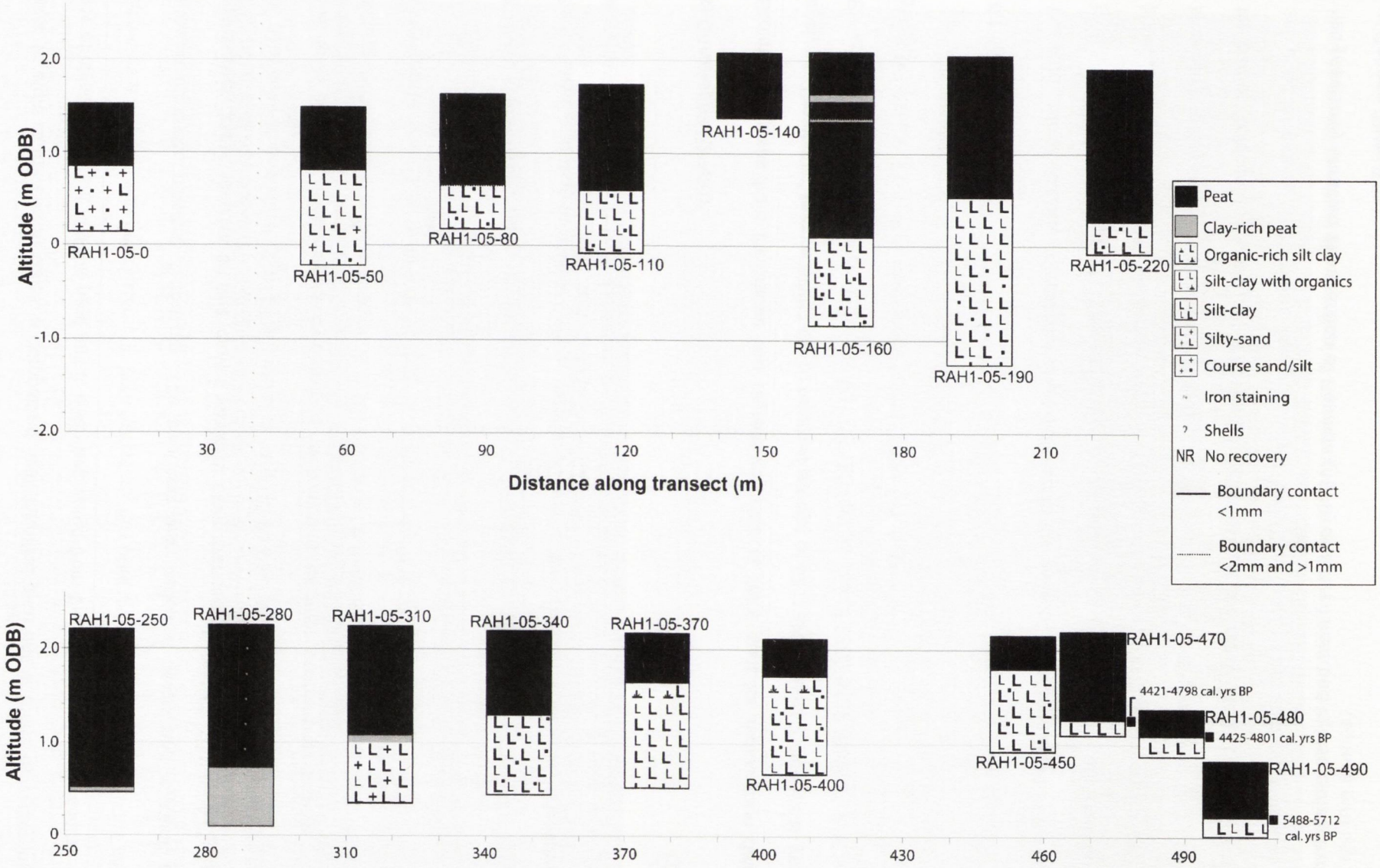


Figure 6.22a Summary lithostratigraphic diagram for transect RAH1 (Rahone). See figure 6.21B for transect locations)

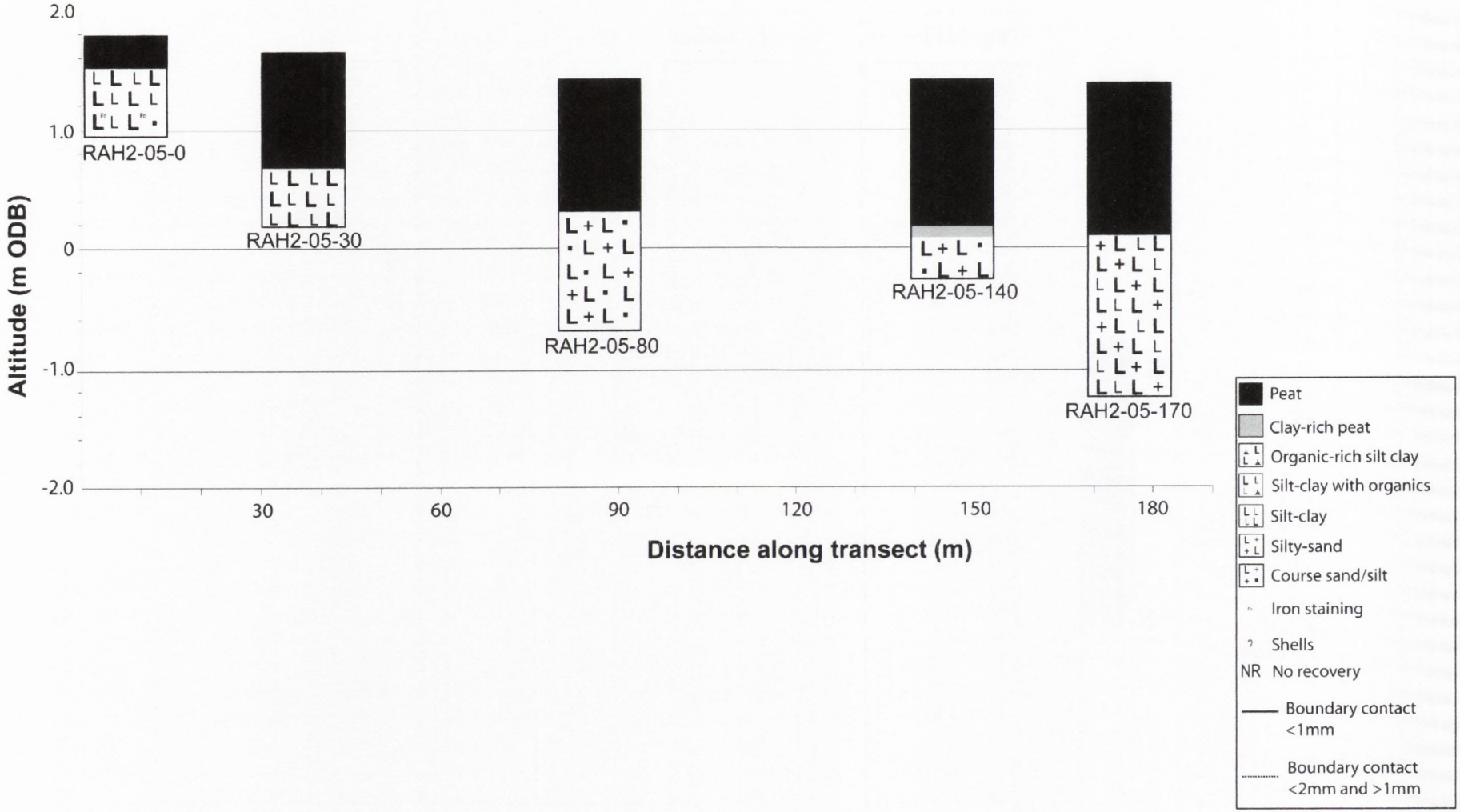


Figure 6.22b Summary lithostratigraphic diagram for transect RAH2 (Rahone). See figure 6.21B for transect locations)

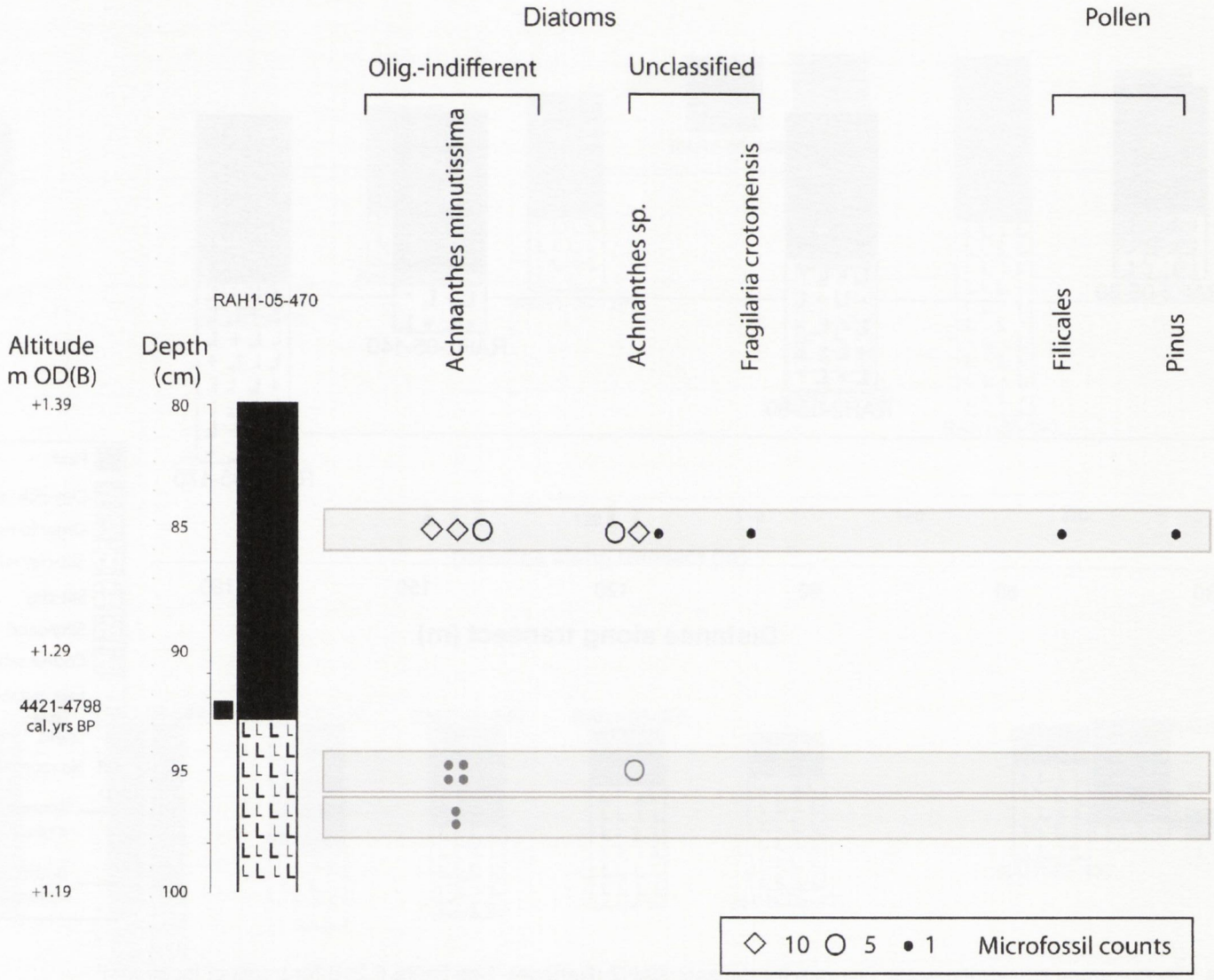


Figure 6.23a Rahone fossil diatoms (and pollen) from core RAH1-05-470 with associated lithostratigraphy and dated horizons. (Analyst: Caroline Hillier)

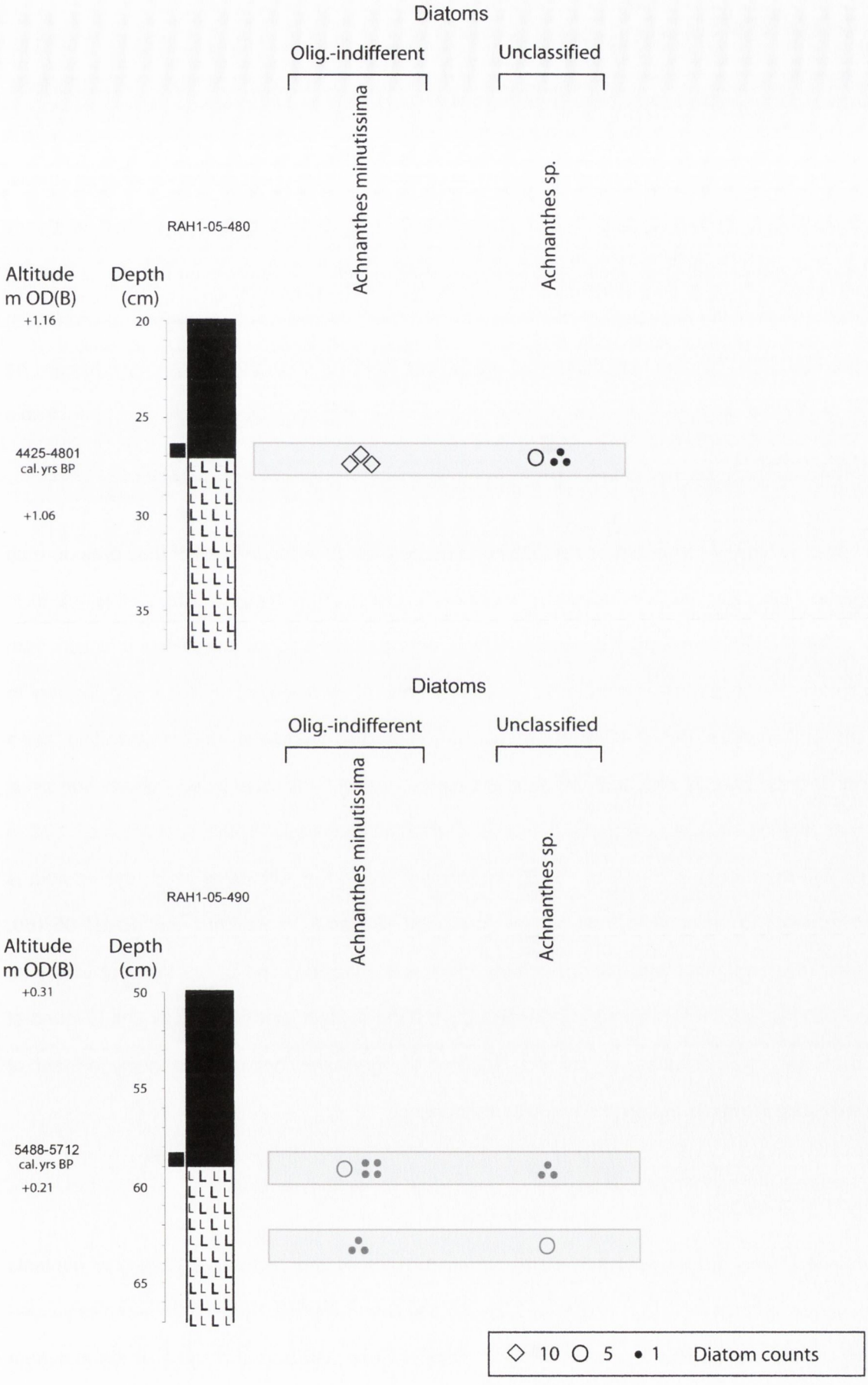


Figure 6.23b Rahone fossil diatoms from core RAH1-05-480 and core RAH1-05-490 with associated lithostratigraphy and dated horizons. (Analyst: Caroline Hillier)

a) Lithostratigraphy

Two transects made up of over twenty cores have been collected from the study site at Rahone. Details of core lithostratigraphy are displayed in Figure 6.22a,b. The stratigraphy of the site is straightforward and each core is essentially characterized by two contrasting sedimentary units. The lower section of the sequence is comprised of a silt-clay sediment association which frequently contains larger clasts. This clastic unit overlies bedrock (or impenetrable substrate) at varying altitudes across both transects, ranging from +1.4 m OD(B) in core RAH1-05-140 to -1.4 m OD(B) in core RAH2-05-170. The high altitude at which core RAH1-05-140 terminates may however, be explained by the presence of recumbent tree trunks within the peat layer which are impenetrable by auger corer.

The upper section is comprised of partially humified peat which is found in all of the cores on both transects. This peat unit is however, of variable thickness: at its thickest, it measures c.2 m in depth (RAH1-05-160) although it thins in both a northeasterly and southwesterly direction from core RAH1-05-160. On the foreshore (at the seaward end of the transect), this peat unit is closer to 50 cm in thickness (RAH1-05-490). In all of the cores, this peat unit is underlain by a (predominantly) silt-clay sediment unit although the contact with this clastic sedimentary horizon is found at differing altitudes: the highest altitude at which this regressive contact occurs is c.+1.75 m OD(B) and the lowest is c.+0.2 m OD(B). In transect RAH1, the altitude at which this contact is found appears to increase as one moves north east and south west from core RAH1-05-160. However, this straightforward pattern is interrupted at the western end of the transect where in, cores RAH1-05-470, RAH1-05-480 and RAH1-05-490 the contact with the silt/clay unit is found at progressively lower altitudes. In transect RAH2, the regressive contact is generally evident at higher altitudes towards the southern end of the transect.

b) Biostratigraphy

Foraminiferal and diatom analyses were carried out across the regressive peat–clay contacts encountered in cores RAH1-05 470, RAH1-05-480 and RAH1-05-490. None of the samples yielded any foraminifera tests and only low counts of diatoms were discovered in some of the sediment samples. These are displayed in Figure 6.23a,b. By far the most common diatom species is *Achnanthes minutissima* which has been classified oligohalobous-indifferent (de Wolf, 1982; Vos

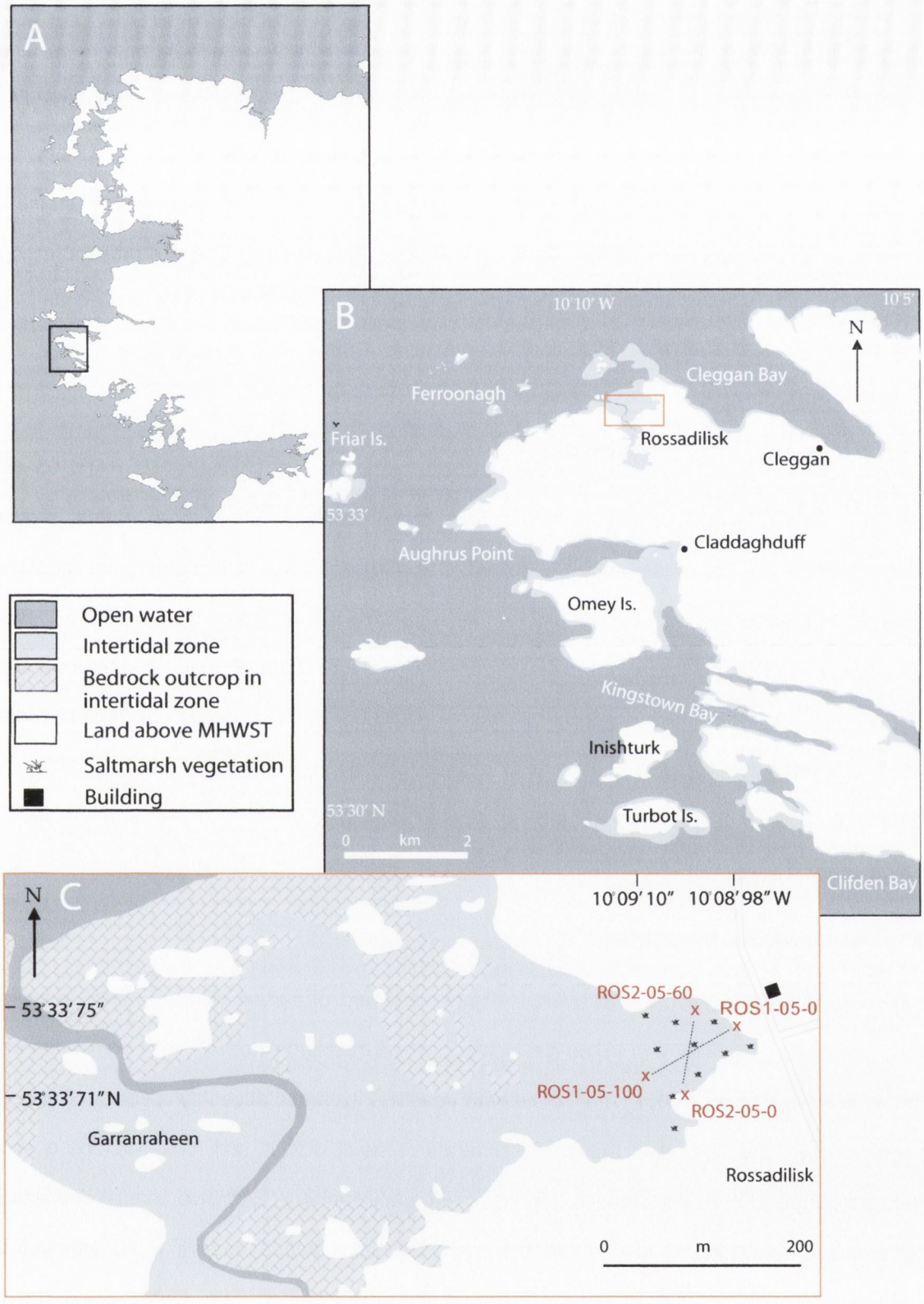


Figure 6.24 Site map showing **A** Rossadilisk in a regional context; **B** the location of the study site in west Connemara; **C** the position of the core transects

and de Wolf, 1993). This species is encountered in samples from each of the three cores: In cores RAH1-05-480 and RAH1-05-490, its presence is confined to the silt-clay unit and the organic-clay unit on the (regressive) contact where up to 30 counts have been recorded. In core RAH1-05-470, this diatom species is also observed within the peat unit. A number of *Achnanthes* diatom species are also present although they cannot be identified to the species level and are therefore unable to be employed as indicators of salinity. Pollen grains from both fern and pine are present in core RAH1-05-470 although only single counts of each taxa are found.

c) Chronostratigraphy

Sediment samples from the three regressive contacts found in cores RAH1-05-470, RAH1-05-480 and RAH1-05-490 were sent for radiocarbon dating analysis. The three dates proved chronologically coherent and are shown in Table 6.6 and Appendix V. The lithostratigraphic contact in core RAH1-05-470 yielded a date of 4421-4798 (Beta-216640) cal. years BP (CALIB 5.0.1; 95% confidence limits) whilst a date of 4425-4801 (Beta-216641) cal. years BP was returned from core RAH1-05-480. Finally, the regressive contact in core RAH1-05-490 was dated to 5488-5712 (Beta-216642) cal. years BP.

6.2 CONNEMARA

6.2.1 Rossadilisk Site Description

Connemara is situated on the high-energy Atlantic seaboard of Ireland (Figure 6.24a), a region where topography, together with linked geological controls, result in extensive rock dominated and cliffed coastlines (Devoy, 2000). The Connemara coastline is characterized by a series of narrow, topographically controlled inlets along with a series of small islands and rock outcrops a short distance offshore to the west. Despite the high energy environment, extensive coastal deposits of Holocene age have been mapped in this region (e.g. Jessen, 1949; O'Connell, 1988), although by far the most common of these deposits are sand and gravel barriers (e.g. Delaney and Devoy, 1995; Devoy *et al.*, 1996).

Whilst, the geology of Connemara as a whole is dominated by Dalradian metasediments (e.g. schists, psammites and marbles), Rossadilisk is situated on a small extension of Siluro-Devonian granite (Graham, 2005). Central western Ireland was extensively glaciated at (and probably well



Figure 6.25 Drumlins and associated Last Glacial phenomena in Connemara. (Taken from Orme, 1967)



Plate 6.7 The study site at Rossadilisk **A** looking southwest; **B** looking northeast

Photos taken August 2005

before) the last Glacial maximum (e.g. Bowen *et al.*, 2002) although the glacial stratigraphy of the region currently lacks geochronological control (Farrell *et al.*, 2005). During the last glaciation, the site was ice covered with ice moving in a general east-west direction (e.g. Synge and Stephens, 1960; Finch, 1977). Much of the land in the immediate vicinity of the site is characterized by exposed bedrock, sometimes with a thin topsoil cover although till and drumlins are present nearby (Figure 6.25).

	HAT	MHWST	MHWNT	MTL	MLWNT	MLWST	LAT
Clifden Bay	-	1.7	0.7	-	-	-	-
Bofin Harbour	1.9	1.4	0.4	-0.8	-1.1	-2.2	-2.7
Average	-	1.55	0.55	-	-	-	-

Table 6.7 Summary tidal characteristics for the site at Rossadilisk, Connemara

The salt-marsh found at Rossadilisk has been described as fringing in nature, overcapping a peat/sand substrate (Curtis and Sheehy-Skeffington, 1998). Although this region is exposed to high wave energy (Orford, 1989; Carter, 1990; Devoy, 1992) and macro tidal conditions (Table 6.7), the site is reasonably well sheltered: To the south west of Rossadilisk, the Aughrus peninsula provides protection from prevailing wave attack whilst to the west, a series of small islands (Ferroonagh west, Ferroonagh east) help dissipate wave energy (Figure 6.24b). At the site-scale, the entrance to the inlet at Rossadilisk is characterized by a number of exposed rock platforms which provide a further buffer to wave attack (Figure 6.24c). This protection has created a (localized) low energy environment in which halophytic vegetation has been able to develop (Plate 6.7a + b). However, the thin veneer of sand over much of the salt-marsh suggests that despite the protection to the south and west, during storm events the salt-marsh is overtopped by waves.

Anthropogenic influence on the Connemara coastline is minor in comparison to many other areas of northwest Europe and the region has been described in the past as one of the most desolate in Ireland (Jessen, 1949). The area surrounding Rossadilisk is sparsely populated with the nearest village (Cleggan) 3km to the south east. However, evidence of prehistoric settlement in northwest Connemara is abundant and dates back to the Mesolithic (Gibbons and Gibbons, 2005). By the early Neolithic period, widespread settlement existed, evidenced by a series of Megalithic tombs in the nearby Cleggan-Ballynakill Bay area.

Limiting Date Tier	Region	Site	Site Location (Latitude)	Site Location (Longitude)	Labcode	C-14 Age BP	‰ C-13	Calibrated Age BP (2σ)	Midpoint	±	Altitude m OD(Belfast)	Material	Microfossil Evidence	Extent of Marine Influence	Tidal port 1	Tidal port 2	HAT	MHWST	MHWNT	MTL/MSL	MLWNT	MLWST	LAT	Indicative Meaning	Max/Min Height of msl	Error Term (m)	Reference
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-56064	100.6		Invalid			1.52	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	0.28	0.3	Delaney and Devoy, (1995)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53296	104.7		Invalid			1.63	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	0.39	0.3	Delaney and Devoy, (1995)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53927	10±60		0-268	134	134	1.58	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	0.34	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53926	120±70		0-284	142	142	1.44	Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	0.2	0.3	Delaney and Devoy, (1995)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-56066	230±60		0-459	229.5	230	1.04	Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.2	0.3	Delaney and Devoy, (1995)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53929	420±70		308-543	425.5	118	1.12	Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.12	0.3	Delaney and Devoy, (1995)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-56065	600±50		533-662	597.5	65	1.1	Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.14	0.3	Delaney and Devoy, (1995)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-56067	760±50	-25	570-782	676	106	1.1	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.14	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-56069	1250±60	-25	1016-1293	1154.5	139	0.51	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.73	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53931	1290±80	-25	1008-1343	1175.5	168	0.2	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-1.04	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-56068	1260±60	-25	1060-1294	1177	117	0.55	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.69	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-56070	1370±60	-25	1175-1386	1280.5	106	0.46	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.78	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53930	1620±80	-25	1351-1701	1526	175	0.3	Sandy Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-0.94	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53932	2970±100	-25	2876-3372	3124	248	-0.73	Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-1.97	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53933	3610±80	-25	3699-4147	3923	224	-0.83	Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-2.07	0.3	Devoy et al., (1996)
1	11	Carrownisky, (Mayo)	53.71	-9.90	Beta-53934	3990±80	-25	4163-4810	4486.5	324	-1.12	Peat		MHWST	Clare Island	Killary Harbour		1.51	0.56	0.27	-0.99	-2.09		1.24	-2.36	0.3	Devoy et al., (1996)
1	11	Roundstone, (Galway)	53.4	-9.9	(Pollen)	11000-12000		13000-14000	13500	500	-3.5	Peat	Pollen	MHWST	Roundstone			1.5	0.5	-0.4	-1.2	-2.4		1.9	-5.4	1	Jessen, (1949)

Table 6.8 Existing sea level data from Connemara (database region 11)

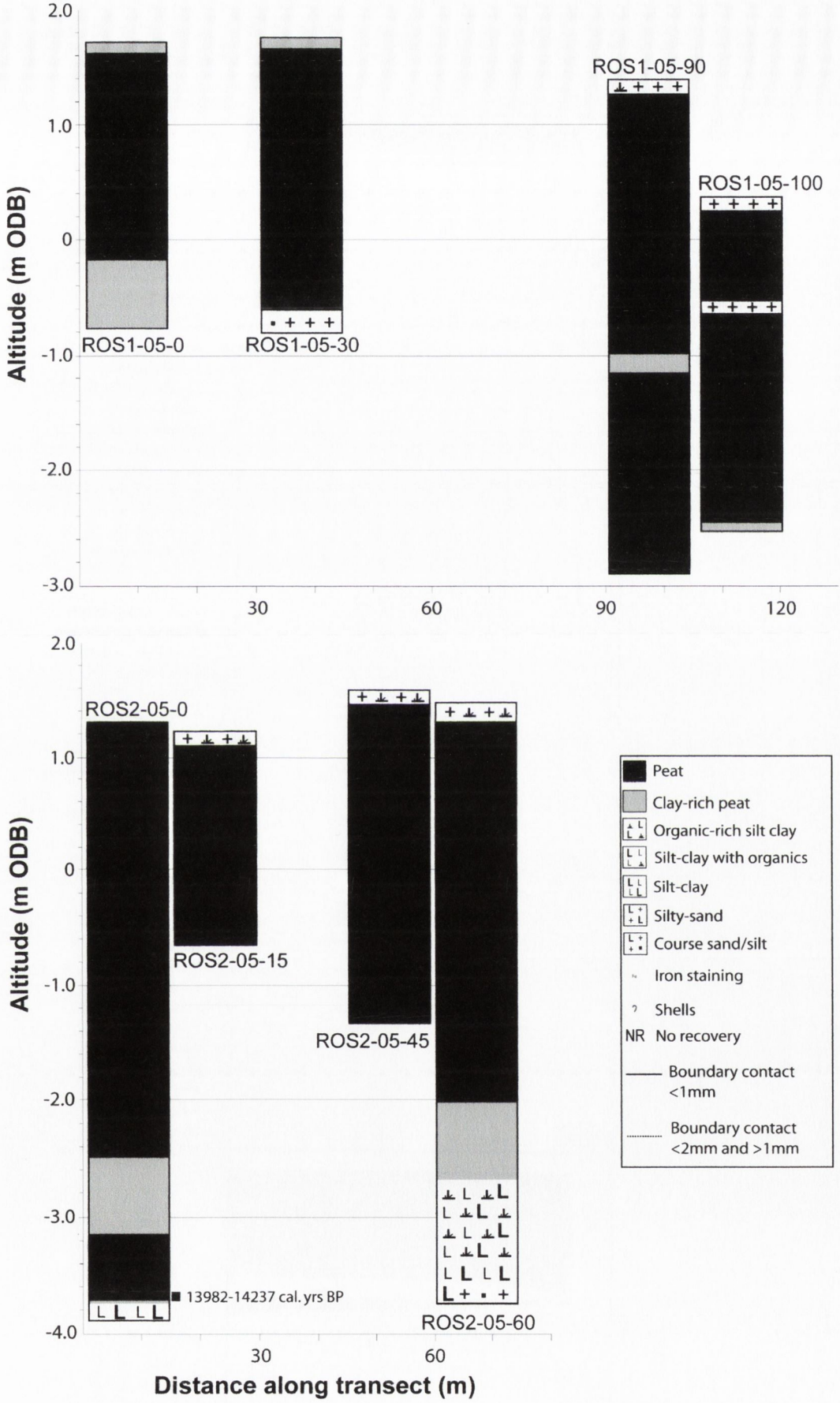


Figure 6.26 Summary lithostratigraphic diagram for transects ROS1 and ROS2 at Rossadilisk. (See figure 6.24C for transect locations)

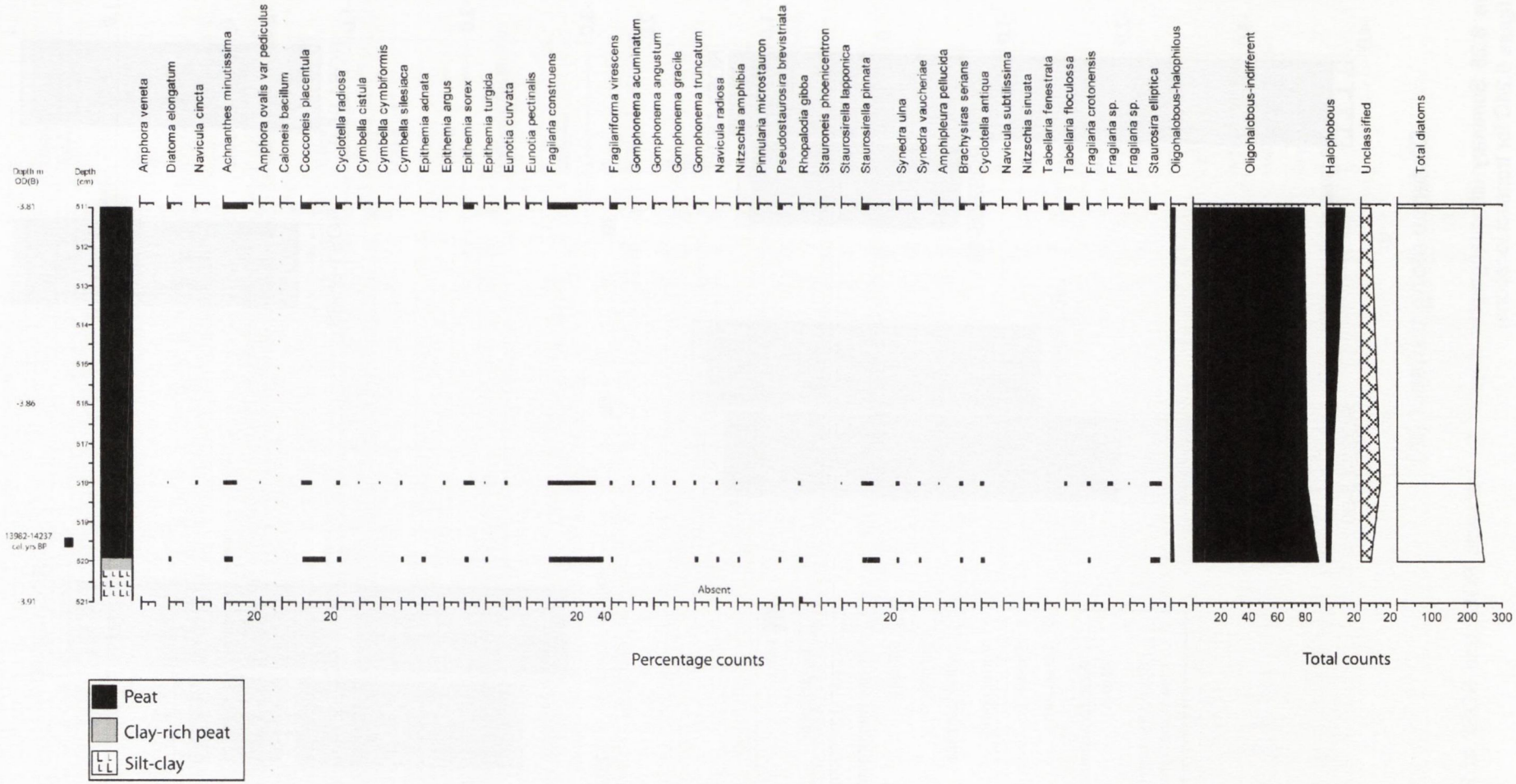


Figure 6.27 Rossadilisk fossil diatoms from core ROS2-05-0 with associated lithostratigraphy and dated horizons. (Analyst: Caroline Hillier)

Along this stretch of coastline, movements of sea level during the postglacial are largely unknown (Chapter 2). All available sea level data are limiting in form and may be viewed in Table 6.8.

6.2.2 Rossadilisk Stratigraphic Characteristics

6.2.2(i) Lithostratigraphy

At Rossadilisk, eight cores were taken from two transects orientated across the salt-marsh and adjacent inter-tidal flat. Details of these cores are shown in Figure 6.26. All of the cores terminate on bedrock and evidence very uneven basal topography. The cores are all dominated by a dark peat unit which varies in thickness from c. 1.5 m (ROS1-05-0) to almost 5 m (ROS2-05-0). With the exception of cores ROS1-05-30, ROS2-05-0 and ROS2-05-60, these peat units overlie bedrock which is encountered at an altitude of between +1.3 and -2.9 m OD(B) below these cores. In cores ROS1-05-30. ROS2-05-0 and ROS2-05-60, this peat unit is underlain by a clastic unit although organic content and particle size varies between cores. There is no consistent trend in core depth along either of the transects, emphasizing the undulating nature of the bedrock at the site. The peat unit itself displays subtle variation in its degree of humification and is generally more humified with depth. At no point does it become well humified and wood fragments along with plant macrofossils are present. These peat units are nearly all capped by a (surface) veneer of organic sand which measures c.10 cm in thickness.

6.2.2(ii) Biostratigraphy

Foraminiferal and diatom analyses were undertaken across the regressive contact found in ROS2-05-0. No foraminifera were found in the core although high concentrations of diatoms were present in samples taken from the peat section and underlying clay-rich horizon. Counts in excess of c.200 valves were obtained from samples contained within these two sediment units although samples taken from the basal light grey clay unit appear barren of diatoms (Figure 6.27). All of the samples taken from core ROS2-05-0 have a high concentration of the oligohalobous-indifferent diatom species *Fragilaria construens*. This species accounts for 20-40% of the sample population whilst oligohalobous-indifferent forms commonly make up c.80% of the total diatom population in all samples. Within the peat unit found at a depth of 511 cm, the diatom assemblage also contains c.15% halophobous species and these species are indicative a (totally) freshwater environment (Table 3.1). However, halophobous species become less abundant with depth and comprise less

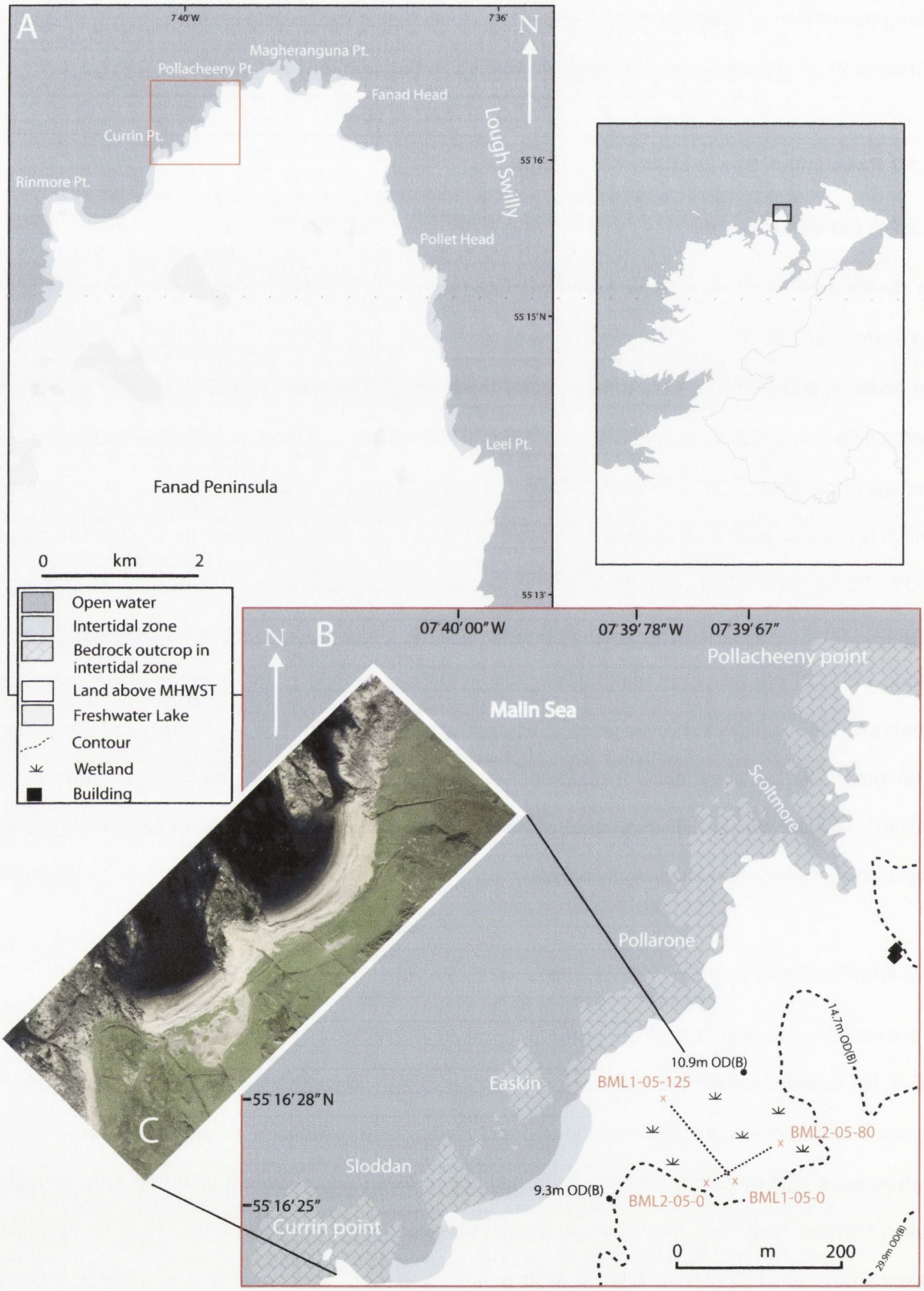


Figure 6.28 Site map showing **A** the study site at Ballymichael in relation to the Fanad Peninsula; **B** the location of the core transects; **C** aerial photograph of the study site

than 5% of the total diatom assemblage in the clay rich peat unit encountered at –5.20 m. Neither mesohalobous nor polyhalobous diatoms (which are indicative of saline conditions) are present in any of the samples taken from core ROS2-05-0.

6.2.2(iii) Chronostratigraphy

A sediment sample from the regressive contact in ROS2-05-0 was sent for radiocarbon dating analysis and returned an age of 13982-14237 cal. years BP (Beta 211391) (CALIB 5.0.1; 95% confidence limits) (Appendix V).

6.3 DONEGAL

"Northwest Ireland is an area of particular interest to students of sea-level variations as it lies within a area especially sensitive to both short – (10^1 to 10^2 years) and long-term (10^3 to 10^4 years) environmental changes" (Shaw and Carter, 1994; pp 74).

HAT	MHWST	MHWNT	MTL	MLWNT	MLWST	LAT
1.85	1.33	0.33	-0.41	-1.17	-2.17	-2.7

Table 6.9 Summary tidal characteristics for the site at Ballymichael, Donegal

6.3.1 Ballymichael Site Description

The small hamlet of Ballymichael is situated on the north coast of the Fanad peninsula in north east Donegal (Figure 6.28a). Donegal forms a distinct area of high terrain with an irregular, deeply embayed coastline (Stephens, 1970). Much of the coastline is bedrock controlled and is characterized by several major sea loughs including fjards (e.g. Lough Foyle) and fjords (Lough Swilly) (Shaw and Carter, 1994). Extensive beach ridge and dune systems are also commonly found along the coastline of north west Ireland (e.g. Horn Head, Malin Head, Magilligan) and these have been considered in a number of recent investigations (e.g. Carter and Wilson, 1990; Wilson and Braley, 1997; Carter, 1982b). The northwest Ireland seaboard is one of the most exposed and windiest parts of Europe (Troen and Petersen, 1989) with mean annual wind speed at nearby Malin head exceeding 8 m s^{-1} (Rohan, 1986). On exposed stretches of coastline (such as the north Fanad coastline), wave energy is high: The 50 year extreme of wave height is 30 m (Orford, 1989; Carter, 1990) and the wave climate is dominated by North Atlantic swell waves which have a

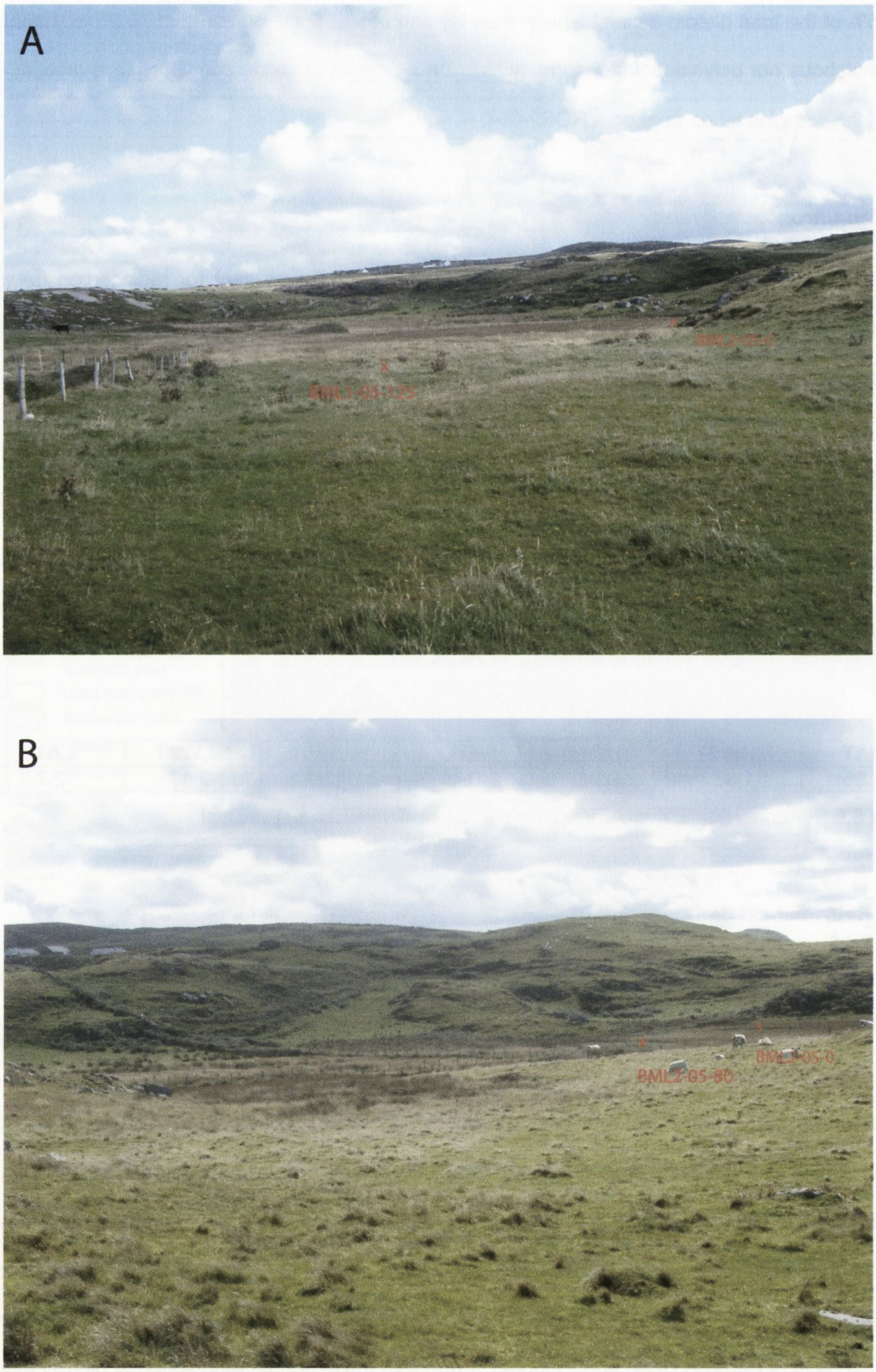


Plate 6.8 The study site at Ballymichael **A** looking southeast; **B** looking southwest)

Photos taken August 2004

strong westerly component in their approach direction (Shaw, 1985). Tides in this region are meso to macro in nature with the mean spring range of (the semi-diurnal) tides being 3.5 m at Fanad Head (Table 6.9, Figure 6.28).

This region of north Donegal falls within the NE-SW trending Dalradian metasediments of Precambrian age which include quartzite, pelite and granite surface rock types (Pitcher and Berger, 1972). These rocks experienced extensive glacial scour during the Pleistocene glaciations which engulfed this region leaving a largely ice-eroded surface in many parts of Donegal (Carter and Wilson, 1993). However, in this region, movements of ice during the last glaciation are imperfectly understood. The extent to which ice extended beyond the current coastline has been debated (e.g. Stephens and Synge, 1965; Evans *et al.*, 1980; McCabe and Clark, 2003) (see section 5:5.1) although the most recent estimates for ice thickness suggest the Fanad Peninsula was overridden by approximately 450 m of ice at the LGM (Ballantyne *et al.*, submitted). Deglaciation of northwest Ireland is suggested to have occurred rapidly and relatively early although the region was overridden by ice once again during the major ice readvance episode associated with the Killard Point stadial at c.18 000 cal. yrs BP (McCabe and Clark, 2003; Clark, 2006). The site at Ballymichael is situated at the northerly limit of this readvance.

The site at Ballymichael faces northwest onto the Malin Sea and is partly fronted by a large storm beach composed of coarse clastic material (Figure 6.28c). This beach is buttressed by two exposed rock promontories (Sloddan and Easkin) with the rock outcrop at Easkin providing a degree of protection from (prevailing) westerly wave attack (Figure 6.28b). The basin itself is flanked by relatively steep relief on both its west and south side which partially enclose a tract of marshland dominated by euryhaline vegetation (Plates 6.8a + b). This site is currently used for pastoral farming and moderate alterations have been made to the surrounding landscape. The most significant of these is the construction of a small (<1 m) drainage ditch that bisects the site and passes across the sill of the basin. (This is clearly visible on the aerial photograph seen in Figure 6.28c).

As with the site at Rossadilisk, the local area is very sparsely populated. However, reference to the Ordnance Survey (1840) 1:6 inch map suggests that the Fanad area was more densely populated in the past. Indeed, the abundance of old sea walls, fence-lines and embankments running through

Index Point Tier	Region	Site	Site Location (Latitude)	Site Location (Longitude)	Labcode	C-14 Age BP	% C-13	Calibrated Age BP (2σ)	Midpoint	±	Altitude m OD(Belfast)	Material	Microfossil Evidence	Reference Water Level	Indicative Range (m)	Tidal port 1	HAT	MHWST	MHWNT	MTL/MSL	MLWNT	MLWST	LAT	Indicative Meaning	RSL (Relative to msl)	Error Term (m)	Reference
1	16	Clonmass	55.153	-7.913	UB-2679	3300±95	-27.7	3348-3824	3586	238	1.24	Saltmarsh Peat	Pollen, Foraminifera	(MHWNT+HAT)/2	0.74	Fanad Head	1.8	1.3	0.3	-0.4	-1.2	-2.2		1.5	-0.24	0.3	Shaw and Carter, (1994)

Limiting Date Tier	Region	Site	Site Location (Latitude)	Site Location (Longitude)	Labcode	C-14 Age BP	% C-13	Calibrated Age BP (2σ)	Midpoint	±	Altitude m OD(Belfast)	Material	Microfossil Evidence	Extent of Marine Influence	Tidal port 1	HAT	MHWST	MHWNT	MTL/MSL	MLWNT	MLWST	LAT	Indicative Meaning	Max/Min Height of msl	Error Term (m)	Reference
1	16	Clonmass	55.167	-7.883	SRR-2399	800±40	-29	672-783	727.5	56	3.96	Organic-rich	Pollen	MHWNT	Fanad Head	1.8	1.3	0.3	-0.4	-1.2	-2.2		0.7	3.22	0.3	Shaw and Carter, (1994)
1	16	Clonmass	55.167	-7.883	SRR-2398	3600±50	-30	3725-4082	3903.5	179	3.41	Organic-rich	Pollen	MHWNT	Fanad Head	1.8	1.3	0.3	-0.4	-1.2	-2.2		0.7	2.67	0.3	Shaw (1985)
1	17	New Bridge	55.143	-7.541	Gif-5483	5810±140		6311-6939	6625	314	-0.1	Peat	Pollen	MHWST	Rathmullen		1.6	0.5	-0.2	-0.8	-2.2		1.8	-1.9	2	Shaw and Carter, (1994)
2	16	Clonmass	55.167	-7.883	SRR-2395	1670±50		1417-1702	1559.5	143	-0.2	Wood	Pollen	MHWST	Fanad Head	1.8	1.3	0.3	-0.4	-1.2	-2.2		1.7	-1.9	?	Shaw (1985)
2	16	Clonmass	55.167	-7.883	SRR-2396	3990±70		4239-4803	4521	282	1.5	Wood	Pollen	MHWST	Fanad Head	1.8	1.3	0.3	-0.4	-1.2	-2.2		1.7	-0.2	?	Shaw (1985)
2	16	Clonmass	55.168	-7.883	SRR-2397	4540±100		4876-5467	5171.5	296	2.63	Wood	Pollen	MHWST	Fanad Head	1.8	1.3	0.3	-0.4	-1.2	-2.2		1.7	0.9	?	Shaw (1985)
2	17	Drumskellan	55.094	-7.227	UB-206	6955±100		7615-7960	7787.5	173	1.3	Tree trunk		MHWST	Culmore Point	1.3	0.8	0.2	0.0	-0.6	-1.2		0.8	0.5	?	Colhoun et al., (1973)
2	17	Corvish	55.282	-7.255	AA32315	14445±100		16275-17150	16712.5	438	2.59	Foraminifera		HAT	Portmore	2	1.5	0.5	-0.4	-0.7	-1.8		2.4	0.2	?	McCabe and Clark (2003)
2	17	Corvish	55.282	-7.255	AA33831	15425±95		18027-18634	18330.5	304	-0.91	Foraminifera		HAT	Portmore	2	1.5	0.5	-0.4	-0.7	-1.8		2.4	-3.3	?	McCabe and Clark (2003)
2	17	Corvish	55.282	-7.255	AA45967	15590±150		18057-18806	18431.5	375	0.59	Foraminifera		HAT	Portmore	2	1.5	0.5	-0.4	-0.7	-1.8		2.4	-1.8	?	McCabe and Clark (2003)
2	17	Corvish	55.282	-7.255	AA45968	16120±160		18718-19273	18995.5	278	1.59	Foraminifera		HAT	Portmore	2	1.5	0.5	-0.4	-0.7	-1.8		2.4	-0.8	?	McCabe and Clark (2003)
2	17	Corvish	55.282	-7.255	AA45966	16460±430		18566-20139	19352.5	787	-0.41	Foraminifera		HAT	Portmore	2	1.5	0.5	-0.4	-0.7	-1.8		2.4	-2.8	?	McCabe and Clark (2003)
2	17	Corvish	55.282	-7.255	AA33832	17540±110		19975-20524	20249.5	275	-2.41	Foraminifera		HAT	Portmore	2	1.5	0.5	-0.4	-0.7	-1.8		2.4	-4.8	?	McCabe and Clark (2003)

Table 6.10 Existing sea level data from the Fanad peninsula (database region 16-17)

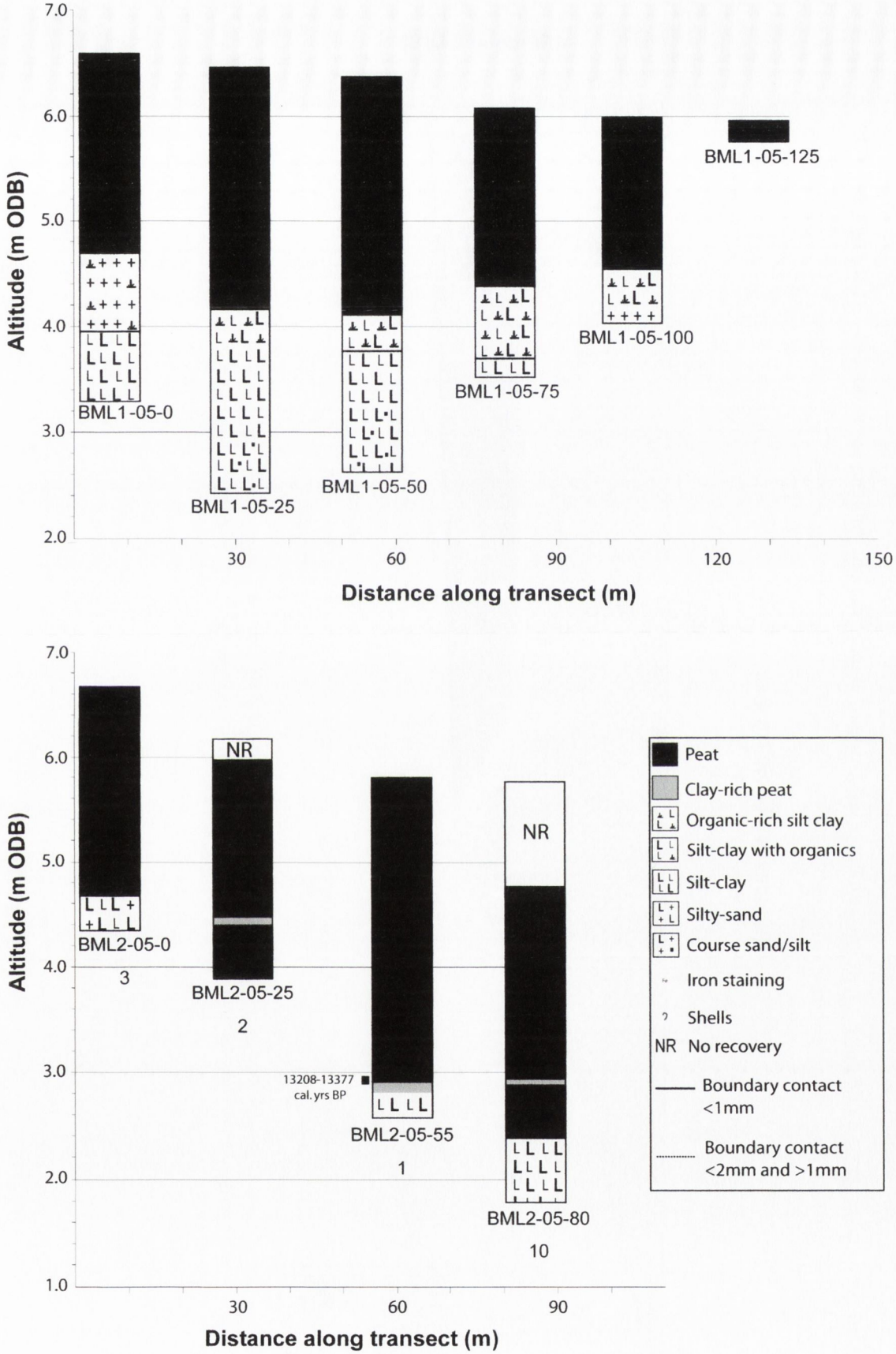


Figure 6.29 Summary lithostratigraphic diagram for transects BML1 and BML2 at Ballymichael. (See figure 6.28B for transect locations)

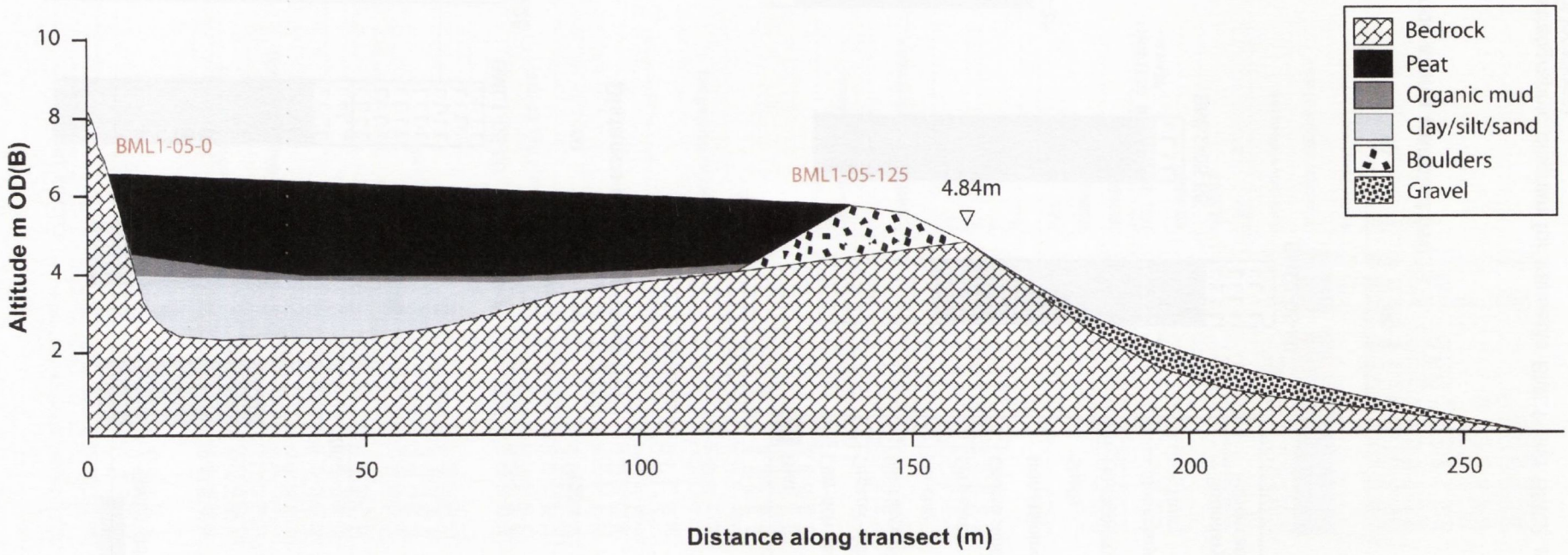


Figure 6.30 Cross – section of the marginal marine basin at Ballymichael along transect BML1. (See figure 6.28B for core locations)

many of the salt-marshes in this region suggest much of the land in the coastal zone was more intensively used in the past (Curtis and Sheehy-Skeffington, 1998). In Donegal, evidence for human activity dating back to the Mesolithic is extensive (Lacy, 1983) and this human occupation resulted in the clearance of much of the natural vegetation, particularly from the Neolithic onwards (Mallory and McNeill, 1991).

The Fanad peninsula lies on the Holocene 0m isobase proposed by Shaw (1985); a line that divides areas to the east which have experienced Holocene sea levels above present msl and areas to the west where the entire Holocene sea-level story has been acted out below present msl. Ballymichael is contained within region 17 (Lough Swilly) of the sea-level database, and data from this area along with data from the adjacent region 16 are contained within Table 6.10.

6.3.2 Ballymichael Stratigraphic Characteristics

6.3.2(i) Lithostratigraphy

Ten cores orientated across two transects were collected from the study site at Ballymichael. Detailed lithostratigraphic information of these cores is contained in Figure 6.29 and summarized in cross section in Figure 6.30. Together, these cores evidence a topographic hollow in-filled by both organic and minerogenic sediments to an altitude of between c. +5.9 to 6.6 m OD(B). With the exception of core BML2-05-25, all of the cores have a clastic unit at their base and in a number of the cores, small (<5 mm) clasts are encountered within this clastic unit. The altitude at which the clastic unit rests upon bedrock differs across the basin: towards the margins of the basin, it is found at c.+4.2 m OD(B) (BML2-05-0) although is closer to +1.8 m OD(B) nearer to the centre of the basin. In all of the cores, a substantial depth of saturated, poorly humified peat containing abundant wood fragments and plant macrofossils is encountered. This peat unit is commonly c. 2 m in thickness although it tapers towards the basin sill where it becomes less than 20 cm in depth (BML1-05-125). The contact between the peat and underlying clastic unit occurs at varying altitudes across the basin and ranges between +4.65 m OD(B) and +2.4 m OD(B). In cores BML2-05-0 and BML2-05-80, a silt-clay unit is encountered immediately below the peat. However in core BML2-05-55 and in all of the cores in transect BML1, the gradation from peat into the underlying silt-clay unit is less abrupt and a transitional organic silt-clay unit of up to 75 cm thickness is present.

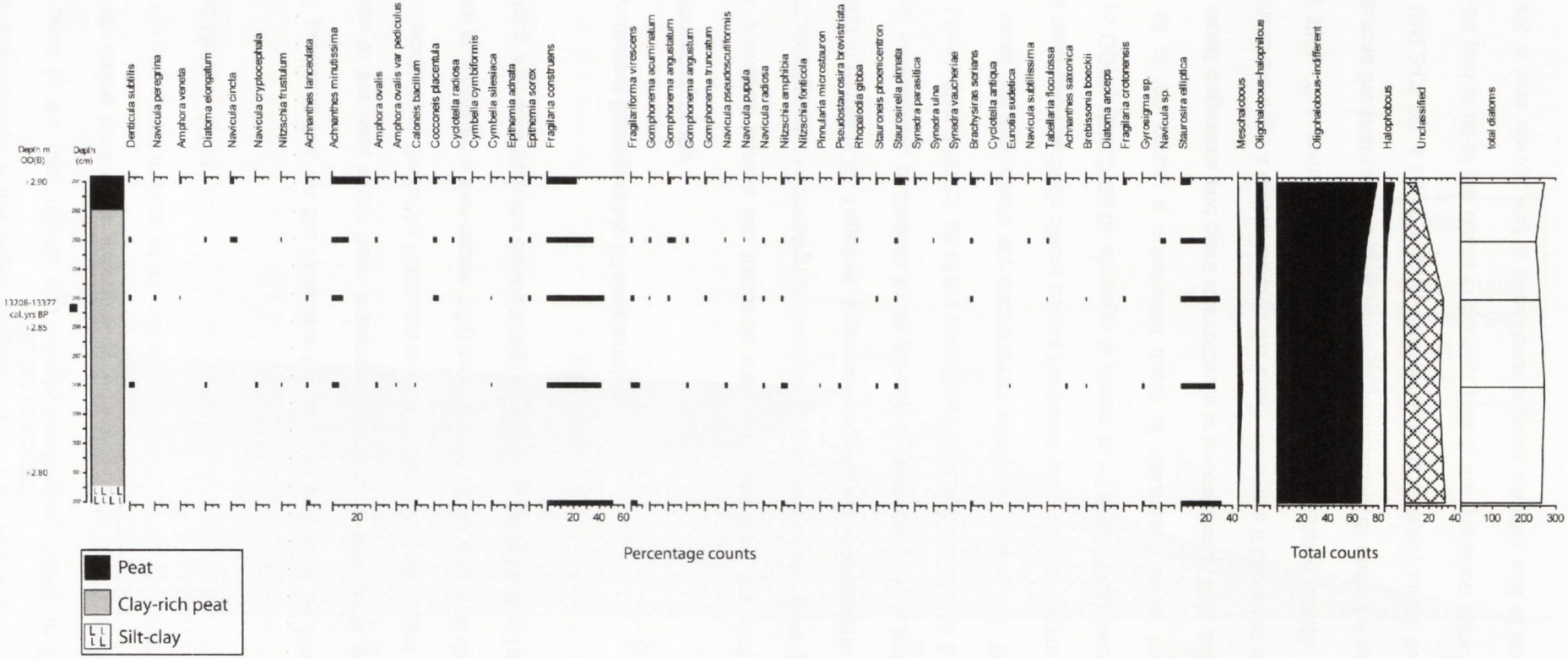


Figure 6.31 Ballymichael fossil diatoms from core BML2-05-55 with associated lithostratigraphy and dated horizons. (Analyst: Caroline Hillier)

A series of large clasts and boulders are present across the entrance to the basin. However, the construction of a drainage channel across the basin sill necessitated their removal and this greatly facilitated determination of the (bedrock) sill height.

6.3.2(ii) Biostratigraphy

Foraminiferal and diatom analyses were undertaken across the regressive contact found in BML2-05-55. No foraminifera were found in the core although high concentrations of diatoms were present in samples across the contact. This permitted counts of c.250 frustules to be obtained from each sample and the results are presented in Figure 6.31. It is evident that all 5 of the samples taken from across the lithostratigraphic contact have a high concentration of the oligohalobous-indifferent diatom species *Fragilaria construens*. This species accounts for 20-50% of the sample population whilst oligohalobous-indifferent forms commonly make up over 60% of the total population in all samples. The sample from the peat unit (291 cm) also contains low counts of both oligohalobous-halophilous and halophobous species, the latter being intolerant of saline conditions. In the clay-rich peat unit found between 292-301 cm, both oligohalobous-halophilous and halophobous species largely disappear and are replaced in the sample at 298cm by low counts of mesohalobous diatoms which account for c.5% of the population. Neither mesohalobous, oligohalobous-halophilous or halophobous diatoms are encountered in the silt-clay unit underneath the clay-rich peat and only oligohalobous-indifferent species may be identified.

6.3.2(iii) Chronostratigraphy

A sediment sample from the regressive contact in BML2-05-55 was sent for radiocarbon dating analysis and returned an age of 13208-13377 cal. years BP (Beta 216639) (CALIB 5.0.1; 95% confidence limits) (Appendix V).

6.4 SUMMARY

- Eight study sites from the Shannon estuary that were believed to hold information on past sea-level change were identified and are presented in this chapter. Collectively, the 30 radiocarbon dates taken from these sites provide a firm chronological framework for the coastal sedimentary sequences dating back to the start of the Holocene period at 11 000 BP.

- The studies sites from the outer Shannon (Rahone, Blackweir Bridge, Moyasta) and mid Shannon (Islandmagrath) were investigated using hand coring equipment and all reveal intercalations of both organic and minerogenic sediments. The cores taken from the sites in the inner Shannon (Meelick Creek, Coonagh, Ballinaccurra, Clonmacken) were commissioned and extracted by the National Roads Authority (although the litho-, bio- and chronostratigraphical analyses have been undertaken by the author).

- All the cores taken from sites located in the inner Shannon reveal abundant microfossil evidence with high counts of both agglutinated and calcareous foraminifera evident in the sediment column. However, the same is not true for those cores extracted from sites in the mid and outer Shannon: Diatom counts are either extremely low (or absent) in all fossil sediment samples whilst none of the core samples taken from sites located in the outer Shannon have any foraminiferal tests present.

- In addition to the Shannon estuary sites documented here, a number of other sites were also visited. However, these did not contain sequences that were suitable for the study of (Holocene) RSL change and thus have not been discussed in detail. Instead, lithostratigraphic evidence from these sites has been included in Appendix IV.

- Only one unaltered, contemporary salt-marsh from the inner Shannon estuary was discovered: The salt-marsh at Ringmoylan Quay appears relatively unaffected by anthropogenic activities such as embankment and this meant continuous foraminifera sampling could be achieved across the high marsh-upland transition. Cluster analysis showed the marsh surface foraminifera at Ringmoylan Quay to be zoned with respect to elevation and a high marsh and a mid-low marsh assemblage were identified.

- Elsewhere along the west coast of Ireland, very few sites containing intercalations of organic and clastic material were discovered (Appendix IV). However, at Rossadilisk (Connemara) one such site was identified and radiocarbon dating evidence reveals a Lateglacial age for the sedimentary sequence.

- Similarly, very few sedimentary sequences that appeared diagnostic of isolation basins were discovered in northwest Ireland (Appendix IV). However, the lithostratigraphy of the marginal marine basin located at Ballymichael (Donegal) identifies it as a potential isolation basin. As with the site at Rossadilisk, radiocarbon dating evidence reveals a Lateglacial age for the sequence.

CHAPTER SEVEN

Field Evidence for Postglacial RSL Change in Ireland

7.1 The Evidence for Sea-Level Change in the Inner Shannon and Fergus Estuary

Together, the litho-, bio- and chronostratigraphic information from four of the sites located in the inner Shannon (Ballinacurra Creek west, Coonagh west, Meelick Creek, Coonagh south) enables fifteen SLIPs to be established (Table 7.1; Figure 7.1). Two techniques have been used to establish separate estimates for the indicative meanings of the inner Shannon SLIPs: firstly, an assemblage zone approach based upon the contemporary distribution of foraminifera at Ringmoylan Quay (section 3.8.1(i)); secondly, the Horton and Edwards (2006) SWLI-2006 UK foraminifera-based transfer function for tide-level (section 3.8.1(ii)). The altitude of past MSL (relative to OD(B)) has been established via equation 3.2 whilst equation 3.3 has been used to obtain estimates of altitudinal error (section 3.9). The remaining eight dates from the inner Shannon sites possess no clear indicative meaning and so may only provide limiting information on the former position of palaeo mean sea level (Table 7.2).

Also included in Figure 7.1 are the two radiocarbon dates from Islandmagrath in the Fergus estuary. Although Tooley (1978) warns against the amalgamation of sea level data from separate geographical areas, this approach may be justified by the close geographical proximity of the inner Shannon and Fergus estuaries (Figure 6.4) allied with the late Holocene age of the dated peat horizons at Islandmagrath (Table 6.5). Combined, the litho-, bio- and chronostratigraphic information from Islandmagrath enables two index points to be established (Table 7.1). The indicative meaning of the two index points has been calculated by the assemblage zone approach based upon the contemporary foraminifera data from Ringmoylan Quay. In neither of the two dated samples do foraminifera counts reach 40. Since previous studies have shown that the incorporation of low sample counts in statistical analyses can potentially yield misleading results

						Calibrated age (yrs BP)				Assemblage Zone Approach (AZA)				Transfer Function Approach (TFA)				
Number	SL database Category	Site	Lab. Code	Material	^{14}C age	Max.	Mean	Min.	Alt. (m ODB)	RWL (m)	IR (m)	RSL (m)	Error ± (m)	RWL (m)	IR (m)	RSL (m)	Error ± (m)	Tendency
1	SLI 1	Island Magrath	Beta 211393	Peat	2510±40	2743	**2557**	2370	0.20	1.65	0.3	**-1.45**	0.37	-	-	-	-	5
2	SLI 1	Island Magrath	Beta 211392	Peat	3170±40	3470	**3372**	3273	0.43	1.65	0.3	**-1.22**	0.37	-	-	-	-	1
3	SLI 1	Meelick Creek	Beta 211398	Organic sediments	3480±50	3883	**3759**	3634	0.85	2.15	0.3	**-1.3**	0.37	3.07	1.00	**-2.22**	0.61	8
4	SLI 1	Meelick Creek	Beta 203549	Peat	3660±40	4139	**4007**	3874	0.69	2.15	0.3	**-1.46**	0.37	3.14	1.00	**-2.45**	0.61	8
5	SLI 1	Meelick Creek	Beta 203547	Peat	4030±40	4784	**4601**	4417	-0.5	2.15	0.3	**-2.65**	0.37	3.07	1.00	**-3.57**	0.61	1
6	SLI 1	Ballinacurra Creek west	Beta 211394	Peat	4390±50	5274	**5062**	4849	-4.07	2.15	0.3	**-6.22**	0.37	2.99	1.00	**-7.06**	0.61	1
7	SLI 1	Ballinacurra Creek west	Beta 203540	Peat	4590±40	5460	**5258**	5055	-1.76	2.15	0.3	**-3.91**	0.37	3.04	1.00	**-4.80**	0.61	8
8	SLI 1	Ballinacurra Creek west	Beta 203539	Peat	5530±50	6432	**6323**	6214	-2.83	2.15	0.3	**-4.98**	0.37	3.15	1.01	**-5.98**	0.61	8
9	SLI 1	Coonagh West	Beta 203544	Peat	5660±50	6563	**6437**	6311	-1.21	2.15	0.3	**-3.36**	0.37	3.07	1.00	**-4.28**	0.61	8
10	SLI 1	Coonagh West	Beta 203546	Organic sediments	5670±40	6559	**6440**	6321	-1.40	2.15	0.3	**-3.55**	0.37	3.03	1.00	**-4.43**	0.61	1
11	SLI 1	Coonagh West	Beta 211397	Organic sediments	5790±40	6713	**6600**	6486	-1.03	2.15	0.3	**-3.18**	0.37	3.02	1.00	**-4.05**	0.61	1
12	SLI 1	Meelick Creek	Beta 203548	Peat	5860±40	6779	**6671**	6562	-1.29	2.15	0.3	**-3.44**	0.37	3.07	1.00	**-4.36**	0.61	5
13	SLI 1	Coonagh West	Beta 203545	Organic sediments	6400±50	7425	**7340**	7254	-2.11	2.15	0.3	**-4.26**	0.37	3.03	1.00	**-5.14**	0.61	8
14	SLI 1	Ballinacurra Creek west	Beta 203538	Peat	6500±50	7552	**7430**	7308	-3.63	2.15	0.3	**-5.78**	0.37	3.10	1.00	**-6.73**	0.61	8
15	SLI 1	Coonagh south	Beta 203537	Peat	8070±50	9129	**8949**	8769	-11.72	2.9*	0.5	**-14.62**	0.37	-	-	-	-	5

* Reference water level inferred from lithostratigraphy

Table 7.1 Sea level Index points from the Inner Shannon and Fergus estuary. (Tendency codes based upon Shennan, 1994. See Appendix I).

Number	SL database Category	Site	Material	Lab. Code	^{14}C age	Calibrated age (yrs BP)			Alt (m) OD(B)	Max/Min extent of marine influence	Indicative Meaning (m)	Max/Min height msl (m)	Error (m)
						Max.	Mean	Min.					
16	Limiting date I	Ballinacurra Creek west	Peat	Beta 203536	6290±50	7321	**7173**	7025	-4.31	MHWST	2.9	-7.21	0.34
17	Limiting date I	Coonagh south	Peat	Beta 203541	9170±40	10482	**10360**	10237	-12.40	MHWST	2.9	-15.30	0.34
18	Limiting date I	Coonagh south	Peat	Beta 203542	9350±70	10742	**10520**	10298	-12.56	MHWST	2.9	-15.46	0.34
19	Limiting date I	Coonagh south	Peat	Beta 203543	9580±40	11122	**10932**	10741	-12.96	MHWST	2.9	-15.86	0.34
20	Limiting date II	Meelick Creek	Shell	Beta 211399	7120±50	7687	**7590**	7492	-11.69	MHWST-60	2.3	-13.99	-
21	Limiting date II	Meelick Creek	Forams	Beta 216647	7390±70	7982	**7837**	7691	-11.71	MHWST-60	2.3	-14.01	-
22	Limiting date II	Coonagh south	Shell	Beta 211396	7620±40	8170	**8074**	7978	-12.87	MHWST-60	2.3	-15.17	-
23	Limiting date II	Coonagh south	Organic sediments	Beta 211395	7800±40	8646	**8550**	8453	-12.87	MHWST-60	2.3	-15.17	-

Table 7.2 Limiting dates from the Inner Shannon estuary

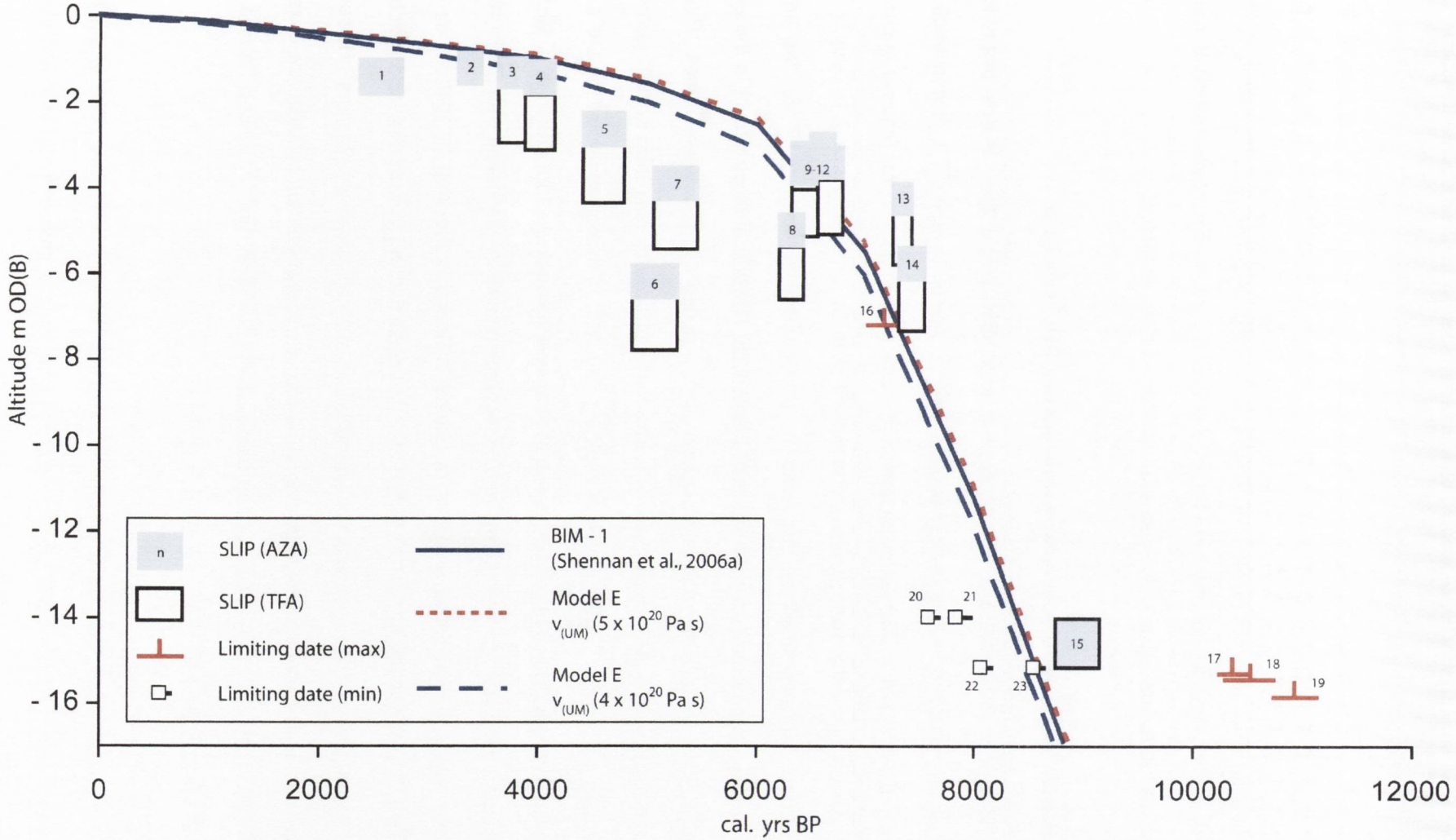


Figure 7.1 Reconstruction(s) of Holocene RSL for the inner Shannon and Fergus estuary. GIA model BIM -1 (Shennan *et al.*, 2006a); Model E v_{UM} (5 x 10^{20} Pa s) and Model E v_{UM} 4 x 10^{20} Pa s are also shown. Numbers relate to those given in Tables 7.1 and 7.2

(e.g. Patterson and Fishbein, 1989), the indicative meanings from the two dated Islandmagrath foraminifera samples were not interpreted through the use of the SWLI-2006 transfer function.

Collectively, the 23 radiocarbon dates taken from the inner Shannon and Fergus estuaries evidence changes in RSL back to 9000 BP and place limits upon the maximum position of RSL back to c.11 000 BP. The reliability of the reconstructed sea level data are discussed in detail below and the implications for the veracity of the (newly developed) glacial rebound model predictions of RSL for the region presented in Chapter 5 are also considered.

7.1.1 Trends in the Data and Comparison with Existing Sea Level Data

Both the index point and limiting data have been combined into a single age – altitude plot of RSL for the inner Shannon and Fergus estuary and this may be viewed in Figure 7.1. Reconstructions of msl derived from both the assemblage zone approach (AZA) and transfer function approach (TFA) are shown and data point numbers correspond to those given in Table 7.1. Despite the scatter in these data (which will be discussed in greater detail in section 7.1.2(iii)), two broad episodes of RSL change may be identified from these data. The first of these occurs in the early Holocene (c. 9000 – 7000 BP) and is associated with a period of rapid sea-level rise. This is evidenced by SLIPs (9) – (15) which together reveal that RSL rose by over 10m in 2000 years; a rate of around 5-6 mm yr^{-1}. This phase of rapidly rising RSL is further supported by the early Holocene limiting data (17) - (19) which show that sea level was below –15m OD(B) at c. 10 500 BP. At this juncture, it is worth noting that the biostratigraphic analysis associated with SLIP (15) is inconclusive: whilst very tentative evidence of marine influence is provided by (extremely) low counts of the diatom species *Nitzschia navicularis*, it is notable that the sample was lacking in both Chenopodiaceae or Armeria pollen which is characteristic of a salt-marsh environment[1]. However, even if this regressive contact is only limiting in nature, it still provides strong evidence of rapid early Holocene RSL rise since the altitude of palaeo mean sea level (pmsl) cannot be above the height of this peat unit.

The second phase of RSL change is evidenced by SLIPs (1) – (8) and takes place in the mid-Holocene between c. 7000 – 3000 BP. This episode of change is characterized by much slower

[1] Pollen analysis undertaken by Professor Pete Coxon

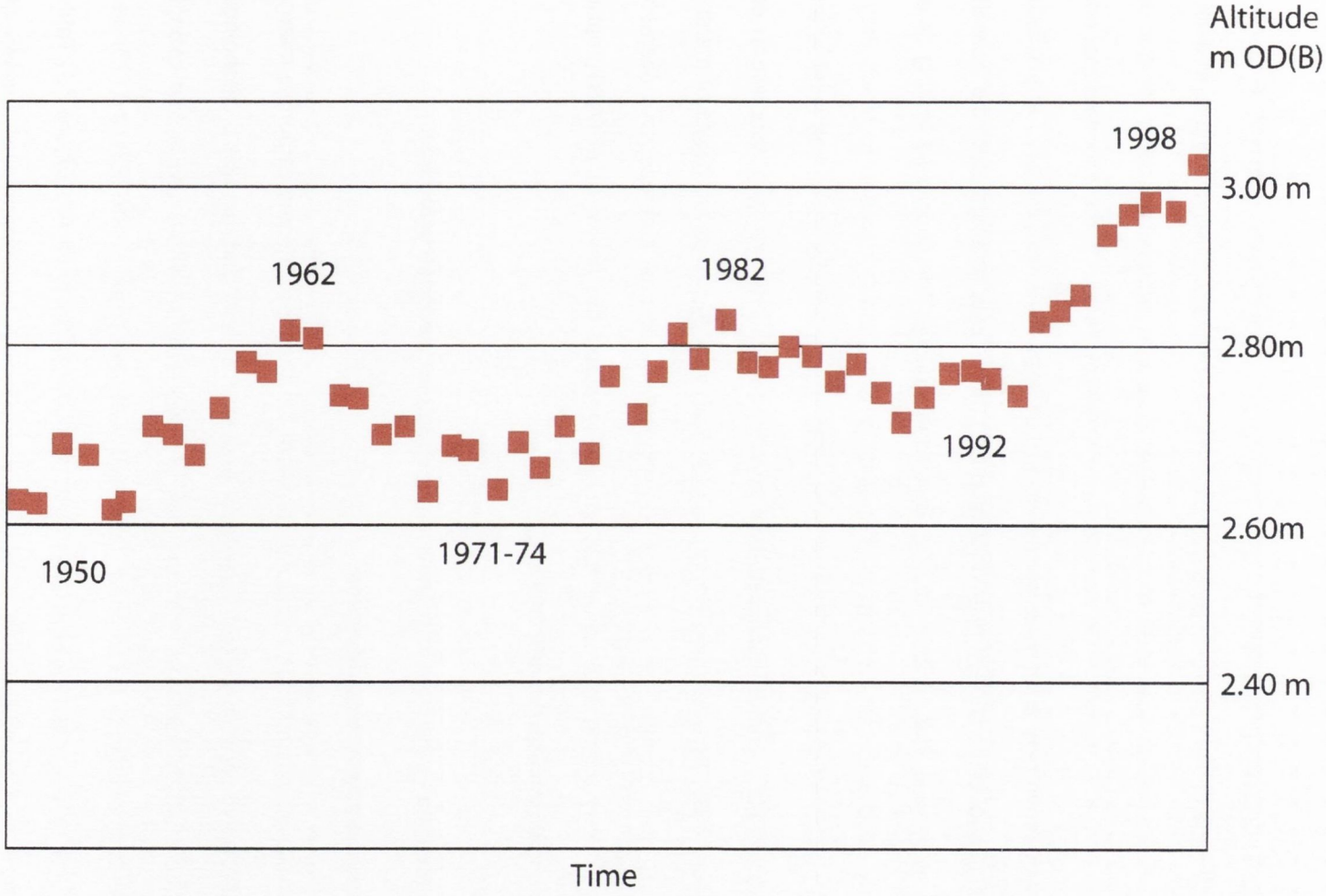


Figure 7.2 The trend of water level movements at Limerick Docks, Shannon estuary based on uncorrected tide gauge records 1949 to 1998. (Adapted from Healy, 2002)

rates of RSL rise with the data suggesting that only around 4m of RSL rise took place within this period: This equates to a rate of c. 1.0 mm yr^{-1}. This overall trend of rapid early Holocene RSL rise followed by a much-reduced mid to late Holocene rate is typical of an area at (or beyond) the margins of the BIIS (e.g. Shennan and Horton, 2002).

Whilst the archaeological investigations undertaken by O'Sullivan (2001) delivered a number of dates from the banks of the Shannon estuary, there still exists a lack of reliable sea level data with which to compare these new data to. The inner Shannon radiocarbon dates presented in O'Sullivan (2001) are contained in the database and shown in Table 6.2: the only data that enable sea level to be fixed in time and space are the dated fish traps from the Deel and Fergus Estuary, yet these are associated with very large vertical error components and post date the new SLIPs presented here by over 1000 years. The remaining dates occupy the secondary limiting date tier and thus only enable a maximum age to be fixed for the sediment horizon in which they are contained. A number of studies have chosen to consider geological records of late Holocene RSL change alongside 20th century mean sea-level trends from tidal records (e.g. Woodworth *et al.*, 1999; Lambeck, 2002; Shennan and Horton, 2002): A 50 year record of the height of mean high tide is available from Limerick dock (Figure 7.2) although substantial anthropogenic alteration to channel morphology during this period has grossly amplified the height of MHWST, nullifying meaningful comparison with the new SLIPs.

7.1.2 Complications in the Data and Their Implications for Reconstructed RSL

7.1.2 (i) Chronostratigraphic anomalies

Overall, the AMS radiocarbon dates sourced from both the peat horizons and carbonate material in the MOSTAP cores reveal very good agreement and, for the most part, provide a chronologically coherent pattern which conforms with the law of superposition. There are a few exceptions however; the most notable of these being the paired shell and organic date (22) and (23) derived from the organic mud unit in core MS03 at an altitude of c. –12.85cm OD(B) (Figure 7.1; Table 6.4 and Table 7.2). This core section is of especial interest since biostratigraphical analysis of the organic clay horizon reveals abundant calcareous foraminifera evidencing marine conditions. If the age of this marine clay unit is consistent with dates from the peat horizons above and below, the data would strongly contradict the newly established glacial rebound model predictions for the

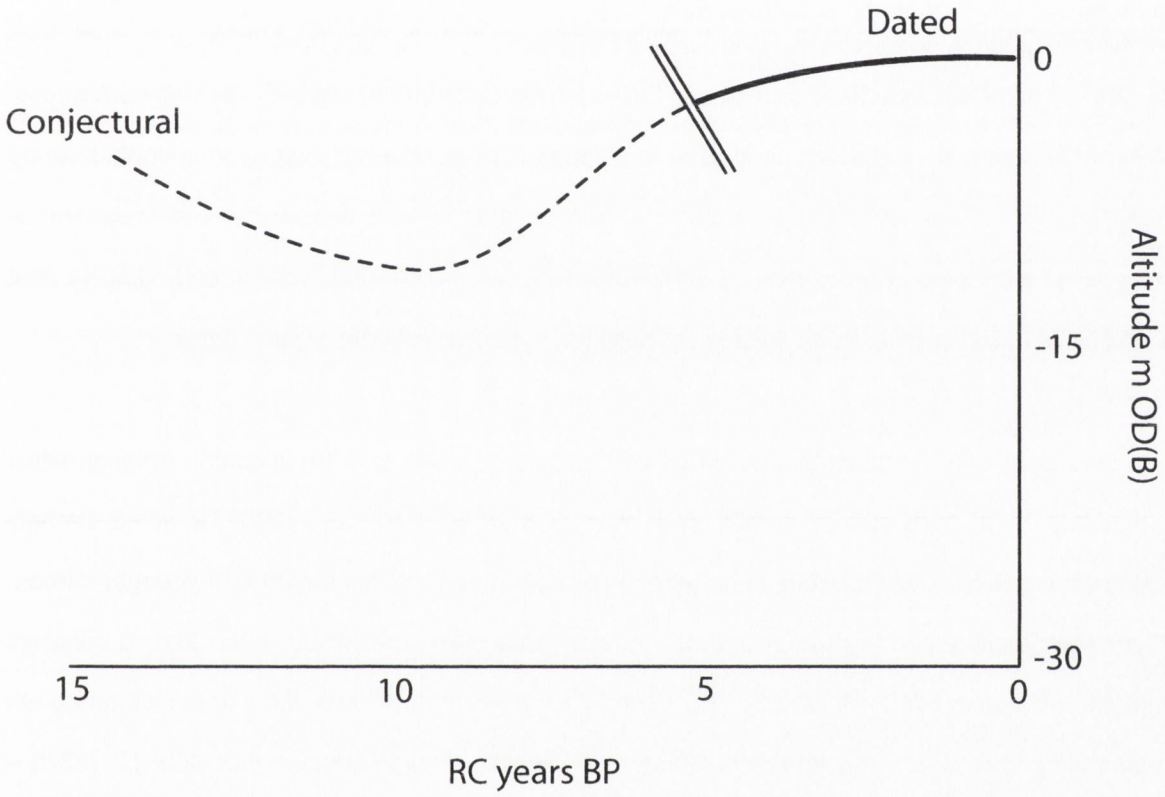


Figure 7.3 Taylor *et al.* (1986) RSL curve for southwest Ireland

region and provide support for the Taylor *et al.* (1986) RSL curve for southwest Ireland (Figure 7.3). However, whilst the (freshwater) peat horizons which bracket the organic clay horizon date to c. 10 200 – 11 000 BP, the paired shell and bulk organic dates return an age of c. 8000 – 8500 BP. The three peat dates sourced from above and below the organic clay horizon are in sequence (Figure 6.11) and their validity is not in question. On the other hand, radiocarbon dates from both carbonate material and predominantly minerogenic sediments are notoriously unreliable (see section 3.7.1) and it may be the case that the paired shell and bulk organic dates are erroneous. More likely however, is the possibility that this section of the core has been contaminated during sampling by cavings of material introduced from further up the sequence. This assertion is supported by both the abrupt nature of the contact between the clay-rich horizon and adjacent peat units (Table 6.3c) and the broad agreement between the shell and bulk organic dates.

The two remaining irregularities involve SLIP (6) (Beta 211394) and the inversion present within core IMG3-05-25 from Islandmagrath. With regards SLIP (6) from core MS02, it would strongly appear that this date is erroneous and has most probably been contaminated with younger carbon. This assertion is supported by the remaining four dates from core MS02 which form a coherent sequence between 5000 – 7300 BP. The inconsistency between the two dates from Islandmagrath is less easy to resolve. Consideration of Figure 7.1 would perhaps suggest that SLIP (1) (2370 – 2743 cal. yrs BP; Beta 211393) under estimates the age of the peat horizon although insufficient evidence exists to verify this claim. (Interestingly, at Islandmagrath O'Sullivan (2001) recorded a late Bronze age post-and-wattle structure along the foreshore which returned an almost identical age to SLIP (1) (2540±20 RC yrs BP GrN-20974). Unfortunately, the lack of detailed stratigraphic and levelling information meant this find was of little help in resolving the dating inconsistencies in core IMG3 –05 –25).

7.1.2(ii) Interpretation of the indicative meaning

Imbrie and Webb (1981) and Birks *et al.* (1990) outlined a number of requirements that are inherent to quantitative palaeoenviromental reconstruction (section 3.8.1). Of these stipulations, two are of particular significance when considering the validity of the reference water level reconstructions made by both the AZA and TFA: firstly, the contemporary distribution(s) of foraminifera used to quantify the indicative meaning are representative of the local environment,

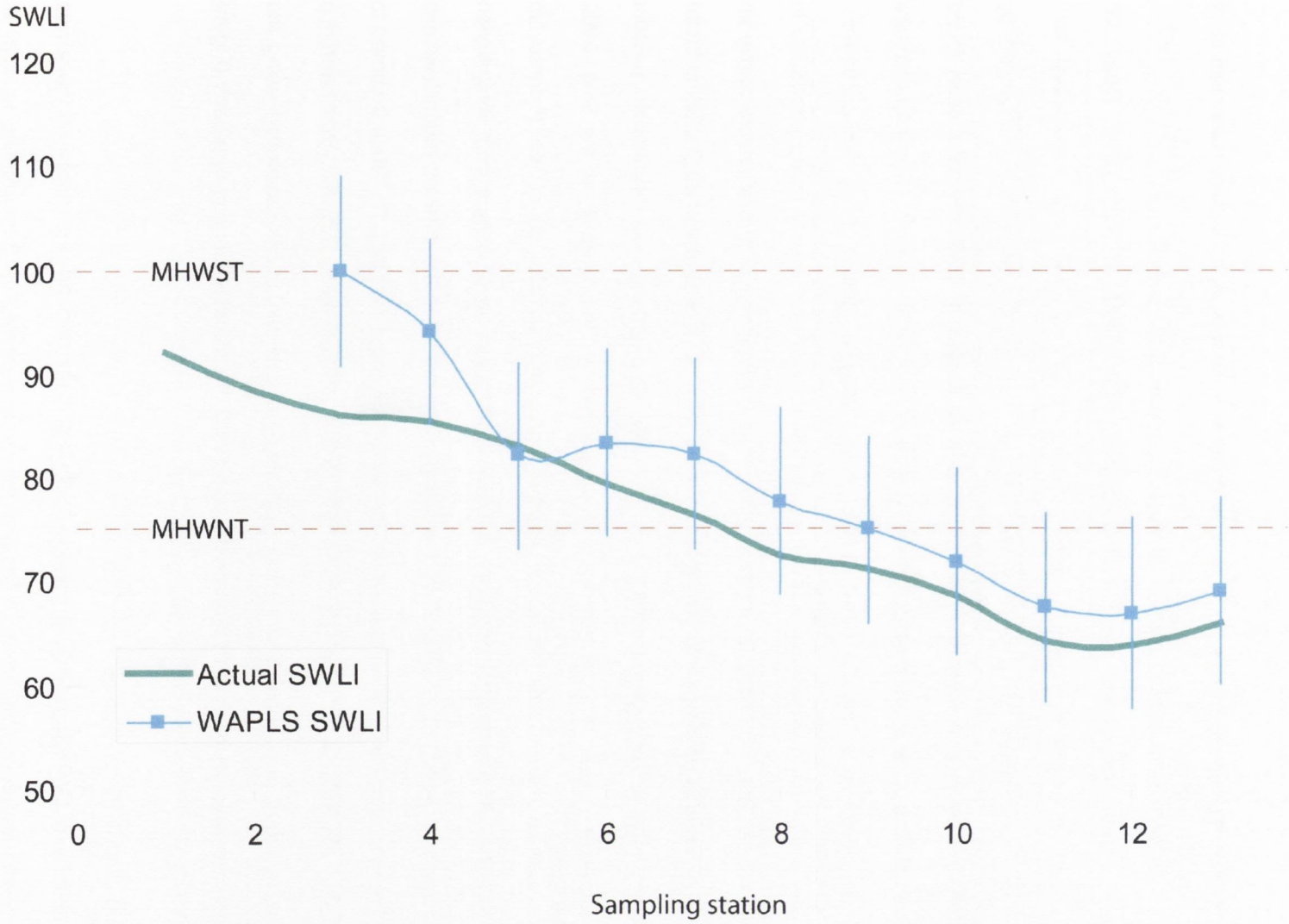


Figure 7.4 The relationship between observed SWLI and foraminifera transfer function predicted SWLI of marsh surface samples at Ringmoylan Quay

both for the present and the past; secondly, the composition of the fossilized foraminiferal assemblages found within the core samples remains unchanged since the time of deposition. The likelihood that these two criteria have been met is considered in more detail below.

a) Suitability of the modern analogues

Whilst a number of previous studies from the west coast of Ireland reveal a lack of halophytic plant zonation in salt-marshes (Sheehy Skeffington and Wymer, 1991, Duffy and Devoy, 1999), reference to Figure 6.6 shows that the contemporary foraminifera at Ringmoylan Quay are vertically zoned (albeit not strongly). A bipartite division between a low marsh assemblage with relatively high abundances of calcareous taxa and a high marsh assemblage dominated by agglutinated species is in evidence, even though only weak salinity gradients have been found across the inter-tidal zone in the inner Shannon (Wilson *et al.*, 1993). However, whilst foraminifera may be vertically zoned across the marsh at Ringmoylan Quay, a number of key differences exist between the foraminiferal distribution found here and those more commonly found elsewhere in northwest Europe. The most notable of these differences is the absence of foraminifera above an elevation of c. MHWST–60cm; a situation that contrasts with many other marshes (including those incorporated in the SWLI-2006 transfer function), which have agglutinated foraminifera present between MHWST – HAT. This discrepancy is demonstrated by the propensity of the SWLI-2006 transfer function to overestimate the actual height of high marsh samples taken from Ringmoylan Quay (Figure 7.4). The altitudinal difference between high marsh assemblages from Ringmoylan Quay and those included in the SWLI-2006 transfer function results in the differing reconstructions of the reference water level for samples from the MOSTAP cores (Table 7.1). When it comes to reconstructing the altitude of pmsl (Equation 3.2), these differences in reconstructed reference water level manifest in a consistent vertical displacement between SLIPs developed using the assemblage zone approach and SLIPs constructed using the transfer function approach (Figure 7.1).

The magnitude of this vertical offset cannot easily be dismissed as an artifact of the errors associated with surveying the marsh at Ringmoylan Quay. Neither can the sampling strategy at Ringmoylan Quay be questioned since samples were taken across the high marsh-upland transition and those taken from above this elevation (within a zone of euryhaline vegetation)

revealed an absence of foraminifera. Instead, an explanation may be sought by examining the nature of the prevailing environmental conditions operating within the inner Shannon. The tidal characteristics of the estuary are of particular interest here since other studies from macro-tidal settings have reported a similar 'barren' zone of foraminifera across the upper marsh surface (Charman *et al.*, 1998; Haslett *et al.*, 1998b). In their study from the severely macro-tidal Severn estuary, Haslett *et al.* (1998b) find that this barren zone is broadly equivalent to MHWST-HAT and may be explained by a reduction in the duration of high marsh inundation in comparison to marshes situated in meso/micro tidal estuaries. The barren zone at Ringmoylan Quay extends even further down the marsh since no foraminifera are found above MHWST–60cm. However, this may be explained in terms of the salinity of the water body at this point in the estuary. In their study on nutrient distribution within the inner Shannon, Wilson *et al.* (1993) find that estuary salinity at Ringmoylan Quay is only 15‰ compared with nearly 30‰ in the lower reaches of the Shannon. This will result in reduced salinity across the marsh surface and the elevation of the salinity threshold (which determines the presence/absence of foraminifera) will be lowered.

Another irregular feature of the marsh at Ringmoylan Quay is the prevalence of *Jadammina macrescens* across the marsh surface and onto the fringing tidal flats (Figure 6.6). Although concentrations rarely exceed 20% in the middle and lower marsh, this figure is likely to rise significantly if the (probable) allochthonous component of the assemblage is removed (see below). This situation contrasts with that more commonly experienced in other marshes from the east coast of Ireland (Brooks – unpublished data; Appendix VII) and elsewhere in the UK, which tend not to have such high counts of *Jadammina macrescens* in the mid *and* low marsh (Table 3.3). (Despite this, other studies from the British Isles have documented high counts of *Jadammina macrescens* across the marsh and these studies are also from macrotidal coasts (e.g. Charman, *et al.*, 1998; Haslett *et al.*, 2001; Edwards, 2006a). These conditions are suggested to cause prolonged periods of sub-aerial exposure of the marsh surface inducing harsh conditions which favour the proliferation of *Jadammina macrescens* (Edwards, 2006a)). In the MOSTAP cores, *Jadammina macrescens* also dominate most of the fossil assemblages and, as with the situation at Ringmoylan Quay, are found alongside relatively high counts of species indicative of mid to low marsh environments (Figure 6.12a-d). However, this situation is atypical of the contemporary marshes included in the SWLI-2006 transfer function and none of the marshes included in the regional training set have samples with high abundances of *Jadammina macrescens* alongside *Miliammina*

fusca. As a consequence, the modern analogue technique (which is used to quantify the similarity (or dissimilarity) between fossil assemblages and the modern (regional) training set) shows that none of the fossil assemblages from the MOSTAP cores have modern analogues.

Taking into consideration the above, it would appear that the marsh at Ringmoylan Quay provides a better analogue for conditions in the inner Shannon than the Horton and Edwards (2006) transfer function. Although SWLI-2006 transfer function predictions for the elevation of Ringmoylan Quay surface samples are in surprisingly good agreement with actual sample elevation, significant departures emerge in the (important) high marsh section (Figure 7.4). This is because none of the marshes incorporated in the SWLI-2006 transfer function exhibit this barren zone of foraminifera above MHWST which means the transfer function approach tends to over predict the height of the reference water level. However, whilst reconstructions of sea level based upon the assemblage zone approach appear to be more accurate than those delivered by the transfer function approach, two principle uncertainties remain with the AZA. These uncertainties require further consideration before the SLIPs reconstructed via this approach can be interpreted with confidence:

(1) Clearly, the presence of the barren zone of foraminifera is an important characteristic of inner Shannon marshes and has a direct and significant bearing upon reconstructions of sea level. The marsh at Ringmoylan Quay is situated approximately 10km up estuary from the sites in the inner Shannon (Figure 6.5) although it is unclear whether the foraminiferal distribution in this marsh offers a representative analogue for the inner Shannon sites. Reference to Wilson *et al.* (1993) shows that the salinity gradient of the inner Shannon decreases exponentially as one moves up estuary, with salinity falling from 15‰ to 5‰ between Ringmoylan and Meelick Creek. This marked decrease in salinity is very likely to lower the elevation of the barren zone of foraminifera. If this barren zone is lower in the inner Shannon, the SLIPs displayed in Figure 7.1 are likely to underestimate the position of palaeo msl (Equation 3.2). This is hard to investigate further since no salt-marshes currently exist up estuary from Ringmoylan (Curtis and Sheehy-Skeffington, 1998). Furthermore, at sites where nascent salt-marsh vegetation does exist alongside extensive reed bed communities, the upper portion of the inter-tidal zone has been heavily modified and embanked.

The picture becomes increasingly complicated when one considers that the inner Shannon estuary was considerably wider in the past than it is today (Figures 6.2, 6.3). Indeed, the reclamation and embankment of much of the inner Shannon will have significantly altered the hydrodynamics of the estuary and this includes its salinity regime (Healy and Hickey, 2002). Thus even if unmodified salt-marshes were presently flourishing adjacent to where the MOSTAP cores have been extracted, they would most probably be unrepresentative of the Holocene palaeoenvironment. Instead, a salt-marsh presently located on a wider stretch of the estuary may potentially offer a better analogue for conditions in the past.

(2) Reference to Figure 6.6 reveals that the foraminiferal samples from Ringmoylan Quay possess very high percentages of the calcareous taxa *Brizalina sp.* and *Fursenkoina sp.* as well as the agglutinated species *Trochammina ochracea*. Both *Brizalina sp.* and *Fursenkoina sp.* are shelf species, which are characteristic of sub-tidal environments and are likely to have been advected onto the marsh surface by wave action (Murray, 1979; Horton and Edwards, 2006). No species were found alive in the surface samples and calcitic species such as these are readily dissolved in the acidic high marsh environment (Scott and Medioli, 1980a, b; Green *et al.*, 1993; Edwards, 2006a). Similarly, *Trochammina ochracea* is an epiphytic shelf species that is commonly found attached to seaweed and pebbles and is generally not considered indicative of high marsh environments (e.g. Murray, 1979). Its presence on the high marsh surface is likely to reflect transport and deposition on to the marsh around the strandline where seaweed collects after the retreat of the tide (Horton and Edwards, 2006). All of these three species are absent from the organic units recovered in the MOSTAP cores yet between them, they frequently account for over 50% of the sample population at Ringmoylan Quay. These high counts of (probable) allochthonous species at Ringmoylan Quay potentially mask more subtle changes that may exist across the contemporary marsh surface. Indeed, if counts of *Brizalina sp.*, *Fursenkoina sp.* and *Trochammina ochracea* are screened out (as they are in the SWLI-2006 transfer function), a mid marsh zone (+1.4 to +1.9m OD(B)) characterized by fairly high percentage counts of *Miliammina fusca* may tentatively be recognized. For the most part, this makes no difference to reconstructions of the reference water level made by adopting the AZA. However, it may make a difference to SLIPs (4) + (8) as they have relatively high counts of *Miliammina fusca* (Figure 6.12a, 6.12d and 7.1). Revised reference water levels would be lower and the SLIPs would subsequently plot higher on the age –

altitude plot (see Equation 3.2). In the case of SLIP (8), this would serve to reduce the discrepancy with the glacial rebound model predictions of RSL.

b) Post-depositional alteration of foraminiferal assemblages

The second principal assumption underpinning the validity of the reconstructed water levels is that the composition of the fossilized foraminiferal assemblages found within the core samples has remained unchanged since the time of deposition. However, often this does not hold true and this is particularly the case in temperate salt-marshes where calcareous taxa are prone to post depositional alteration (Green *et al.*, 1993). Dissolution results in the overrepresentation of agglutinated species and this may contribute to the underestimation of the altitude of pmsl. In the MOSTAP cores, no calcareous tests are discovered within the intercalated organic-clastic units found at the top of cores MS02, MS04 and MS05. This situation is wholly expected for the organic rich horizons since loss on ignition analysis from these units often revealed over 80% organic content which is strongly indicative of a high marsh (or supra-tidal) setting. However, in the minerogenic units that are immediately above and below a number of the peat horizons, clay often comprises over 75% of the sediment matrix and one would expect to find calcareous tests present. Instead, they are absent and the only calcareous foraminifera that are encountered are located at a significant depth below the brackish peat horizons, within the lengthy clastic units found in cores MS03 and MS05 (Figure 6.11). This distribution pattern is consistent with the suggestion that de-calcification of foraminiferal assemblages has taken place within certain sections of the sediment column. This may have been exacerbated by the percolation of organic acids from the overlying peat horizons and could explain why MOSTAP core biostratigraphy does not always mirror core lithostratigraphy (Figures 6.12a-d).

The significance of calcareous test dissolution to salt-marsh palaeoenvironmental studies will depend upon whether calcareous foraminifera are diagnostic species for any of the vegetated salt-marsh environments. However, it is the case that in the Ringmoylan Quay transect, calcareous species represent a negligible contribution to the high marsh assemblage (zone I) which is relied upon to reconstruct the indicative meanings for SLIPs (1) - (14). In some previous studies, calcareous test linings have been observed within the fossil samples and these may be incorporated into a modern training set to minimise the loss of palaeoenvironmental information

(e.g. Edwards and Horton, 2000). However, no calcareous test linings were encountered within the MOSTAP cores therefore this does not provide a viable avenue for further investigation.

Finally, De Rijk (1995b) notes that agglutinates may be at risk from compaction and drying, often leading to post depositional alteration such as the collapse of chamber walls. This is evident in a number of the samples from the MOSTAP cores where high numbers of test linings are present (Figures 6.12a-d). These test linings are likely to belong to either *Jadammina macrescens* or *Haplophragmoides sp.* but the morphology of the test has been altered to such an extent that it is no longer possible to differentiate between the two. This does not present a significant caveat to the reliability of reference water level reconstructions derived from the assemblage zone approach: Only two broad foraminifera zones (I and II) are recognized in the contemporary marsh at Ringmoylan Quay and all fossil samples dominated by either *Jadammina macrescens* or *Haplophragmoides sp.* are assigned an indicative meaning based upon the high marsh zone (zone I) (Figure 6.6 and Table 7.1). However, the high abundance of test linings makes more of a difference to reference water level reconstructions made by the SWLI-2006 transfer function since reconstructed SWLIs take into account individual species SWLI optimum *and* their relevant abundance.

7.1.2(iii) Compaction of the sediment column

In addition to the potential errors already highlighted in section 7.1.2(i) and (ii), it is important to acknowledge that fine-grained minerogenic and organogenic sediments commonly undergo volume reduction after deposition (e.g. Jelgersma, 1961; Terzaghi and Peck, 1967; Greensmith and Tucker, 1971, 1973; Tooley, 1978; Heyworth and Kidson, 1982; Shennan, 1986; Allen 1996; Haslett *et al.*, 1998a; Shennan *et al.*, 2000b; Edwards, 2006a). This is particularly relevant in the inner Shannon since land reclamation is likely to have exacerbated this process (e.g. Greensmith and Tucker, 1987). The potential influence of compaction on Holocene RSL records is long established and widely acknowledged although has commonly been set aside due to the lack of a formal means of correcting for its influence and the difficulty in calculating changing sediment pore water pressure through time (Allen, 2000b; Shennan and Horton, 2002; Brain, 2006). Sediment compaction is complex, involving a variety of processes that alter the arrangement, composition and physical characteristics of an accumulating sedimentary body (Edwards, 2006a). However,

irrespective of the processes in operation, the overall outcome is the reduction in altitude of the sediment volume and the consequent lowering of SLIPs sourced from the affected sediment units.

Reference to Figure 7.1 reveals that there is a fairly significant degree of (vertical) scatter between SLIPs developed from the inner Shannon. In particular, SLIPs (5) – (8) plot especially low on the age – altitude graph and reveal over 3m of vertical scatter between them. Some of the scatter within the data may be explained in terms of the problems associated with reconstructing the indicative meaning and associated indicative range (section 7.1.2(ii)). However, the magnitude of variation within the data suggests other factors are in operation here and consideration of the position of the SLIPs within the sediment column would suggest compaction is a likely candidate (Figure 6.11). This deduction is lent credence by the biostratigraphic record since a number of the MOSTAP foraminifera samples have very high counts of agglutinated test linings and their presence is often indicative of compression of the sediment matrix (e.g. De Rijk, 1995b).

Although quantitative estimates of compaction have proved notoriously hard to obtain, previous investigations by Shennan *et al.* (2000b) and Edwards (2006a) have achieved first order estimates of decompaction. These have been arrived at by comparing the altitude and stratigraphic context of SLIPs with predictions of RSL delivered by glacial rebound modelling. In both of these studies, all of the observed sea level data fall below the altitude of GRM predictions of RSL and a linear trend is evident between the residuals from these GRM predictions and both sediment overburden and underlying sediment. However, here this approach is of less value since the relationship between the reconstructed altitude of the SLIPs, the stratigraphic position of the sample and misfits with GRM predictions is less straightforward: Many of the SLIPs lie above the predicted altitude of pmsl therefore significant, additional errors associated with either the empirical dataset (e.g. underestimation of the indicative range) or the glacial rebound model (see section 7.1.3) must also be present.

In section 7.1.3, the validity of GRM predictions for the Shannon are discussed and particular focus is placed upon possible inaccuracies in the global ice model between c.6000 and 8000 BP. However, it is worth pointing out here that even if residuals between GRM predictions and the new SLIPs are only considered in the time interval between 6000 BP and present, there is still no linear trend between the misfit with the GRM RSL predictions and the amount of sediment overburden or

the thickness of underlying sediment. This is probably because the above approach adopted by Shennan *et al.* (2000b) and Edwards (2006a) does not account for variation in both inter- and intra-core lithostratigraphy. Indeed, organic-rich units are far more prone to both consolidation (and particularly) compression than clastic horizons (e.g. Boswell, 1961; Greensmith and Tucker, 1987) and therefore an alternative approach is to consider the position of SLIPs in relation to the organic layers from which they are obtained. Indeed, if this approach is adopted, it becomes apparent that the biggest departures from the GRM predictions are associated with SLIPs sourced from the top of the thickest organic sediment horizons (Table 7.1; Figures 6.11 and 7.1): The greatest outliers in Figure 7.1 are SLIPs (6) – (8) and these are all sourced from the upper section of the lengthy peat horizon encountered in core MS02 (Figure 6.11; Table 6.3b). This observed relationship questions the veracity of SLIPs (6) – (8).

7.1.2(iv) Changes in Holocene tidal range

The indicative meaning method inherently assumes that significant changes in tidal range have not occurred over Holocene timescales. In reality this is almost certainly untrue, as evidenced by a number of tidal modelling studies from around the British Isles (e.g. Austin, 1991; Shennan *et al.*, 2000c, 2003). These studies suggest that the greatest changes in tidal range generally occurred in the early Holocene due to the dramatic change in coastal configuration caused by rapidly rising RSL. After 6000 BP, changes in tidal range have been significantly smaller due to less pronounced alteration of the coastal geometry. Reference to the work of Austin (1991) suggests that along the west coast of Ireland, changes in Holocene tidal range have been minimal although these inferences are applicable to the open coast and may not hold true for locations within narrow estuarine inlets such as the Shannon. Of greatest concern however, are the affects that 18th and 19th century reclamation have had on tides within the inner estuary. Reclamation activities are known to alter the characteristics of the tidal prism and are likely to magnify tidal range. If the contemporary tidal range of the Shannon is greater than it was in the past, the SLIPs shown in Figure 7.1 will underestimate the altitude of pmsl because the reconstructed reference water levels shown in Table 7.1 will be too high. However, the actual extent to which the tidal range has been altered is unclear. It is worth noting though that present day MHWST at Mellon Point is +2.8m OD(B) in comparison to +3.0m OD(B) at Limerick Dock. At Mellon Point, the estuary is still fairly wide (c.2km) whereas at Limerick dock it is very narrow (c. 200m) (Figure 6.5). Thus despite

considerable narrowing of the estuary between these two tidal stations, the increase in tidal range is actually small. This might suggest that the (possible) effects of changes in Holocene tidal range will be subordinate to the errors introduced by compaction and uncertainties over the accuracy of the indicative meaning.

7.1.3 New Sea Level Data From the Inner Shannon and Fergus Estuary and Their Implications for Glacial Rebound Model Predictions of RSL

Reference to Figure 7.1 reveals the new index point data from both the inner Shannon and Fergus estuaries alongside model predictions of Holocene RSL. These three predictions are generated by the original Shennan *et al.* (2006a) model and the best fit ice-Earth model solutions advanced in Chapter 5. However, despite differences in both the ice and Earth model input parameters, surprisingly little variation occurs between the three model predictions during the Holocene. Overall, the fit between the SLIPs generated by the assemblage zone approach and the model predictions is reasonable and *most* of the significant departures from the model simulations (such as SLIPs 5 – 8) may be convincingly explained in terms of deficiencies with the quality of the sea level data rather than errors inherent within the model itself.

However, despite the overall goodness of fit, several disagreements between the observed and predicted RSL histories should be addressed. It is evident that all three of the predictions pass c.1-2m below SLIP (15) and it should be noted that the large thickness of underlying sediment and sediment overburden in core MS03 mean this index point may have been lowered to some extent by compaction (Figure 6.11). Taking this factor into account will serve to amplify the discrepancy with the model predictions. This departure from the model may be considered from one of two perspectives. Firstly, it is possible that SLIP (15) is in fact only limiting in nature, providing a maximum position for msl instead of fixing it in time and space (See section 7.1.1). If this is the case, the model remains in good agreement with the available empirical data. Conversely, the model itself may underestimate the height of pmsl at this time, necessitating a revision of the model input parameters. Either a small increase in the local-regional ice load or a delay in the timing of regional ice retreat would serve to raise RSL predictions throughout the Holocene, delivering a better fit with SLIP (15). This amendment to the model need not contradict the mid-Holocene data: All of the potential problems discussed in section 7.1.2 will contribute to an underestimation of the

actual altitude of pmsl therefore the data will not contradict the model if they lie below the predicted RSL curve.

Whilst making very minor modifications to the local-regional ice model will enhance the overall fit with the data, such changes are unlikely to be able to reconcile model disagreement with SLIPs (13) and (14). This is particularly the case since the observed misfits must be seen as minima: both compaction and (probable) recent increases in tidal range will serve to raise the reconstructed altitude of these SLIPs yet further above modelled RSL (See sections 7.1.2(iii) and 7.1.2(iv)). Instead a possible solution to this problem may be found by reconsidering the (mid) Holocene component of the global ice model as this input parameter drives eustatic change (which is the dominant signal in the Shannon Holocene RSL record). This approach is justified since the exact timing of the mid-Holocene slow down in RSL rise is unclear: in the model employed in these analyses, the most pronounced slow down occurs at 6000 BP although others have suggested that this occurred earlier at c.7000 BP (e.g. Peltier, 2002; Lambeck, 2002; Milne *et al.*, 2005). If the global ice model adopted in these analyses was amended to incorporate pronounced slow down in RSL rise at 7000 BP, the fit with SLIPs (13) and (14) would be greatly improved and may also reduce the model misfit with SLIP (15).

7.1.4 Postglacial Landscape Evolution of the Inner Shannon and Fergus Estuary

Despite the caveats on data reliability outlined in section 7.1.2, the good agreement between the MOSTAP core stratigraphy, the pre-existing (NRA) lithostratigraphic evidence and predictions of RSL made by the glacial rebound model enable a broad model of landscape evolution to be advanced.

Contrary to the inferences of Taylor *et al.* (1986) (Figure 7.3), the glacial rebound modelling analyses undertaken in Chapter 5 suggests that during the Lateglacial period the inner Shannon estuary occupied a position well above pmsl with landscape morphology controlled by terrestrial processes (Figure 7.1). During this period the topography of the inner estuary was characterized by a series of low limestone hills (which have subsequently been covered by thick Holocene minerogenic and organic sequences) (Figure 6.10a-c). River morphology is likely to have been strongly influenced by these bedrock knolls and river flow would have been enhanced by a series

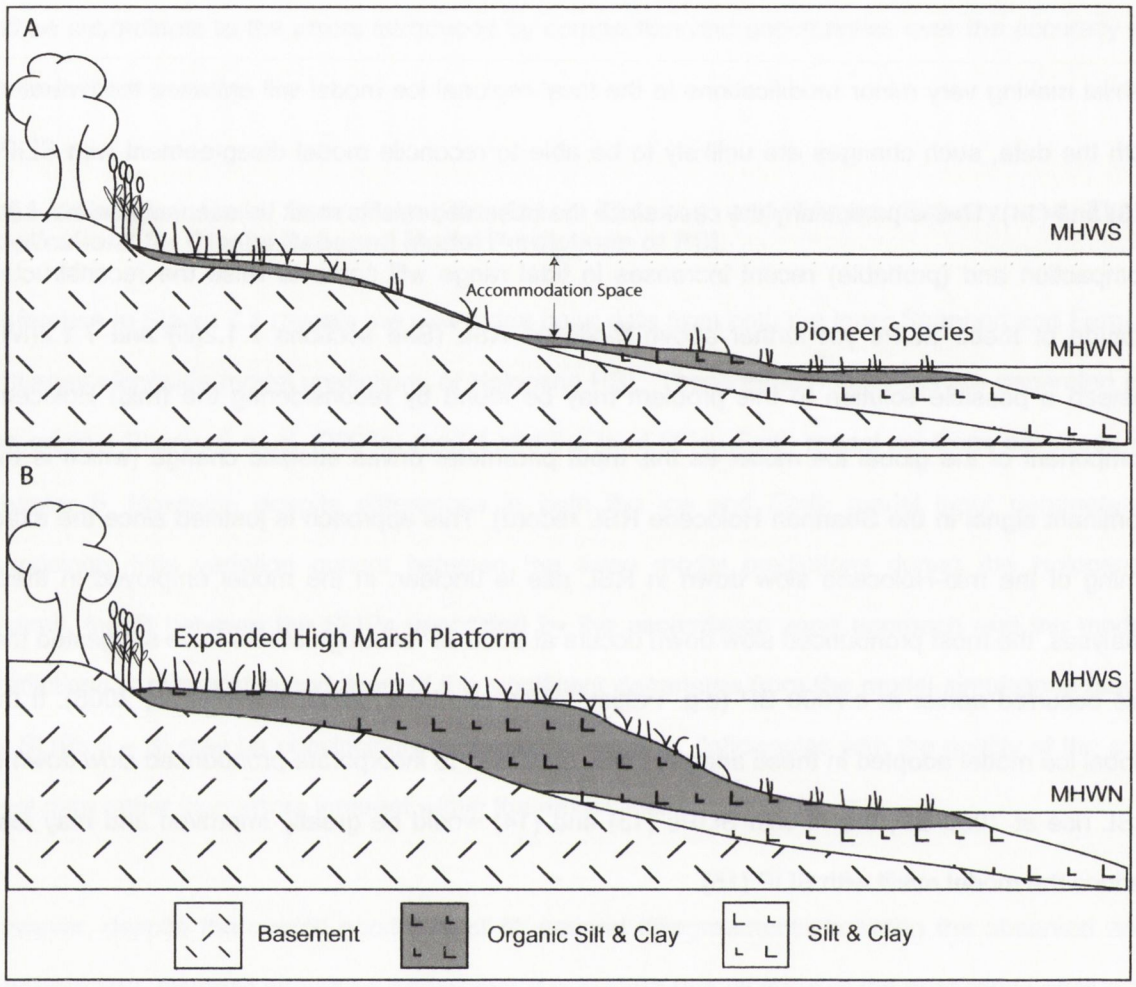


Figure 7.5 A sketch of the evolution of a typical minerogenic salt-marsh: **A** Pioneer species colonise unvegetated inter-tidal flats, enhancing sediment accumulation, and promoting infilling of available accommodation space. **B** Sediment accumulation results in an elevated salt-marsh surface and an expanded platform colonised by high marsh species. An increased proportion of high marsh environments indicates a mature salt-marsh. (Taken from Edwards, 2006b)

of well-defined tributaries (such as at Ballinacurra and Meelick) which converged with the main river channel. Fluvial erosion by these adjoining river systems probably accounts for the absence of till in several of the limestone depressions seen in Figure 6.10c. However, a subsequent change in river energy regime and/or sediment availability has bought about the deposition of the (very) course clastic units seen directly overlying the till in several of the existing NRA cores (Figure 6.10a) and at the base of the MOSTAP cores (Figure 6.11). These sand and gravel units which frequently support large clasts are barren of foraminifera and appear strongly indicative of a terrestrial environment.

At the beginning of the Holocene, the inner Shannon still occupied a position upstream of the tidal head of the river. Scrubland vegetation would have been pervasive since the region had yet to be colonized by woodland species such as oak, pine and elm (Mitchell, 2006). Species such as birch would soon have flourished in the warmer conditions associated with the termination of the Younger Dryas stadial at 11 450 BP (O'Connell *et al.*, 1999) whilst at this time, the topographic bedrock hollows impeded drainage and led to the extensive development of peat beds adjacent to the Shannon river (Figure 6.10a and Figure 6.11). The first unequivocal evidence for the invasion of marine conditions into the inner Shannon comes from the paired shell and foraminifera dates (20 and 21) in MS05. These carbonate deposits date to c. 7800 BP although the date from the regressive contact in core MS03 (SLIP 15) suggests the encroachment of the sea into the inner Shannon may have occurred by 9000 BP. Regardless of the exact time at which marine inundation occurred, the sea level data (augmented by the glacial rebound modelling evidence) suggest sea level was rising quickly at this time (Figure 7.1). Rapidly rising sea level would have raised the local ground-water table, encouraging waterlogging of the land on the margins of the estuary. These conditions would have promoted (freshwater) peat growth in supra-tidal settings and evidence of this is perhaps seen at the base of core MS02 (Figure 6.11 and 6.12a). Throughout this period, the rapid rate of RSL rise would also have provided large volumes of accommodation space within the estuary (Figure 7.5) and this may have caused much of the inter-tidal zone to be dominated by low-tidal mudflats during the early Holocene as the estuary expanded. Subsequently, this accommodation space was rapidly infilled with fine-grained clastic sediment, particularly as the rate of RSL rise slowed during the mid-Holocene. Evidence of this is found in both MS03 and MS05 where lengthy (c. 10m) and largely homogenous silt/clay units are a major element of the stratigraphic architecture. Combined, the foraminifera, shell and peat dates from MS05 suggest

that between c.7500 BP and 6500 BP, over 10m of sedimentation occurred which equates to an accumulation rate of 10 mm yr^{-1}. Almost identical units are also seen in the deep auger cores from Islandmagrath in the Fergus estuary (Figure 6.14a-c). However, as with the lengthy clastic units found in MS03 and MS05, the foraminiferal evidence is unable to distinguish between an inter-tidal and sub tidal setting for these sequences.

As the rate of RSL rise slowed and as the accommodation space (created by early Holocene RSL rise) became in filled, an episode of organic sedimentation occurred between c. 7000 - 3500 BP. Similar phases of mid-Holocene peat and brackish peat formation have been reported in numerous other studies from both intermediate and near field localities and also commonly form a dominant component of the mid Holocene stratigraphy (e.g. Devoy, 1979; Allen, 1987; Long *et al.*, 2000; Long, 2001). Evidence of this organic depositional phase can be seen in Figure 6.10a-c (existing NRA logs) although is best shown in Figure 6.11. The organic sequences are intercalated with minerogenic horizons and reflect a fluctuating balance between peat growth and mineral sediment deposition. The abundance of reed fragments found within many of these organic units suggest that during this period, reed bed communities would have been a familiar feature of the upper estuary (Tables 6.3a-f). After 3500 BP, the litho- and biostratigraphic evidence contained within the MOSTAP cores reveals a return to a minerogenic mode of sedimentation which persists until present (Figure 6.11). This perhaps suggests that the dominant estuarine environments of the late Holocene were un-vegetated mud-banks accompanied by low salt-marsh, with high salt-marsh and/or reed-swamp communities restricted to the margins of the estuary. However, this reduction in organic accumulation after c. 3500 BP may or may not relate to a restriction of vegetation within the inter-tidal zone. Numerous other studies considering late Holocene coastal sequences in the UK have observed a similar trend in reduced organic content after c. 3000 BP (Devoy, 1979; Allen, 1991; Beets *et al.*, 1992; Shennan, 1994; Baeteman, 1998; Long *et al.*, 2000). This phenomenon has variously been interpreted in terms of a reduction in organic preservation (relating to stagnation in RSL) along with greater anthropogenic activity or climatic changes promoting an increase in minerogenic sedimentation (Long *et al.*, 2000). Since the 18th century, embankment of much of the estuary is likely to have reduced the flooding frequency of the upper part of the inter-tidal zone which will have retarded salt-marsh accretion (Delaney *et al.*, in press).

Number	Sea Level database Category	Site	Material	Lab. Code	Calibrated age (yrs BP)				Alt (m) ODB	Max/Min extent of marine influence	Indicative Meaning (m)	Max/Min height msl (m)	Error (m)
					^{14}C age	Max.	Mean	Min.					
1	Limiting date I	Rahone	Peat	Beta 216640	4050±40	4798	**4610**	4421	1.26	MHWST	2.17	**-0.91**	0.34
2	Limiting date I	Rahone	Peat	Beta 216641	4060±40	4801	**4613**	4425	1.09	MHWST	2.17	**-1.08**	0.34
3	Limiting date I	Rahone	Peat	Beta 216642	4880±40	5712	**5600**	5488	0.22	MHWST	2.17	**-1.95**	0.34
4	Limiting date I	Blackweir Bridge	Peat	Beta 216646	7530±40	8411	**8311**	8211	0.2	MHWST	2.36	**-2.16**	0.34
5	Limiting date I	Blackweir Bridge	Peat	Beta 216645	8090±40	9132	**8957**	8782	-1.37	MHWST	2.36	**-3.73**	0.34
6	Limiting date I	Moyasta	Peat	Beta 216643	8850±40	10158	**9951**	9744	0.27	MHWST	2.36	**-2.09**	0.34
7	Limiting date I	Moyasta	Peat	Beta 216644	9070±40	10276	**10228**	10180	0.03	MHWST	2.36	**-2.33**	0.34

Table 7.3 Limiting dates from the outer Shannon estuary

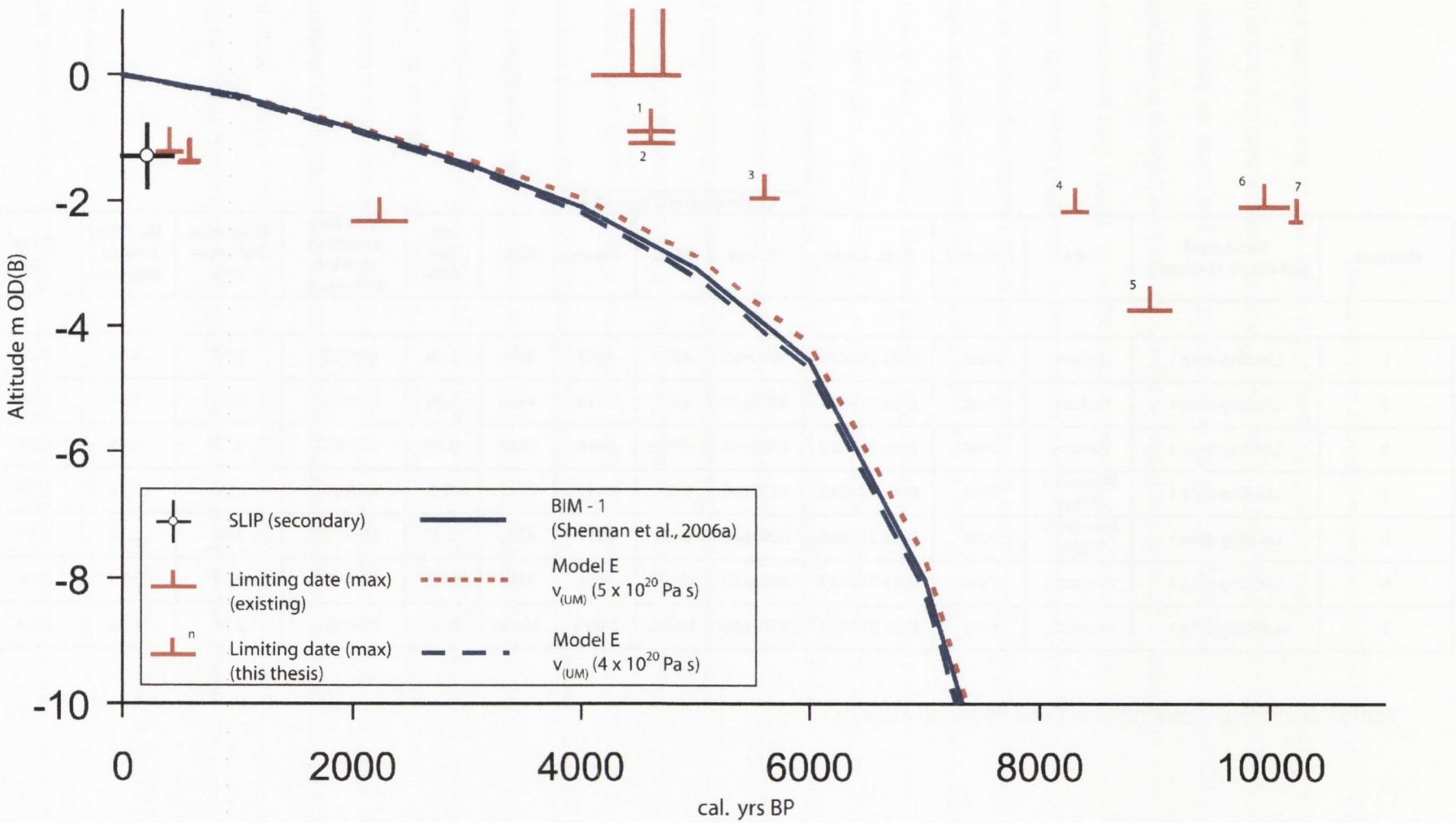


Figure 7.6 Reconstruction of Holocene RSL for the outer Shannon estuary. GIA model BIM -1 (Shennan *et al.*, 2006a); Model E v_{UM} (5 x 10^{20} Pa s) and Model E v_{UM} 4 x 10^{20} Pa s are also shown. Numbers relate to those given in table 7.3

7.2 The Evidence for Sea-Level Change in the Outer Shannon Estuary

At all of the three sites from the outer Shannon (Blackweir Bridge, Moyasta and Rahone), biostratigraphical evidence is either absent or does not demonstrate an unequivocal relationship between the dated deposit and sea level (Figures 6.18, 6.23a-c; Table 6.6). Instead, all seven of the dates only provide a maximum altitude for pmsl. These seven dates are all derived from peat horizons which are said to have formed no lower than MHWST (Table 7.3). These data are plotted alongside existing sea level data and newly developed GRM RSL predictions from the outer Shannon (Figure 7.6). The implications of these findings are now discussed.

The intercalated horizons of peat and clastic sediments found at Moyasta and particularly Blackweir Bridge appear typical of sites encountered along the submergent coastlines of southern Ireland and southern Britain (e.g. Long *et al.*, 2000; Sinnott, 1999; Edwards and Horton, 2000; Devoy *et al.*, in press). In these studies, accompanying chrono- and biostratigraphical evidence has shown that changes in lithostratigraphy are related to movements of Holocene RSL and that these changes in RSL have been the dominant forcing mechanism on sediment accumulation and associated site evolution. However, at both Blackweir Bridge and Moyasta, this relationship between site evolution and RSL is not evidenced by the biostratigraphical record: no foraminifera are encountered at either of the two sites whilst only extremely low counts of diatoms were found at Moyasta (Figure 6.18). The majority of these diatom counts were either indicative of a freshwater setting or unclassified. Although two diatom frustules symptomatic of a marine setting were recovered, as previously mentioned such low counts are unable to provide unequivocal evidence of the depositional environment. Accordingly, the dates from both Moyasta and Blackweir Bridge only provide limiting information on the maximum position of pmsl.

A compelling explanation for the observed patterns in the biostratigraphy comes from consideration of the chronostratigraphic evidence since all four of the dates from Poulnasherry Bay return early Holocene ages for the stratigraphic sequences. At this time, it is extremely likely that sea level was well below the height of the dated peat horizons (Figure 7.6): nearby in the inner Shannon, sea level can have been no higher than c. –15m OD(B) at 11 000 BP and differences in (Holocene) GIA between the outer and Inner Shannon are unlikely to have been significant (Figure 7.1).

It is important to note that the early Holocene age assigned to the peat unit at Moyasta is inconsistent with the previous findings of O'Sullivan (2001) which reveal a mid- to late Holocene age for the exposed inter-tidal peat horizon found approximately 200m to the south of transects MOY1 and MOY2 (Figure 6.16b). O'Sullivan (2001) interprets the changes in lithostratigraphy in terms of both climatic fluctuations *and* rising sea level which contrasts with the stratigraphic evidence laid out in section 6.1.5(i). However, all four of the dates presented here from Poulnasherry Bay form a chronologically consistent and coherent pattern which lends support to their authenticity (Table 6.6). Furthermore, O'Sullivan (2001) dated (submerged) Scots Pine tree stumps nearby to transects MOY1 and MOY2 but importantly, no pine tree stumps are encountered in the dated peat sections described in Figure 6.17. This is significant since Pine woodland was not found in this region at c.10 000 BP and its presence in the peat unit would necessarily have challenged the validity of the two new dates (Mitchell, 2006). It therefore appears that the exposed inter-tidal peat unit at Moyasta is diachronous although the lack of levelling data accompanying the work of O'Sullivan (2001) does complicate direct comparison with the findings of this study.

Thus the lithostratigraphic changes observed at Moyasta (Figure 6.17) and Blackweir Bridge (Figure 6.20) are most likely to have been forced by terrestrial environmental processes and cannot be directly related to changes in sea level. The clay and peat units appear to relate to lake and mire settings, which probably would have characterised much of Poulnasherry Bay at the start of the Holocene. None of the available data contradict either of the three Earth-ice model combinations advanced in Chapter 5 although are of little use in constraining input parameters employed in the glacial rebound model.

At Rahone, the litho- and biostratigraphy is similar to the sites located in Poulnasherry Bay, although here the age of the sequence does not provide such a ready explanation for patterns observed in the biostratigraphy. The three radiocarbon dates taken from the peat horizon at Rahone provide a firm chronological framework for the deposit and are in good agreement with the radiocarbon dates collected by Pearson (1979) from tree stumps in nearby Rinevalla Bay (Table 6.2 and 6.6). However, the biostratigraphic record offers no evidence for a marine depositional environment for the light grey clay deposit consistently found beneath the peat unit: as with the site at Blackweir Bridge and Moyasta, no foraminifera are discovered whilst diatom counts are extremely low and inconclusive. As a result, the three dates from Rahone also only provide limiting

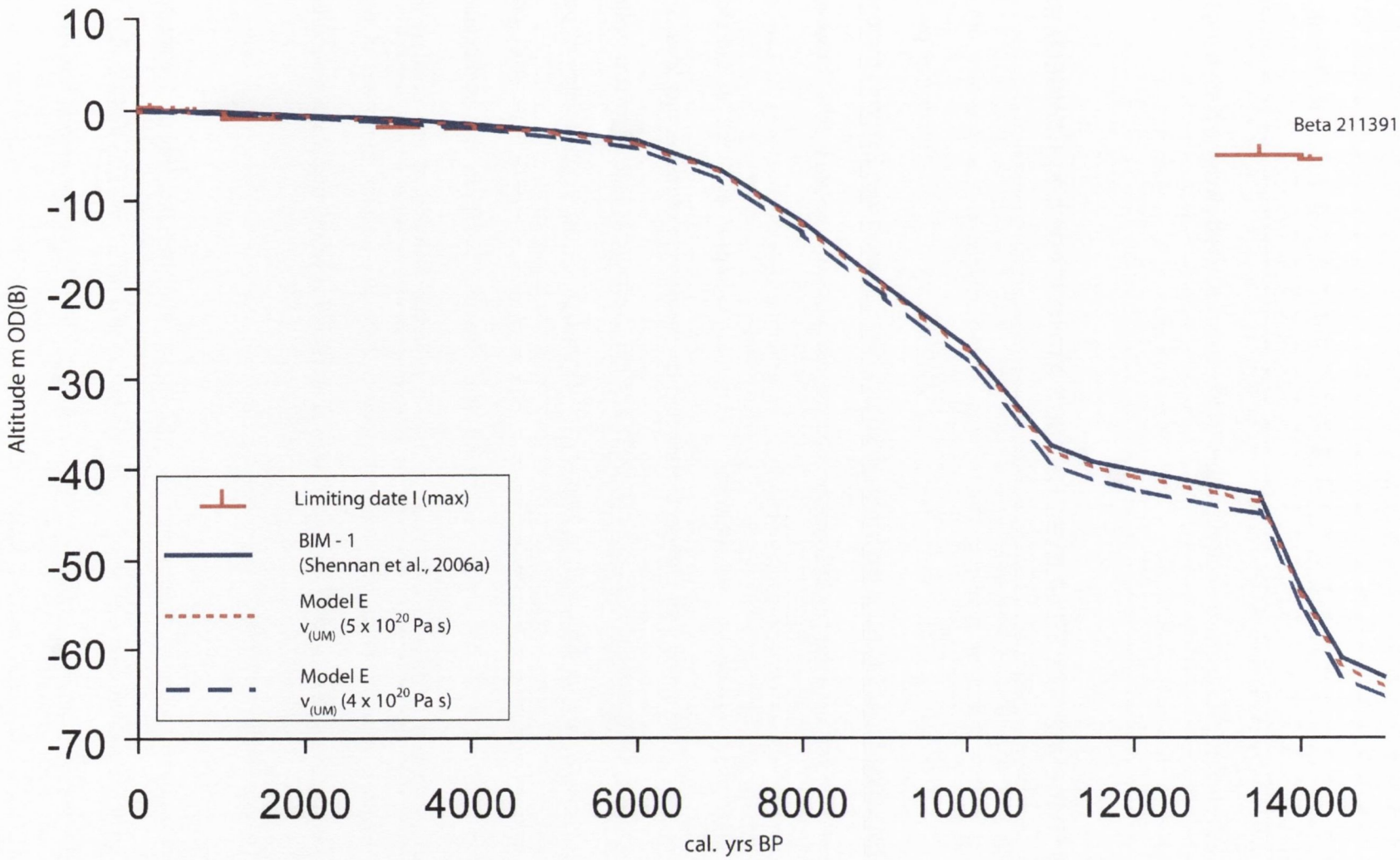


Figure 7.7 Reconstruction of postglacial sea level for Connemara. GIA model BIM -1 (Shennan *et al.*, 2006a); Model E v_{UM} (5 x 10^{20} Pa s) and Model E v_{UM} 4 x 10^{20} Pa s are also shown

information, placing a maximum altitude on the position of pmsl. However, although comparisons between the new data from Rahone and the glacial rebound model predictions of RSL change are consistent with the suggestion that this peat horizon was deposited in a supra tidal setting (Figure 7.6), poor constraints on the model in this region guard against attaching too much significance to this comparison. Moreover, sampling for contemporary foraminifera across the nearby marsh at Querrin (Figure 6.4) revealed very low counts of mid-high marsh foraminifera and both this and (possible) calcareous test dissolution would account for the absence of tests within the stratigraphic column.

7.3 The Evidence for Sea-Level Change at Rossadilisk, Connemara

The combined litho- and chronostratigraphical evidence from Rossadilisk evidences over 5m of mainly organogenic sedimentation, which began accumulating back in the Lateglacial period (Figure 6.26). Of particular interest is the nature of the regressive contacts found towards the base of several of the cores since this lithostratigraphic change may relate to changes in RSL. However, whilst diatom analysis revealed an abundance of counts within the peat unit overlying the clay, no frustules were found within the clay unit itself. The biostratigraphic evidence is therefore inconclusive and the date may only provide a maximum for the position of msl at this time. The peat unit is said to have formed at or above MHWST and the maximum altitude of pmsl is plotted alongside existing data from the region in Figure 7.7. Interestingly, similar lengthy suites of peat found in the coastal zone have been identified from elsewhere in Connemara with peat inauguration also dated to the Lateglacial period (e.g. Jessen, 1949). In the investigation undertaken by Jessen (1949), no evidence for marine influence is found at any point in the sequence which is believed to be continuous up until c. 2500 BP. Instead, the onset of peat formation can probably be related to (regional) climatic amelioration which took place around this time (Coxon 2001).

The glacial rebound model predictions of RSL support the idea that the clay-peat transition encountered at Rossadilisk has been driven by climatic factors and not by changes in RSL. At c. 14 000 BP, all three ice-Earth model solutions suggest msl was c. 50m below present and sea level is suggested to have been lower than the altitude of the regressive contact until mid-Holocene times (Figure 7.7). However, this interpretation places a strong emphasis on the veracity of Late

Devensian rebound model predictions yet the recent work of Thomas and Chiverrell (2006) puts forward evidence from this region that is grossly contrary to model predictions for this time period. Indeed, this study argues in favour of RSL at a height of up to +80m present msl and supports the earlier work of McCabe *et al.* (1986) for the existence of higher than present Late Devensian RSL along this stretch of coastline. Whilst the data presented here is consistent with the newly developed glacial rebound model predictions of RSL, the limiting date also need not contradict the findings of McCabe *et al.* (1986) and Thomas and Chiverrell (2006). However, if the grey clay unit which underlies the dated peat horizon (Figure 6.26) is found to be indicative of a marine setting, the limiting date would then be 'upgraded' to a sea level index point. This new SLIP would then be greatly at odds with GRM predictions of RSL in this locality and offer strong support to those workers advocating the occurrence of higher than present Late Devensian RSL along this stretch of coastline.

7.4 The Evidence for Sea-Level Change at Ballymichael, Donegal

Reference to the work of Shaw (1985), Lambeck (1996), Lambeck and Purcell (2001) and McCabe and Clark (2003) suggests that higher than present sea levels have occurred along this stretch of coastline during the postglacial period and these findings encouraged the search for isolation basins in this region. The visual characteristics of the site at Ballymichael, coupled with the encountered lithostratigraphy appears to be very similar in appearance to the isolation basins found along the west coast of Scotland and documented in the work of Shennan *et al.* (1993, 1994, 1995b, 1996c, 1999, 2005, 2006b). However, the interpretation of the site at Ballymichael as an isolation basin is not borne out by the biostratigraphical evidence, as diatom assemblages from the peat-clastic transition appear indicative of a freshwater depositional environment (Figure 6.31). A possible caveat to this interpretation is provided by the occurrence of mesohalobous diatoms at a depth of 298cm which may evidence occasional basin inundation by extremely high tides. However, the low percentage counts of mesohalobous taxa (<5%) and their absence elsewhere in the clay-rich peat and clastic unit suggests that their presence is more readily explained in terms of a storm event. This may have introduced saline water into the basin even if the basin sill resided at a (supra-tidal) altitude many meters above the height of pmsl. The above means that, as with the data from the outer Shannon and Rossadilisk, the date may only provide a maximum altitude for pmsl. The dated sediment sample is said to have been deposited at or above HAT (e.g. Shennan

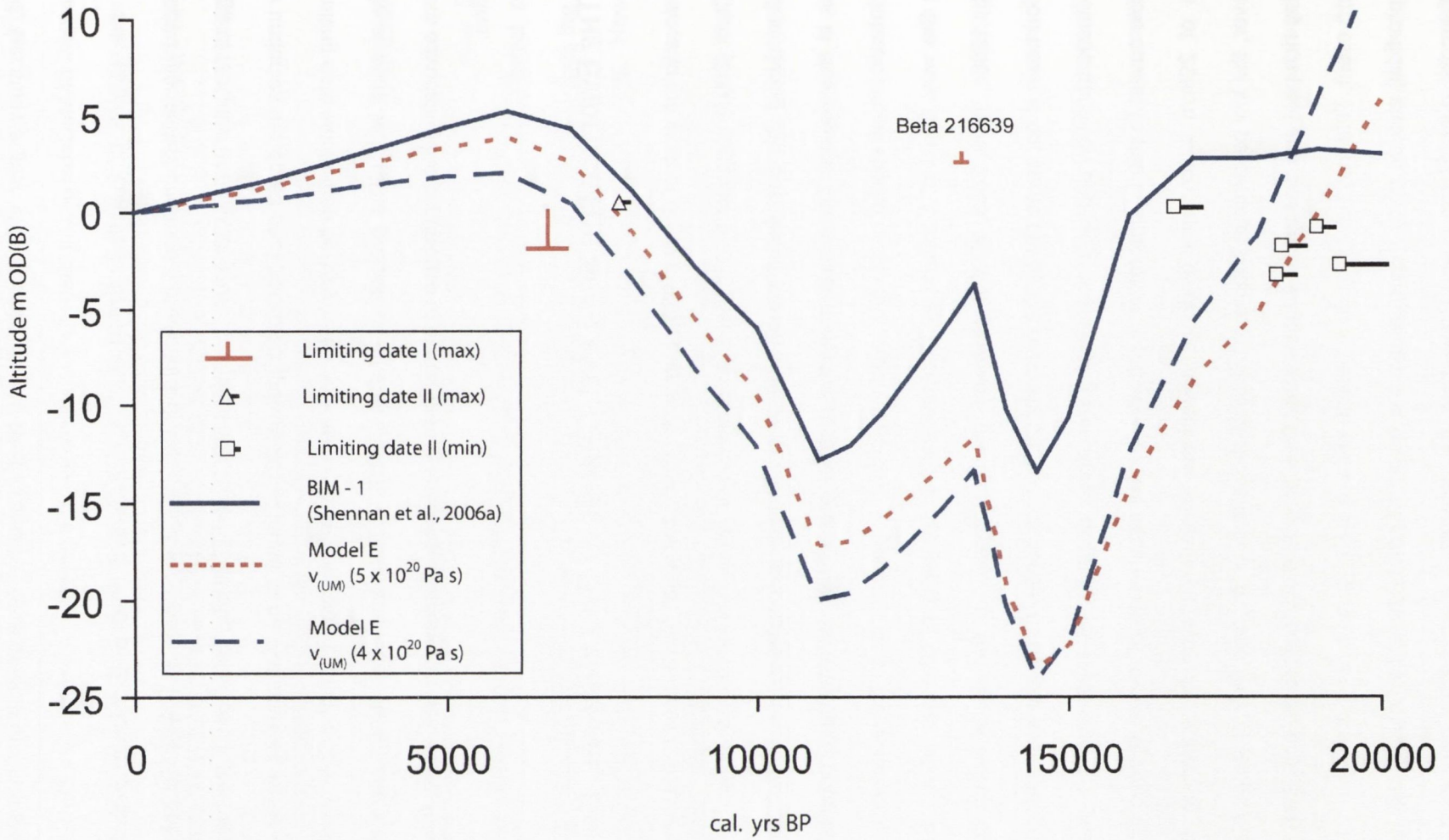


Figure 7.8 Reconstruction of postglacial sea level for the Fanad region. GIA model BIM -1 (Shennan *et al.*, 2006a); Model E v_{UM} (5 x 10^{20} Pa s) and Model E v_{UM} 4 x 10^{20} Pa s are also shown

et al., 2005) and is plotted in Figure 7.8 alongside other sea level data from the area. The interpretation of the date as limiting in nature is in good agreement with the glacial rebound model predictions of postglacial RSL change. In fact, if the available biostratigraphic data had enabled an index point instead of a limiting date to be fixed, best fit model predictions would have underestimated the height of msl by around 15m. McCabe and Clark (2003) have previously considered the deglacial history of this region and have presented a series of AMS radiocarbon dates from foraminifera which are interpreted as evidence of higher than present Lateglacial RSL. The RSL data presented by McCabe and Clark (2003) are not in total agreement with GRM model predictions of RSL, emphasizing the need for more (Late Devensian) RSL data from this region. However the new data collected here post-dates the youngest AMS radiocarbon date presented by the authors (*idem*) by c.3500 BP and so is of little help in validating their findings.

7.5 SUMMARY

- Combined litho-, bio- and chronostratigraphic evidence enables fifteen SLIPs and eight limiting data to be established from Inner Shannon and Fergus estuaries. These SLIPs have been developed using i) an assemblage zone approach (based on the contemporary foraminifera data from Ringmoylan Quay) and ii) a transfer function founded upon a regional training set of contemporary salt-marsh foraminifera (Horton and Edwards, 2006).

- The AZA based upon the modern foraminifera encountered at Ringmoylan Quay appears to be the more accurate of the two techniques used to interpret the inner Shannon (Holocene) palaeoenvironment. This is primarily because none of the contemporary marshes incorporated in the SWLI-2006 transfer function exhibit a barren zone of foraminifera in the high marsh, which is present in the contemporary marsh at Ringmoylan Quay.

- Together, the SLIP and limiting data reveal two distinct phases of RSL change in the inner Shannon during the Holocene: the first of these occurs in the early Holocene (c. 9000 – 7000 BP) and is associated with a period of rapid sea-level rise of around 5-6 mm yr^{-1}; the second phase of RSL change takes place in the mid Holocene (c. 7000 – 3000 BP) and is characterized by a much slower rate of RSL rise of c. 1.0 mm yr^{-1}.

- There exists a reasonable agreement between the newly acquired sea level data and the glacial rebound model predictions of RSL (developed in Chapter 5). No strong evidence exists to reject the basic tenets of the glacial rebound model in favour of previous (conjectural) models of postglacial RSL change for this region (e.g. Taylor *et al.*, 1986).

- However, significant departures between observed and predicted RSL occur around the time of the mid Holocene which are hard to explain in terms of errors inherent within the empirical dataset. Instead, these discrepancies *may* be related to inaccuracies in the mid-Holocene global meltwater component of the glacial rebound model.

- All the sea level evidence collated form the outer Shannon is only limiting in nature: the available biostratigraphical evidence provides no evidence for a direct link between the transitions observed in the lithostratigraphy and changes in RSL. These limiting data are in agreement with GRM predictions of RSL although are of little use as model constraints.

- Similarly, the dated (Lateglacial) sediments from both Rossadilisk (Connemara) and Ballymichael (Donegal) only provide limiting information on the maximum position of pmsl: at Ballymichael, the available diatom evidence strongly points to a terrestrial depositional environment. However, at Rossadilisk, the diatom evidence is inconclusive and further sampling is required before a definitive statement on the depositional environment of the dated sediment can be made.

- Both the limiting data from Rossadilisk and Ballymichael do not contradict the newly developed GRM predictions of RSL. At the same time however, neither of these data need contradict the suggestion that at times during the Late Devensian RSL was significantly higher than present along the west coast of Ireland.

CHAPTER EIGHT

Conclusions

8.1 THESIS SUMMARY

In contrast to the situation in Britain, Ireland suffers from a general paucity of accurate records of postglacial RSL change and many of the previously undertaken sea-level investigations from Ireland do not meet internationally recognized criteria for the collection and interpretation of RSL data (Chapter 2). Much of this work has been documented in relatively inaccessible archives and has sometimes escaped the attention of Quaternary scientists working in the coastal zone. The development of a sea-level database for Ireland (Chapter 4) has made the findings of these studies far more accessible and also permits researchers to readily identify the most reliable trends in the available information, and discriminate from potentially erroneous data. In the past, the fragmentary nature of the Irish RSL record has deterred researchers from using these data to constrain either the ice history or sub-surface viscosity structure for this region. Here, the new database has been used to help constrain glacial rebound model estimates of the spatial and temporal evolution of the Irish ice sheet (Chapter 5). The collection of empirical RSL data from the Shannon estuary and elsewhere along Ireland's west coast (Chapter 6 and 7) significantly improves the quality and quantity of sea level information available from the west coast of Ireland in the database. These new data have also been used to provide an independent test for the validity of the new Irish ice model developed in Chapter 5.

8.2 METHODOLOGICAL EVALUATION

In Chapter one, three research aims were established. The extent to which these aims have been accomplished is considered in the following section.

Aim 1: '*To compile a database of standardised sea level information from Ireland.*'

The database presented in Chapter four is the first attempt at compiling a standardized sea-level database for Ireland which is consistent with established protocols for the evaluation of sea level information. The available data are of varied quality, yet limitations associated with this are

mitigated by applying a rigorous appraisal of data errors and the consequent adoption of a quality tier system. This quality tier system discriminates four classes of sea level data on the basis of the range of fields for which information is available, and the associated precision of these data. These four classes represent the different quality of existing sea level data and reflect the extent to which the samples meet the stipulated requirements of a sea level index point. Primary index points represent the highest quality sea level data currently available from Ireland, and possess information on location, age, altitude and indicative meaning. Importantly, the relationship between the dated sea-level indicator and the environment in which it formed is quantified and error estimates for age and altitude are included. Secondary index points offer lower quality information on the former position of sea level than Primary Index Points. They are derived from dated sea-level indicators, but one or more of the core variables is unquantified or associated with significant uncertainty. The limiting date (type I) tier includes samples of known age which are associated with a known environment but this environment has no quantifiable relationship with sea level. As a consequence, it is only possible to infer whether sea level was above or below a certain altitude at a given time. Limiting date type II data differ from type I dates in that they are derived from material whose source environment is unclear or contested. Accordingly, the date may only provide a 'maximum age' for the deposit since the dated material may have been deposited then reworked several times before finally occupying its current position in the stratigraphic column.

Each database entry includes over 20 fields of information such as the geographical location of the dated sample, the age of the dated sample, the type of sea-level indicator used and the altitude of the dated sample (relative to ODB). A complete version of the database is now available online at http://www.naturalscience.tcd.ie/SL_Database.php.

Aim 2: *'To collect new, reliable RSL data from sites around the Irish coastline.'*

In Chapter six, 32 radiocarbon dates sourced from 10 field sites are presented. Combined, these dates directly evidence changes in RSL back to 9000 BP and place limits on the maximum position of RSL back to c. 14 000 BP. These data significantly augment the number of existing SLIPs from Ireland and now represent some of the oldest sea level data contained within the newly formed Irish sea-level database. However, the search for and collection of empirical sea level data was beset with problems, a fact evidenced by the large number of entries in Appendix IV ('Cored sites containing unsuitable sedimentary sequences for detailed sea-level research'). No sites suited to

the collection of postglacial RSL information were discovered on Ireland's east coast whilst very few potential isolation basins were discovered along the glacially uplifted coastline of NE Donegal. A similar pattern is observed along Ireland's west coast and the only SLIPs that were collected come from a geographically restricted area encompassing the inner Shannon and Fergus estuary.

Fifteen SLIPs have been collected from the inner Shannon and Fergus estuary and were developed using i) an assemblage zone approach (based on the contemporary distribution of salt-marsh foraminifera in the inner Shannon) and ii) a transfer function incorporating a British Isles training set of contemporary salt-marsh foraminifera (Horton and Edwards, 2006). Together with the limiting data evidence, the newly collected SLIPs enable generalized patterns of RSL change and associated phases of coastal evolution to be inferred. However, none of these SLIPs are derived from basal contexts and so are all prone to compaction. In addition, uncertainties pertaining to the contemporary distribution of foraminifera within the inner Shannon complicates reconstruction of the indicative meaning and impedes reliable estimates of the indicative range (section 7.1.2). Furthermore, those dated samples collected from elsewhere along the west coast of Ireland (outer Shannon; Connemara; Donegal) could not be directly related to movements of RSL and so only permit a maximum position for pmsl to be reconstructed.

Aim 3: *'To develop a new geophysical model capable of predicting postglacial RSL change around Ireland.'*

The development of a sea-level database for Ireland (Chapter 4) along with the emergence of new glaciological data on the spatial extent, thickness and deglacial chronology of the Irish ice sheet has enabled the successful modification of a model for the Irish component of the BIIS. Prior development of the Irish sea-level database proved to be integral to the modelling procedure particularly since it helped discriminate the most accurate and precise sea level data available from Ireland: a number of these sea level data provide useful constraints on the model input parameters yet in previous modelling attempts from Ireland their potential could not be fully realized because of uncertainties surrounding their accuracy. Accordingly, it is the contention of the author that the new model put forward in Chapter 5 represents a significant improvement over existing glacial rebound modelling studies from Ireland (e.g. Lambeck, 1996; Lambeck and Purcell, 2001).

Through combining the modelled altitude of RSL with a digital elevation model compiled from satellite altimeter and echo-sounding data, it was also possible to produce a first order palaeogeographic reconstruction for given time periods since the LGM. These reconstructions are most accurate in regions that have experienced minimal erosion or sediment reworking and provide a useful tool for elucidating the nature and timing of the postglacial colonization of Ireland.

8.3 Overview of the Main Findings of the Thesis

Uncertainties surrounding the precision and accuracy of existing RSL data from Ireland have been greatly reduced through the development of an Irish sea-level database (Chapter 4), which is based upon internationally recognized methods for sea-level research. This new database should represent a valuable tool to a range of user groups including coastal scientists, archaeologists and geophysicists interested in understanding patterns of GIA in Ireland. However, users of the database should be aware of the limitations (in precision and accuracy) of the data points it contains, and should interpret these in light of established sea-level methodology. For example, of the 200 records contained within the database, less than a quarter are primary SLIPs and the majority of the existing data must be considered as 'limiting' in nature. A number of previous studies have treated these limiting data as index points, and in some instances this has contributed to misleading conclusions being reached. Furthermore, there is a marked degree of spatial and temporal bias in the dataset. Those data occupying the primary SLIP tier are almost exclusively found along the south coast of Ireland and are nearly all of mid/late Holocene age.

It has been shown that this newly compiled Irish RSL database does provide an important data set with which to test GIA models for Ireland and the UK (Chapter 5). The information contained within the sea-level database has been used to show that the Irish ice sheet depicted in Shennan *et al.* (2006a) is largely incompatible with observations of Late Devensian and Holocene RSL from Ireland. Instead, in order to deliver a better fit between both the glaciological *and* RSL data from Ireland, a 'best fit' ice model comprised of a thick, spatially extensive Irish ice sheet of around 700m over much of north and central Ireland at the LGM is required. The sensitivity analyses undertaken in Chapter 5 also suggested that predictions of RSL are responsive to the specific rate and timing of deglaciation prior to the LGM with the Irish RSL data strongly favouring very rapid deglaciation after 21 000 BP.

Whilst modifications to the existing Shennan *et al.* (2006a) ice model have bought about significant improvements in the fit between the empirical RSL data and modelled RSL, only a limited overall fit with the available RSL data could be achieved. Important discrepancies with RSL data could not be resolved, particularly those in relation to the geographical position of the 0m Holocene isobase in both Donegal and the Irish Sea basin. However, it was possible to solve this problem by adopting a slightly lower value for the viscosity of the upper mantle in the Earth model. Despite the significant changes in RSL predictions brought about through making this adjustment to the Earth model, the results still conclusively demonstrate that the Irish component of BIM-1 is not compatible with the RSL data. The primary conclusion from the ice modelling – the need for a thinner Irish ice sheet soon after the LGM and throughout the deglacial phase - remains a robust requirement of the RSL data irrespective of the adopted viscosity model.

Regardless of which (realistic) Earth-ice model solution is adopted, at no point during the Late Devensian are predictions of RSL along Ireland's west coast higher than present msl as suggested by McCabe (2005); McCabe *et al.* (1986; 2005) and Thomas and Chiverrell (2006). The sensitivity analysis undertaken in Chapter 5 has shown that it is not possible to generate higher than present Late Devensian RSL along this stretch of coastline without contradicting the available sea level data and glaciological constraints from Ireland. The empirical data collected from Rossadilisk (Chapters 6 and 7) does not contradict the GRM predictions of Lateglacial RSL; at the same time however, this new data need not contradict the suggestion that at times during the Late Devensian RSL was significantly higher than present in this locality.

At a glance, the Shannon estuary appears to be an environment that is ripe for the collection and analysis of Holocene RSL information since both shores of the estuary are commonly flanked by mature salt-marsh systems, which are fairly rare along Ireland's west coast. However, consideration of the stratigraphic records contained within these coastal sequences reveals a complex and sometimes fragmentary record of change that appears to evidence both marine *and* terrestrial processes.

Many of the salt-marsh encountered in the outer Shannon are underlain by wholly minerogenic sequences that cannot be used to accurately reconstruct RSL (Appendix IV). Those sites that do contain inter-bedded horizons of organic and minerogenic sediments are characterized by

extremely low microfossil counts and the observed lithostratigraphic transisitions cannot be directly related to changes in RSL. Instead the chronostratigraphic evidence supports the suggestion that terrestrial environmental processes, which are largely unrelated to movements of RSL, have forced many of these sedimentary changes. Accordingly, all of the sea level evidence collated form the outer Shannon may only be considered as limiting in nature. These limiting data are in agreement with the GRM predictions of RSL presented in Chapter 5 although are of little use as model constraints.

In the inner Shannon and Fergus estuary, the evidence for past change in RSL is more ubiquitous: a large number of the salt-marshes which flank the estuary rest upon intercalated peat and silt-clay sediment horizons which commonly contain abundant microfossil evidence. Combined litho-, bio- and chronostratigraphic evidence from five sites has enabled fifteen SLIPs and eight limiting dates to be established. Together, the SLIP and limiting data reveal two distinct phases of RSL change in the inner Shannon during the Holocene: the first of these occurs in the early Holocene (c. 9000 – 7000 BP) and is associated with a period of rapid sea-level rise of around 5-6 mm yr^{-1}; the second phase of RSL change takes place in the mid Holocene (c. 7000 – 3000 BP) and is characterized by a much slower rate of RSL rise of c. 1.0 mm yr^{-1}.

The extraction of a coherent high-resolution record of RSL change from the inner Shannon and Fergus estuary (detailing variation on a sub-millennial and decimeter scale) is complicated by a number of problems. Arguably of greatest concern is the accuracy of the reconstructed indicative meanings. An assemblage zone approach based upon the modern foraminifera encountered at Ringmoylan Quay *appears* to be a more accurate tool for the interpretation of the inner Shannon (Holocene) palaeoenvironment than the Horton and Edwards (2006) SWLI-2006 transfer function. This is primarily because none of the contemporary marshes incorporated in the SWLI-2006 transfer function exhibit a barren zone of foraminifera in the high marsh, which is present in the contemporary marsh at Ringmoylan Quay. However, the relationship between the vertical extent of this barren foraminifera zone and distance up estuary remains uncertain. Similarly, the extent to which reclamation and estuary narrowing has influenced this barren zone is also unclear. Until the contemporary spatial distribution of this barren zone is known and the factors that control its presence better understood, uncertainties will persist over the most appropriate modern analogue to decipher the palaeoenvironment of the inner Shannon.

The fit between the newly acquired sea level data from the inner Shannon and the 'best-fit' glacial rebound model predictions of RSL is reasonable. No compelling evidence exists to reject the basic tenets of the glacial rebound model in favour of previous (conjectural) models of postglacial RSL change for this region (e.g. Taylor *et al.*, 1986). However, significant departures between observed and predicted RSL occur around the time of the mid Holocene which are hard to explain in terms of errors inherent within the empirical dataset. Instead, these discrepancies might be related to inaccuracies in the mid-Holocene global meltwater component of the glacial rebound model.

As with the data from the outer Shannon estuary, the dated (Lateglacial) sediments from both Rossadilisk (Connemara) and Ballymichael (Donegal) only provide limiting information on the maximum position of RSL: no foraminifera were encountered in samples taken from either site and only low (<5%) counts of brackish diatoms were discovered in a single core sample at Ballymichael. It may be stated with some confidence that the biostratigraphical record contained within the deposits found at Ballymichael evidences a freshwater setting. However, the same cannot (yet) be said of the sequence at Rossadilisk and further microfossil sampling would be a useful line of future enquiry.

8.4 Recommendations and Priorities for Future Work in Ireland

8.4.1 Future RSL Work in the Inner Shannon Estuary

Unlike at most other locations along Ireland's Atlantic seaboard, the inner Shannon is characterized by a rich tapestry of sites possessing deep sequences of inter-leaved organic and minerogenic sequences. These sediments have been used here to provide a framework for changes in Holocene RSL. However, the age-altitude graph of RSL change proposed in Chapter 7 (Figure 7.1) is compromised by a handful of issues which must be addressed before movements of Holocene RSL in the inner Shannon can be fully understood.

Firstly, there remains a lack of early Holocene SLIP data and the bulk of dates presented in this thesis post-date 7000 BP. However, previous exploratory coring undertaken by the NRA suggests that in the inner Shannon, deep (> –10m ODB) peat beds are commonly found underlying lengthy clastic units (Figures 6.10a,b,c). These deep organic-clastic sediment transitions should be targeted by future studies and subjected to detailed biostratigraphic analysis.

Secondly, the investigation of surface salt-marsh foraminifera at Ringmoylan Quay revealed a 'barren' high marsh zone that is atypical of most salt-marsh sites from elsewhere in the British Isles. This barren zone is suggested to have been caused by a combination of relatively high freshwater input into the inner estuary coupled with prevailing macro-tidal conditions. Significantly, the majority of the inner Shannon SLIPs are based upon dated high marsh samples and so the altitude of these index points is especially sensitive to the presence of this barren zone of foraminifera. However, only one inner Shannon salt-marsh was investigated and the relationship between the vertical extent of this barren zone and distance up estuary is unknown. Future studies should therefore prioritize the investigation of contemporary salt-marsh foraminifera in the inner Shannon. These studies *must* be restricted to those few remaining areas that are free from embankment since such artificial features commonly prevent the development of a high marsh zone of fauna and flora.

Thirdly, anthropogenic alteration of the inner Shannon and Fergus estuary has been substantial and widespread reclamation has greatly restricted the lateral extent of the estuary. However, it is unclear how this has affected present day tidal parameters in the estuary. This is important since the indicative meaning method inherently assumes that changes in (Holocene) tidal range have not taken place. Data on the spatial extent of land reclaim activities (e.g. Healy and Hickey, 2002) should be combined with tidal models to provide a first order estimate for the effects of this land reclamation on tidal range.

8.4.2 Future RSL Work in the Outer Shannon Estuary

Many of the intercalations of organic and clastic sediments encountered in the inner Shannon and Fergus estuary can be convincingly explained in terms of movements of RSL. However, the same cannot be said for the lithostratigraphic transisitions evident in the sites from the outer Shannon because all of the seven dated samples possess extremely low fossil diatom counts and are barren of foraminifera. In Poulnasherry Bay, the absence of foraminifera is probably due to the dated deposits originating from supra-tidal settings. However, at other locations in the outer Shannon, low counts of foraminifera in the contemporary environment (section 7.2) coupled with calcareous test dissolution may also explain their absence from the stratigraphic record. Given that diatom analysis has proven to be an unsuccessful tool for distinguishing between marine and terrestrial

depositional settings in the fossil record, future RSL investigations from the outer Shannon may have to adopt alternative methodologies in order to develop robust SLIPs. Recently, a number of researchers have assessed the potential of using $\delta^{13}C$ and C/N analysis of sedimentary matter in coastal sequences that posses a similarly fragmentary microfossil record (e.g. Mackie *et al.*, 2005; Lamb *et al.*, 2006). Studies of contemporary high, inter-tidal and sub-tidal surface salt-marsh sediments reveal distinctly different $\delta^{13}C$ and C/N values, reflecting the contrasting sources of organic matter in relation to ground elevation within the tidal frame (Wilson *et al.*, 2005). This (comparatively crude) vertical zonation may be used to reconstruct the indicative meaning of dated sedimentary deposits and also distinguish between freshwater and marine depositional environments (e.g. Deines, 1980). This approach may well be worth considering at sites in the outer Shannon.

8.4.3 Future RSL Investigations in NE Donegal

In the northwest of Ireland, the search for isolation basins proved relatively futile. Only one marginal marine basin possessed lithostratigraphic characteristics indicative of an isolation basin yet biostratigraphical analysis provided evidence to the contrary. However, the search for potential isolation basins was focused upon bog and mire environments yet on the Fanad Peninsula in northeast Donegal, a number of low-lying lakes exist in coastal settings (e.g. Eelburn Lough, Lough Beg, Melmore Lough, Tawny Lough, Fallaneas Lough). These potential isolation basins should be targeted by future RSL studies from this region.

8.4.4 General Suggestions and Observations

The development of the Irish sea-level database has emphasized both the spatial and temporal bias in existing RSL work from Ireland. In particular, very few of the data entries pre-date the Holocene period whilst nearly all of the SLIP data are sourced from the south coast of Ireland. SLIPs presented in this thesis along with forthcoming data from Delaney *et al.* (in press) (Kerry) and Harmon (2006) (Donegal and Down) will go someway to rectifying this imbalance. However, during the search for field sites to investigate as part of this thesis, it became increasingly apparent that Ireland does not possess an abundance of coastal localities containing intercalations of organic and clastic sediments (Appendix IV). This situation contrasts with that found elsewhere in the British Isles where numerous such sequences have developed and been preserved (Shennan

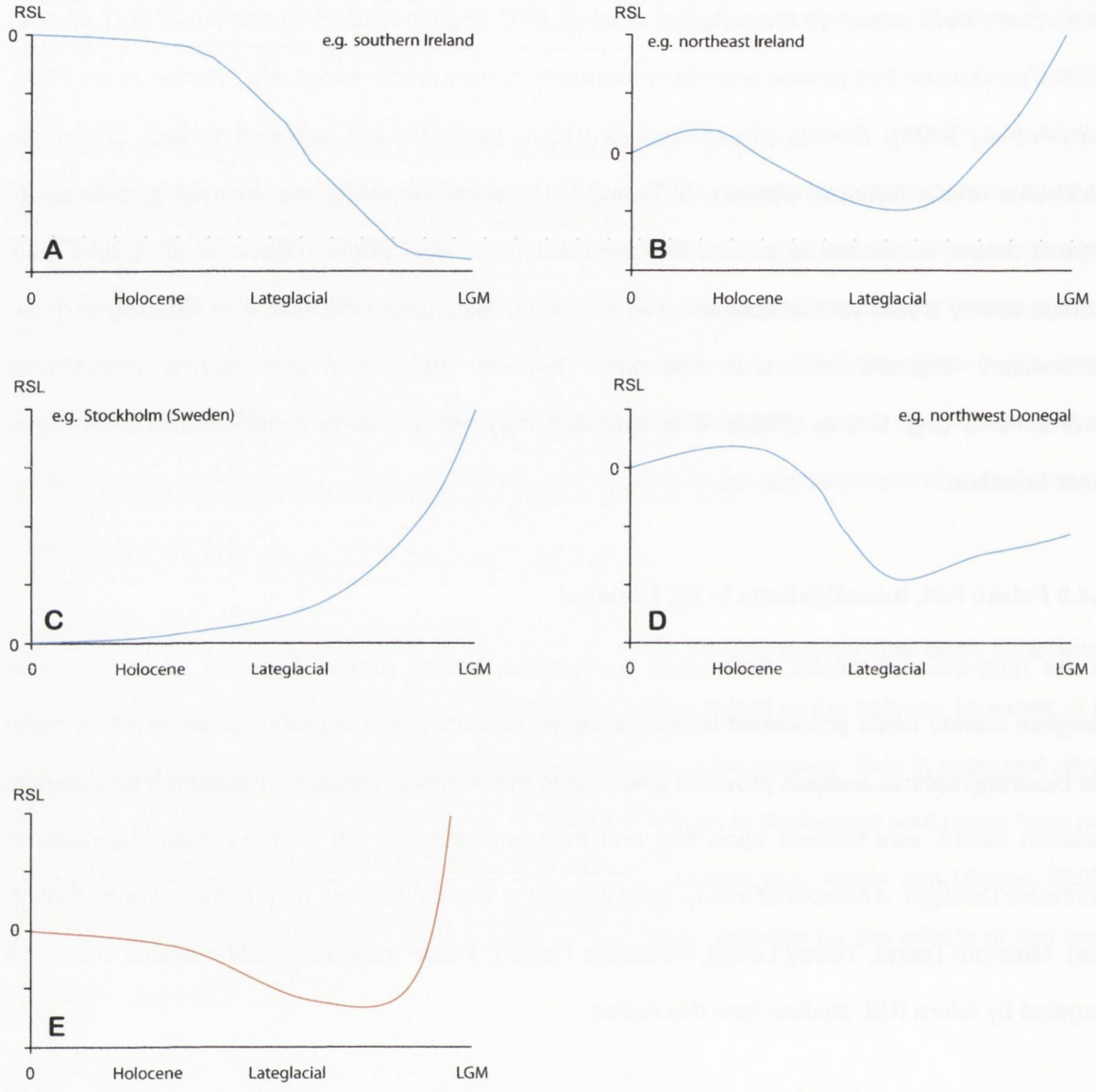


Figure 8.1 Hypothetical GRM predictions of RSL for near and intermediate field localities

and Horton, 2002). Accordingly, the information deficit identified in the database is unlikely to be erased simply by further sea-level research employing similar methods to those adopted in this thesis, which rely upon organic-clastic sediment transitions. Instead, future studies will need to utilise alternative methodologies in order to investigate a wider range of environments, which include sub-tidal and shallow marine contexts. The use of luminescence dating in conjunction with microfossil based transfer functions *may* provide a suitable avenue for future investigation along Ireland's east and west coast's where empirical evidence for change in postglacial RSL is virtually absent.

Although the best fit Earth-ice model solution (advanced in the GRM analyses in Chapter 5) yielded a very good fit with both RSL and glaciological records, a number of authors have argued for caution in the adoption of these glacial rebound models, highlighting contradictions with field data (e.g. McCabe, 1997; Smith, 2005). These dissenters dispute the 'deterministic' nature of the rebound models and argue for the existence of a more complex cryo–lithosphere interaction. One of the greatest challenges to the basic tenets of the glacial rebound modelling process is put forward by field-based investigation from the west coast of Ireland, which suggest higher than present Late Devensian RSL (e.g. McCabe, 2005; McCabe *et al.*, 2005; Thomas and Chiverrell, 2006). The reason these studies represent such a challenge to rebound modelling has less to do with the magnitude of the discrepancy with existing GRM predictions and more to do with the 'shape' of the postglacial RSL curve that they propose: if a realistic Earth-ice model combination is adopted, only four basic RSL curves are identified by rebound models from near and intermediate field localities (Figure 8.1a-d). Nowhere does the model predict Late Devensian RSL above present and Holocene RSL below present yet this is precisely the pattern that is inferred by McCabe along Ireland's west coast (Figure 8.1e). This region therefore provides a critical testing ground for the rebound modelling process and the need for more Lateglacial sea level data from here cannot be over emphasized. It remains to be seen whether the regressive contact dated at Rossadilisk (Connemara) evidences a transition from a marine to a terrestrial environment and more detailed sub sampling of the biostratigraphic record is obligatory. Future work should also involve looking for and coring (potential) isolation basin sites along this stretch of coastline since this type of marine archive is most likely to hold the oldest (postglacial) record of RSL change.

APPENDIX I

The Standardized Sea-Level Database for Ireland (CR-ROM)



APPENDIX II

Reference List of Research Included in the Irish Sea-Level Database

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APPENDIX III

The Troels-Smith (1955) Sediment Classification Scheme

(Adapted from Kershaw, 1997)

1 PHYSICAL PROPERTIES OF THE SEDIMENT

1 DEGREE OF DARKNESS *(NIGROR)*		
	0	**The shade of quartz sand**
	1	**The shade of calcareous clay**
	2	**The shade of grey clay**
	3	**The shade of partly decomposed peat**
	4	**The shade of black decomposed peat**
2 DEGREE OF STRATIFICATION *(STRATIFICATIO)*		
	0	**Complete heterogeneity**
	1	**Intermediate (between 0 – 4)**
	2	**Intermediate (between 0 – 4)**
	3	**Intermediate (between 0 – 4)**
	4	**Thin layers that spilt horizontally**
3 DEGREE OF ELASTICITY *(ELASTCIITAS)*		
	0	**Totally inelastic, plastic**
	1	**Intermediate (between 0 – 4)**
	2	**Intermediate (between 0 – 4)**
	3	**Intermediate (between 0 – 4)**
	4	**Elastic**
4 DEGREE OF DRYNESS *(SICCITAS)*		
	0	**Clean water**
	1	**Thoroughly saturated, very wet**
	2	**Saturated**
	3	**Not saturated**
	4	**Air dry**
5 SHARPNESS OF BOUNDARY *(LIMES SUPERIOR)*		
	0	**>1cm *(diffuses)***
	1	**<1cm and >2mm *(conspicuous)***
	2	**<2mm and >2mm *(manifestus)***
	3	**<1mm and >0.5mm *(acutus)***
	4	**<0.5mm**

2 Physical Components of the Sediment

Component	Latinized term	Code	Description
Clay	*Argilla steatodes*	As	Mineral particles <0.002mm
Silt	*Argilla granosa*	Ag	Mineral particles 0.002-0.06mm
Sand	*Grana minora*	Gmin	Mineral particles 0.06 - 2mm
Gravel	*Grana majora*	Gmaj	Mineral particles >2mm.
Iron oxides	*(Limus ferrugineus*	Lf	Iron oxides of various types and colours
Shells	particulae testarum molluscorum	ptm	Carbonate tests (either intact or broken)
Woody plants	*Turfa lignosa*	Tl	Roots of trees and shrubs together with attached stumps and branches. (*In situ*)
Herbs	*Turfa herbacea*	Th	Roots of herbaceous plants together with attached stems and leaves. (*In situ*)
Woody detritus	*Detritus lignosus*	Dl	Fragments of woody plants >2mm. Non vertical or random alignment. (Allocthonous)
Herb detritus	*Detritus herbosus*	Dh	Fragments of herbaceous plants >2mm.
Fine detritus	*Detritus granosus*	Dg	Fragments of woody or herbaceous plants <2mm.
Humus	*Substantia humosa*	Sh	Completely disintegrated organic substances and precipitated humic acids.

Appendix IV

Dromisign Marsh (Louth) (53 deg. 55' 22.6" N: 006 deg. 22' 31.0" W)

Core Number	Position	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Gmin	GA	Gs	Ptm	Tl	Dh	LF	Description
DRM/01	53 deg. 55' 22.6" N: 6 deg. 22' 31.0" W	0-6	2	0	0	2	N/A	2		1	1								Brown turfa rich silt:-clay (modern salt marsh)
		6-32	3	0	0	2	0	1		2	1				+				Grey-brown silt clay with turfa and rare shell fragments
		32-34	3+	0	0	2	0	++	+	3	1								Black silt clay with rare turfa and humified organic matter
		42-55	3	0	0	2	0	+	+	3	1								Dark grey-black silt: clay with rare turfa and humified organic matter
		55-72	2	0	0	2	0	+		3	1				+				Grey silt clay with rre turfa and shell fragments
		72-95	2	0	0	2	0	+		1	1		2		+				Sandy silt clay with rare turfa and occasional shell fragments
		95-204																	Increasing shells depth
		Terminates in stiff sand																	
DRM/02	30m N of DRM/01	0-14	2	0	0	2	N/A	2		1	1								Modern salt marsh. Brown turfa rich silt clay
		14-28	2+	0	0	2	0	1		2	1								Brown silt clay with turfa
		29-44	3+	0	0	2	0	1	+	2	1								Dark black silt clay with turfa and some humified organic matter
		44-79	2+	0	0	2+	0	1		2	1				+				dark grey silt clay with turfa and some shell fragments
		79-95	2+	0	0	2+	0				1		3		+				Grey silt sand with shell fragments
		95-c.200			0														Stiff sand - no recovery.
		Terminates in stiff sand																	
DRM/03	30 N again	0-8	2	0	0	2	N/A	2		1	1								Modern salt marsh surface
		8-26	3	0	0	2	0	1	+	2	1								Dark grey silt clay with some turfa and humified organic matter
		26-66	2+	0	0	2	0			3	1				+				Grey silt clay with some shell fragments and lenses of sand
		66-92	2	0	0	2	0				1		3		+				Grey silt sand with shell fragments
		92-167																	
		Impenetrable																	
DRM/04	30m N again	0-13	2	0	0	2	N/A	2+		1	1								Modern salt marsh
		13-57	2+	0	0	2	0	1		1	2								Grey clay silt with turfa
		57-90	2	0	0	2+	0	+		1	3				+				Grey clay silt with turfa and shells
		90-166	2	0	0	2+	0	+			2		2		+				Grey clay silt with turfa and shells. Sandier with depth
		Impenetrable																	
DRM/05	30m N again	0-7	2	0	0	2	N/A	3			1								Salt marsh surface
		7-23	3	0	0	1+	0	1	+	1	2								Dark grey clay silt with some turfa and a trace of humified organic matter
		23-63	2	0	0	2	0	1		2	1								Grey clay silt with turfa
		63-100	2	0	0	2+	0	+		1	1		2		+				Grey sand with some silt clay ad broken shell fragments. Trace of turfa
		100-112	2	0	0	2+	0	+		1	1		2		++				Grey sand with some silt clay ad broken shell fragments. Trace of turfa
		112-174																	
		Impenetrable																	
DRM/06	60m N again	0-3	2	0	0	2	N/A	3			1								Salt marsh surface
		3-10	2+	0	0	1+	0	2		1	1								Brown grey silt clay with turfa
		10-15	3	0	0	1+	0	1	+	2	1								Dark grey black silt clay with turfa and some humified organic matter
		15-31	2+	0	0	2	0	1	+	2	1								Same as above but peler in colour
		31-56	2	0	0	2	0			2	2								Grey silt clay
		56-74	2+	0	0	2	0				1		3		+				Stiff sand with a trace of silt and shell fragments
		74-214																	
		Impenetrable																	
DRM/07	60m N again	0-18	2	0	0	1+	N/A	2		1	1								Salt marsh surface. Brown silt clay with abundant turfa
		18-44	2	0	0	2	0	1		2	1							+	Silt clay with iron stained turfa channels
		44-69	2+	0	0	2	0	1		2	1								Grey silt clay with turfa
		69-85	2+	0	0	2	0				1		3		++				Grey silt sand with broken shell fragments
		85-282																	
		Impenetrable																	
DRM/08	60m N again (slightly to seaward)	0-12	2+	0	0	2	N/A	2		1	1		1						Salt marsh surface
		12-41	2	0	0	2	0												Same as above but slightly lighter
		41-62	2+	0	0	2	0	1		2	1							+	Silt clay with turfa and some iron staining around roots
		62-68	2	0	0	2	0	+		3	1								Grey silt clay with rare turfa
		68-74	2+	0	0	2	0			+	2		2						Grey silty sand
		74-82	2+	0	0	2	4			3	1								Dark grey silt clay with abrupt upper and lower contacts
		82-91	2+	0	0	2	4			+	2		2						Grey silty sand
		91-248																	
		Impenetrable																	
DRM/09	(53 deg 55' 33" N; 6 deg. 22' 21" W)	0-11	2	0	0	2+	N/A	2		1	1								Salt marsh surface. Brown silty clay with turfa
		12-40	2+	0	0	1+	0	1		2	1							+	Grey silt clay with some turfa and occasional iron staining
		40-69	2+	0	0	2	0	++		2	2								Silt clay with turfa
		69-100	2	0	0	2+	0			+	3		1					+	Sandy silt with a trace of clay and some broken shell fragments
		100-408																	
		Impenetrable																	

Appendix IV

Ballymascanlan (Louth) (54 deg. 01' 49" N; 006 deg. 21' 26" W

Core Number	Position	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Gmin	GA	Gs	Ptm	Tl	Dh	LF	Description
DUN/04/01	076097	0-12	2+	0	0	2	N/A	2+		1	1							+	Brown/grey organic rich silt/clay with iron staining
		12-54	2	0	0	2	0	+		3	1							++	Grey stained silt/clay
		54-75	2	0	0	2	0			3	1							++	Grey stained silt/clay with less organics than above but more iron staining
		75-136	2	0	0	2	0	+		3	1								Brown/grey silt/clay with occasional detrital plant fragments and lenses of sandy silt.
								(In lenses)		1	2		1						
		136-175	2+	0	0	2	0			1	2		1						Dark grey sandy silt with clay
		Impenetrable																	
DUN/04/02		0-11	2+	0	0	2	0	2+		1	1								Brown/grey organic rich silt clay
		11-95	2	0	0	3	0	+		2	1							1	Grey silt/clay with heavy iron staining and rare turfa
		95-135	2+	0	0	2	0				2		2					+	Grey fine silty sand with some iron staining
		135-260	2+	0	0	2	0				2		2		+			+	Grey fine silty sand with some iron staining and shell fragments
		260-278	2	0	0	2	0						2	2					Grey coasre sand with some silt?
		Impenetrable																	
DUN/04/03	077096	0-10	2+	0	0	2	N/A	2		1	1								Brown/grey organic rich silt clay
		10-110	2	0	0	2	0	+		2	1							1	
		110-120?	2	0	0	2	0			2	2							++	Grey silt/clay with iron staining. (Iron stained concretions around ?root channels?)
		?120-248	2	0	0	2	0	+		4									Grey clay with rare turfa
		248-269	2	0	0	2	0					+	2	2					Grey coarse sand
		Poor recovery																	
		Impenetrable																	

Blackrock Marsh, (Louth)(53 deg. 57' 20.5" N; 006 deg. 22' 39.0" W)

20-30cm of thin eroding marsh over bedrock or impenetrable substrate

Appendix IV

Portmarnock Bridge, (Dublin) (53 deg. 25' 03.2" N): 006 deg. 08' 25.3" W)

Core Number	Position	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Gmin	GA	Ptm	Tl	Dh	Description
PMB/04/01	53 deg 25' 03.2 (N): 6 deg 08' 25.3 (W)	0-11	3	0	0	2	N/A	2		2	+						Dark brown/Grey silt/clay wit habundant turfa
		11-24	2+	0	0	2	0	1		2	1						Brown/grey silt/clay with turfa
		24-40	2	0	0	1	0	1+		2	1						Grey silt/clay with turfa (Spartina roots)
		40-59	2	0	0	2	0	++		3	1						Grey silt/clay with turfa
		59-75	2	0	0	2	0	+		2	2			+			Grey silt/clay with a trace of turfa and shell fragments
		75-119	2	0	0	1+	0			2+	2			++			Grey silt/clay with shell fragments
		119-138	2	0	0	2	0			2	2	+		+			Grey silt/clay with coarse clasts and shell fragments
		138-150	(No recovery)														
		Impenetrable															
PMB/04/02	53 degrees 25' 01.3" (N) 6 degrees 08' 08.5" (W)	0-21	3	0	1	2	N/A	2		1+	1						Brown silt-clay with abundant turfa
		21-50	2+	0	0	2+	0	1		2	1						Grey silt/clay with turfa
		50-78	2	0	0	3	0	+		3	1						Grey clay with a trace of silt and rare turfa
		78-187	2	0	0	1+	0			+	1		3				Grey sand with silt and some clay
		Impenetrable															
PMB/04/03	53 degrees, 25' 01.6" (N) 6 degrees 08' 10.2" (W)	0-20	3	0	0	1	N/A	2	1	1	+						Brown peat with clay and a trace of silt
		20-54	2	0	0	2	0	1		2	1						Grey silt/clay with turfa
		54-78	2+	0	0	1+	0	+		2+	2						Grey silt/clay with black motting and some turfa
		78-89	2	0	0	2	0			1	2		1				Grey sand with silt and clay
		89-105	2	0	0	2	0			1	1		2				Grey sand with silt and clay
		105-133	2	0	0	1	0			1	1		2	++			Grey sand with silt and clay along with shell fragments
		Impenetrable															
PMB/04/04	53 degrees, 25' 01.0" (N) 6 Degrees, 08' 11.9" (w)	0-16	2+	0	0	2	N/A	2		1	1						Brown turfa-rich silt clay
		16-41	2	0	0	2+	0	1		2	1						Brown grey silt/clay with turfa
		41-56	2	0	0	2	0	+		3	1						Grey silt/clay with rare turfa
		56-74	3	0	0	2	0	+		3	1						Mottles black (anoxic?) silt/clay with rare turfa
		74-90	2	0	0	2	0			1	2		1				Grey silt/sand with clay
		90-102	2	0	0	2	0			1	1		2				Grey sand with silt and clay
		102-152	2	0	0	1	0			1	1		2	++			Grey sand with silt and clay and shell fragments
		Impenetrable															
PMB/04/05	53 degrees 25' 00.4" (N) 6 degrees, 08' 13.5" (W)	0-20	3	0	0	2	N/A	2		1	1						Brown turfa rich silt/clay
		20-41	2	0	0	2+	0	+		3	1						Grey silt/clay with rare turfa
		41-56	2	0	0	3	0			3	1						Grey silt/clay
		56-61	2+	0	0	2+	0			2	2						Grey silt/clay
		61-166	2	0	0	2	0			+	2		2				Grey sandy silt with some clay
		Impenetrable															

Seapoint, (Dublin) (53 deg. 27' 58.3" N; 006 deg. 12' 06.7" W)

Core Number	Position	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Gmin	GA	Ptm	Tl	Dh	Description
SPT/04/01	53 degrees, 27' 58.3" (N) 6 degrees, 12' 06.7" (W)	0-17	2	0	0	2	N/A	1		2	1						Brown/grey silt/clay with turfa
		17-21	3	0	0	2+	0	+		2	2						Dark grey silt/clay with rare turfa
		21-32	3+	0	0	2+	0			1	2		1				Dark grey /black sandy/silt clay
		32-49	4	0	0	2	0			3	1						Black silty/clay
		Impenetrable															
SPT/04/02	NE inner side of spit	0-49															
		Impenetrable (gravel)															

Appendix IV

Aughinish Farm (Shannon) (52 deg. 35' 48.8" N; 009 deg. 04' 26.9" W

Core Numl	Position Lat	Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	LF	Description
Transect 1																					
T1/04/0			0-43	2	0	0	2	NA	1		3										Brown grey clay with turfa (modern salt marsh)
			43-74	2	0	0	2	0			4						+				Fine grey clay with shelly fragments
			74-95	2+	0	0	2	0			4						+				Dark grey clay with very rare shells
			95-147	2	0	0	2	0	+		4						+				Brown grey clay with rare turfa and trace of sell fragments
			Impenetrable																		
T1/04/20	52 35 48.8	09 04 26.9	0-36	2	0	0	2	NA	1		3										Brown grey clay with turfa (modern salt marsh)
			36-45	2	0	0	2	0			4						++				Fine brown grey clay with rare shell fragments
			45-55	2	0	0	2+	0			4						+				Firm fine brown grey clay with very rare shell fragments
			55-62	2	0	0	2+	0			4						+				Firm fine brown grey clay with rare shell fragments
			62-84	2	0	0	2+	0			4						+				Very firm fine brown grey clay with very rare shell fragments
			84-95	2	0	0	2	0			3						1				Shell rich clay
			95-100	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			100-200	2	0	0	2	0			4						+				Grey clay with rare shell fragments abd occasional lenses of ptm ++
			200-265	2	0	0	2	0			4						+				Grey clay with rare shell fragments abd occasional lenses of ptm ++
			Impenetrable																		
T1/04/40	52 35 49.3	09 04 26.6	0-66	2	0	0	3	NA	1		3										Modern salt marsh unit. Very wet and limited recovery
			66-100	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			100-200	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			Impenetrable																		
T1/04/60	52 35 49.9	09 04 26.4	0-40	2	0	0	2	NA	1		3										Modern salt marsh unit. Very wet and limited recovery
			40-100	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			100-395	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			395-397	2	0	0	2	0		+	4										Brown clay
			397-464	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			Impenetrable																		
T1/04/80	52 35 50.6	09 04 26.0	0-40	2	0	0	2	NA	1		3										Poor recovery surface saltmarsh sediment
			40-459	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			459-465	3	0	0	2	4		4							+		+		Brown humified peat with traces of shell fragments. Abrupt erosive upper contact with inclusions in overlying unit
			465-478	1++	0	0	2	4			+	1			2		1				Pale grey/white sand with shell fragments, some coarser material and traces of silt/clay
			Impenetrable																		
T1/04/100	52 35 51.1	09 04 25.9	0-55	2	0	0	2	NA													Poor recovery surface saltmarsh sediment
			55-400	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			400-545	2	0	0	2	0			4						+				Grey clay with rare shell fragments. Occasional pale brown layers c.2cm thick which appear to be fine organics
			Impenetrable																		
T1/04/120	52 35 51.9	09 04 25.6	0-53	2	0	0	2	NA	1		3										Poor recovery surface saltmarsh sediment
			53-598	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			Impenetrable																		
T1/04/140	52 35 52.8	09 04 23.5	0-32	2	0	0	2	NA	1		3										Poor recovery surface saltmarsh sediment
			32-484	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			484-487	2	0	0	2	0			4				+		++				Thin, slightly more organic layer
			487-500	2	0	0	2	0			4						+				Grey clay with rare shell fragments
			500-500	2	0	0	2	0			4						+				Grey clay with rare shell fragments. Occasional organic inclusions - pehaps Th 1?
			550-583	2	0	0	2	0		++	4						+		+		Brown clay with humified organic matter detrital herbaceous matter
			583-596	2	0	0	2	0		++	4						+		+		Brown clay with humified organic matter detrital herbaceous matter. Inclusions of (eroded) humified peat?
			596-600	3	0	0	2	1		2	1				1		+				Quite abrupt change to organic clay with sand and shells and peat lenses
			Impenetrable																		

Appendix IV

Aughinish Farm (Shannon)...cont (52 deg. 35' 48.8" N; 009 deg. 04' 26.9" W

Transect 2 (Perpendicular to T1; first core 20m east of T1/04?)

(Sampling direct to base search for uneroded contact)

Core Numl	Position Lat	Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	LF	Description
T2/04/0	52 35 52.4	09 04 25.8	628-630	3+	0	0	2	4			4				++						Dark grey clay with some sand. Abrupt upper contact
			630-631	3	0	0	2	4	+	4											Brown humified turfa
			Impenetrable																		
T2/04/20	52 35 52.6	09 04 24.6	564-567	2	0	0	2	1	+	2					1		1+				Shelly sand with clay and rare turfa
			567-607	2	0	0	2	1			4		+								Brown fine clay with occasional larger clasts
			607-615	2	0	0	2	1			2		+		1		1+				Shelly sand with some clay and large clasts
			615-625	2	0	0	2	2			4		+								Dark clay with large clasts
			625-631	2	0	0	2	0	+		2		+			1	1				Shelly sandy clay with larger clasts
			631–634	2	0	0	2	0			4						+				Brown shelly clay
			Impenetrable																		
T2/04/40	52 35 52.8	09 04 22.8	-584	2	0	0	2	0			4										Buttery clay, no peat, hard base
			Impenetrable																		
T2/04/60	52 35 52.8	09 04 21.5	-551	2	0	0	2	NA			4										Grey, fine clay
			551-564	3	0	0	2	4		4											Well humified peat. Abrupt upper contact
			Impenetrable																		
T2/04/80	52 35 52.9	09 04 20.2	-511	2	0	0	2	0													No peat, coarse base
			Impenetrable																		
T2/04/100+			255-336	2	0	0	2	1	+		3						1				Very shelly clay with rare turfa
			336-345	2	0	0	2	1			4						+				Grey clay with rare shells
			Impenetrable																		
T2/04/150+			125-128	2	0	0	2	0			4						++				Grey fine clay with abundant shell fragments
			165	2	0	0	2	0			4						+				Grey fine clay with shell fragments
			Impenetrable																		

Transect 3 (Other side of channel to north, close to seabank)

T3/04/01			1.63m to base																		Very stiff, dark grey clay. Probable sub tidal deposit
			Impenetrable																		ark grey clay

Transect 4 (5m north of T2/04/60)

T4/04/0	52 35 53.0	09 04 22.5	0-50	2	0	0	2	NA	2		2										Modern salt marsh
			50-548	2	0	0	2	0	+		4						+				Grey clay with trace of shells and turfa
			Impenetrable																		
T4/04/10	52 35 52.6	09 04 22.6	-596																		
	(5m south of T2/04/60)		Impenetrable																		
T4/04/15	52 35 52.5	09 04 22.6	-555																		
			Impenetrable																		
T4/04/20	52 35 52.2	09 04 22.6	Impenetrable																		
T2/04/55	52 35 52.8	09 04 22.9	-605																		
	(5m west of T2/04/60		Impenetrable																		

Appendix IV

Barrigone, (Shannon) (52 deg. 35' 95" N; 009 deg. 03' 40" W)

Core Num	Position Lat	Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
SS/04/01	(64) 285502		0-31	2	0	0	2	NA	2	+	2									+	Iron stained, consolidated turfa with clay (modern salt mash surface)
			31-76	2	0	0	2	0	1		3									+	Turfa rich clay with iron staining
			76-100	2	0	0	2	0	+		4						+				Grey clay with rare turfa and shell fragments
			100-200	2	0	0	2	0			4						+				Grey clay with some large shell fragments
			200-370	2	0	0	2	0			4						+				Grey clay with some large shell fragments. Stiffens to base
			Impenetrable																		
SS/04/02	(64) 284501		0-23	2+	0	0	2	NA	2	1	1									+	Salt marsh surface
			23-65	2	0	0	2	0	1		3									+	Consolidated clay with turfa and iron staining around roots
			65-80	2	0	0	2	0	+		4										Soft grey clay with rare turfa
			80-96	2	0	0	2	0	+		4									+	Grey clay with some rooted turfa and ossociated iron staining
			96-100	2	0	0	2	0			4										Grey clay
			100-128	2	0	0	2	0			4										Grey clay with rare turfa (Th+)
			128-160	2	0	0	2	0			4						+				Silt clay with rare shell fragments
			160-180	2	0	0	2	0			4						++				Silt clay with large shell fragments
			180-390	2	0	0	2	0			4						+				Softer clay with rare shell fragments. Stiffens to base
			Impenetrable																		
SS/04/03	(64) 283501		0-20	2	0	0	2	0	2		2										Modern salt marsh
			20-80	2	0	0	2	0	+		4										Soft grey clay
			80-120	2	0	0	2	0			4						+				Very stiff clay with shell fragments
			Impenetrable																		
SS/04/04	(64) 282502		0-16	2+	0	0	2	NA	1	1	2										Brown modern saltmarsh
			16-81	2	0	0	2	0	+		4										Stiff grey clay with rootlets
			Impenetrable																		
SS/04/05	(64) 282503		0-11	2	0	0	2	NA	1	1	2									+	Iron stained salt marsh surface
			11-43	2	0	0	2	0	1		3										Organic clay
			43-120	2	0	0	2	0	+		4						+				Stiff clay with rare turfa and shell fragments
			Impenetrable																		
SS/04/06	(64) 279508		0-100	2	0	0	2	NA	+		3									1	Heavily iron staned clay with turfa
			Impenetrable																		

Appendix IV

Tyrone House (Galway) (53 deg. 12.48' N; 008 deg. 53.30' W)

Core Number	Position Lat	Position Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
TH/04/0	53 deg 12.48"	008 deg 53.30"	0-150	2	0	0	2	NA	2	2											Saltmarsh peat
			150-200	1+	0	0	2	0	+		4										Light grey clay with trace of organics
			(Bedrock?)																		
TH/04/25			c.200																		
25m			Bedrock																		Light grey clay with trace of organics
TH/04/50			c.200																		
25m			Bedrock																		Light grey clay with trace of organics
TH/04/75			c.200																		
25m			Bedrock																		Light grey clay with trace of organics

Knockgara (Galway) (53 deg. 08.72' N; 008 deg. 59.46' W)

Core Number	Position Lat	Position Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
KN/04/0	53 deg 08.72	008 deg 59.46	0-34	2+	0	0	2	NA	2	2											Saltmarsh peat
			34-90	1+	0	0	2+	0		1	3										Slightly organic clay
			90-101	2	0	0	2	2		+	3			1							Organic clay with sand
			Bedrock																		
KN/04/50			0-30	2+	0	0	2	NA	2	2											Saltmarsh peat
50m			30-90	1+	0	0	2	0		1	3									+	Slightly organic clay
			90-100	2+	0	0	2	2		1	2		+	1							Dark, slightly organic clay with sand/gravel
			Bedrock																		
KN/04/100			0-33	2+	0	0	2	NA	2	2											Saltmarsh peat
50m			33-75	1+	0	0	2	0		1	3										Slightly organic clay
			75-92	2	0	0	2	0		1	3										SAA except slightly lighter coloured
			92-100	1	0	0	2	2		+	4										Lighly clay clay with a trace of organics
			Bedrock																		
KN/04/150			0-110	2+	0	0	2	0	1	3	+										Saltmarsh peat
50m			110-122	1	0	0	2	4	1		3										Very light creamy coloured clay
			122-131	2	0	0	2	1		2	2										organic clay
			Bedrock																		
KN/04/200			0-30	2	0	0	2	0	2	2	+										Saltmarsh peat
50m			30-71	1+	0	0	2	0	+	2	2										Light coloured organic clay
			Bedrock																		

Appendix IV

Kilcaimin (Galway) (53 deg. 14.18' N; 008 deg. 56.73' W)

Core Number	Position Lat	Position Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	LF	Description
KM/04/0	53 deg 14.18"	008 deg 56.73"	0-49	2	0	0	2	NA	2	2											Saltmarsh unit
			49-201	2+	0	0	2	0	1	3											Clay with some organics
			201-243	2	0	0	2	0		4											Soft grey clay
			Bedrock																		
KM/04/40			0-51	2	0	0	2	NA	2	2											Saltmarsh unit
40m; 315 degrees			51-198	2+	0	0	2	0	1	3											Clay with some organics
			198-260	2	0	0	2	0		4											Soft grey clay
			Bedrock																		

Appendix IV

Streamstown (Connemara) (53 deg. 30.37' N; 010 deg. 02.44' W)

Core Number	Position Lat.	Long.	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Gmin	Gmaj	Ggmaj	GA	Gs	Ptm	Tl	Dh	LF	DI	Description
SB/05/1	53 deg. 30.37N	010 deg. 02.44 W	0-5	strat. Conf.																			
			5-39	3	0	0	2	0	2	2												+	partially humified peat
			39-200	3+	0	0	2	0	2	2												+	dark partially humified peat
			bedrock																				
SB/05/2	53 deg. 30.36N	010 deg. 02.47 W	0-20	2+	0	0	2	NA	3	1												+	partially humified peat with a trace of clay
			20-90	3+	0	0	2	0	2	2													partially humified dark peat
			90-102	3	0	0	2	0	2	1	1												very organic clay
			102-182	3	0	0	2	0	2	2												++	Partially humified peat
			bedrock																				
SB/05/3	53 deg. 30.36N	010 deg. 02.49 W	0-20	3	0	0	2	NA	1	3													very organic unit
			20-100	3+	0	0	2	0	2	2													poorly humified peat
			100-170	3	0	0	2	0	1	2												1	Poorly humified dark peat
			imp sub																				
SB/05/4	53 deg. 30.35N	010 deg. 02.50 W	0-20	3	0	0	2	NA	2	2												+	Darkish organic unit
			20-300	3+	0	0	2	0	1	3												+	Dark partially humified peat
			300-365	3	0	0	2	0	+	2												2	wood peat
			365-370	2	0	0	2	0	+	2	2		+										dark organic clay
			370-372	1	0	0	2	0					3	1									light sand unit
			imp sub																				
SB/05/5	15m SW of 4		0-20	3	0	0	2	NA	1	1	2												salt marsh unit
			20-272	3+	0	0	2	0	2	2												+	poorly humified peat (occ. DI 1)
			272-350	3	0	0	2	0	1	2	1												very organic clay
			350-390	3	0	0	2	0		2	1	1											organic clay unit
			390-(418)	1	0	0	2	4				++	4										coarse sand unit
			bedrock																				
SB/05/6	53 deg. 30.36N	010 deg. 02.50 W	0-10	3	0	0	2	NA	2	2													salt marsh unit
			10-150	3+	0	0	2	0	2	2												+	dark partially humified peat
			150-225	3+	0	0	2	0	1	2												1	dark partially humified woody peat
			225-300	3	0	0	2	0	1	2	1											++	very organic unit with wood fragments
			300-344	3	0	0	2	0	1	2	1											+	very oragnic unit
			344-347	1+	0	0	2	4				1	2	1									coarse clastic unit
			imp sub																				
SB/05/7	53 deg. 30.37N	010 deg. 02.53 W	0-20	2+	0	0	2	NA	2	2													organic unit
			20-85	3+	0	0	2	0	1	3												++	Partially humified peat
			imp sub/bedrock?																				
SB/05/8	53 deg. 30.37N	010 deg. 02.43 W	0-20	3	0	0	2	NA	2	2	+												organic unit
			20-28	1+	0	0	2	0				1	2	1									coarse clastic unit
			bedrock																				

Appendix IV

Carney (Sligo) (54 deg. 20.18' N; 008 deg. 32.36' W)

Core Number	Position Lat	Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
CAR/04/0	54 deg 20.18	008 deg 32.36	0-95	2+	0	0	4	NA		++	1			3							Clayey Sand with lenses of organics
			Impenetrable substrate																		
CAR/04/40			0-13	2	0	0	3	NA	3					1							Organic sand
40m; 100 deg			13-82	2+	0	0	3	0	1	1	1			1							Saltmarsh peat
			82-107	2+	0	0	3	0		2	2			+							Organic clay with sand
			107-171	2	0	0	2	0		+	1			3							Clayey sand with a trace of organics
			Impenetrable substrate																		
CAR/04/65			0-18	2+	0	0	2	NA	2	2				+							Saltmarsh peat
25m; 100 deg			18-85	3	0	0	2	0	+	2	2										Organic clay
			85-109	3	0	0	2	0		2	2			+							Dark organic clay with a trace of sand
			109-139	2	0	0	2	0			1			3							Clayey sand
			Impenetrable substrate																		

Drumcliff Bay (Sligo) (54 deg. 19' N; 008 deg. 30.5' W)

DB1/04/0			0-48	2+	0	0	2	NA	2	2				+							Saltmarsh peat
			48-100	2	0	0	3	0			3			1							Very stiff clay/sand
			Impenetrable substrate																		
DB1/04/50			0-53	2+	0	0	2	NA	2	2				+							Saltmarsh peat
50m; 140 degrees			53-100	2	0	0	3	0			3			1							Very stiff clay/sand
			Impenetrable substrate																		
DB2/04/0	G670409		0-40	2+	0	0	2	NA	2	2				+							Saltmarsh peat
			40-150	2	0	0	3	0			3			1							Very stiff clay/sand
			Impenetrable substrate																		
DB2/04/50			0-45	2+	0	0	2	NA	2	2				+							Saltmarsh peat
50m; 270 degrees			45-150	2	0	0	3	0			3			1							Very stiff clay/sand
			Impenetrable substrate																		
DB2/04/100			0-55	2+	0	0	2	NA	2	2				+							Saltmarsh peat
50m; 270 degrees			55-150	2	0	0	3	0			3			1							Very stiff clay/sand
			Impenetrable substrate																		

Appendix IV

Ballysadare Bay (Sligo) (54 deg. 12.55' N; 008 deg. 34.41' W)

Core Num	Position Lat	Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
BB/04/0	54 deg 12.55"	008 deg 34.41"	0-17	2+		0	0	2+	NA	2	2										Salt marsh peat
			17-46	2		0	0	2+	0		2	2									Organic clay
			Impenetrable substrate																		
BB/04/25			0-26	2+		0	0	2+	NA	2	2										Saltmarsh peat
25m			26-61	2		0	0	2+	0	1	2	1									Organic clay
			61-91	2		0	0	2+	0	+	1	3		+							Slightly less organic clay
			Impenetrable substrate																		
BB/04/50			0-130																		SAA but shell lenses at c.120
25m			Impenetrable substrate																		
BB/04/75			0-32	2+		0	0	2+	NA	2	2										Salt marsh peat
25m			32-116	3		0	0	2+	0	+	3	1		+							Clayey organic layer
			116-192	2		0	0	2	0			1		3							Clayey sand
			Bedrock																		
BB/04/100			0-200	2		0	0	2	0			1		3							Clayey sand
25m			Bedrock																		

Ballysadare Bridge (Sligo) (54 deg. 13.1' N; 008 deg. 30.60' W)

Core Num	Position Lat	Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
BBR/04/0	54 deg 13.31"	008 deg 30.28"	0-117	2+		0	0	2	NA	2	2										Saltmarsh peat
			117-176		2	0	0	2	0	1	1	2									Organic clay
			176-215	1+		0	0	2	0			2	1	1							Light grey sandy clay with occasional organics
			215-235	2+		0	0	2	0	1	1	2									Organic clay layer
			235-300	1+		0	0	2	0			2	1	1							Sandy clay layer with some silt
			Impenetrable substrate																		
BB/04/100			c.0-100	2+		0	0	2	NA	2	2										Saltmarsh peat
			c.100-180			0	0			+		3		1				++			Sandy clay with fragments of wood
			180-230			0	0			+		3		1							Sandy clay
			230-380			0	0					1		3							Clayey sand
			Impenetrable substrate																		

Appendix IV

Brughmore (Sligo) (54 deg. 13.62' N; 008 deg. 32.36' W)

Core Number	Position Lat	Position Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
PSB/04/0	54 deg 13.62 N	008 deg 32.36W	0-19	2+	0	0	2	NA	2	2											Saltmarsh peat
			19-41	2	0	0	2	0	1	1	2										Organic clay
			41-100	2+	0	0	2	0		2	2	+		+							Organic clay
			Impenetrable substrate																		
PSB/04/50			0-53	2+	0	0	2	NA													Saltmarsh peat
50m; 180 degrees			53-100						1	1	2			+							Organic clay
			100-125	2	0	0	2	0		1	3			+			+				Slightly organic clay
			125-145	1+	0	0	2	0			1				3		+				clayey sand
			Impenetrable substrate																		
PSB/04/100			0-119	2+	0	0	2	NA	2	2											Saltmarsh peat
50m; 180 degrees			119-144	2	0	0	2	0		1	2	1		+			+				Slightly organic clay
			144-152								1				3		+				Clayey sand
			Impenetrable substrate																		
PSB/04/120			0-40	2+	0	0	2	NA	2	2											Saltmarsh peat
20m; 180 degrees			40-147	2	0	0	2	0		1	2				1						Organic sandy clay
			147-163	1+	0	0	2	0	+		1				3		+				Clayey sand with organics
			Impenetrable substrate																		
PSB/04/320			0-50	2+	0	0	2	NA	2	2											Saltmarsh peat
200m; 135 degrees			50-130	2	0	0	2	0		1	2				1						Organic sandy clay
			130-150	1+	0	0	2	0	+		1				3		+				Clayey sand with organics
			Impenetrable substrate																		

Appendix IV

Doagh Beg (Donegal) (55 deg. 15.35' N; 007 deg. 37.0' W)

Core Number	Position Lat	Position Long	Depth (cm)	Darkness	Stratification	Elasticity	Dryness	Contact	Th	Sh	As	Ag	Ggmaj	Gmin	Ga	Gs	Ptm	Tl	Dh	Lf	Description
	55 deg 15.35"	007 deg 37.0"	0-40	4	0	0	3	NA	1	2	1										Dark organic unit
			40-70	2	0	0	2	0		2	2										light grey organic clay
			70-85	3	0	0	2	0	1	2	1										dark organic unit with some clay
			85-100	4	0	0	3	0	1	2	1										very dark organic unit
			100-150	2	0	0	2	0	+	2	2	+		+							olive coloured organic clay unit
			Bedrock																		

Lough Rinboy (Donegal) (55 deg. 14.54' N; 007 deg. 43.49' W)

LBY/04/0			0-36	3	0	0	2	NA	2	2											Brown humified peat
			36-200	3+		0	0	3	0	1	3										Well humified peat
			200-379	3+		0	0	3	0		4										Very well humified soft peat. (Gyttia?)
			379-398		2	0	0	2	0			1	+		3						Light green clayey sand
			398-400	3+		0	0	3	0		4										Very well humified soft peat. (Gyttia?)
			400-450		2	0	0	2	0			1	+		3						Light green clayey sand
			Bedrock																		
LBY/04/200			-350																		SAA
			Bedrock																		
LBY/04/300			-300																		SAA
			Bedrock																		
LBY/04/400			-100																		SAA
			Bedrock																		

Baloor (Donegal) (55 deg. 16.4' N; 007 deg. 39.0' W)

BAL/04/0			0-380	3+		0	0	3	0		4										Soft Peat
			380-420		3	0	0	3	4				2		2						Gravelly/Sand
			Bedrock																		

Lough Napaste (Donegal) (55 deg. 12.5' N; 007 deg. 46.35' W)

NAP/04/0	55 deg 12.5"	007 deg 46.35"	0-22	3	0	0	3	NA	3			1									Dark Brown organic peat
			22-67	2+	0	0	2	0	2	2		+									Soft brown peat
			67-100	2	0	0	2	0	+	1	+			3							Organic sand layer
			100-120	2+	0	0	2	0	2	2				+							Soft brown peat
			120-210	2	0	0	2	0	+	1	+			3							Organic sand layer
			Impenetrable subtstrate/bedrock?																		
NAP/04/50			0-20	3	0	0	3	NA	3			1									Dark Brown organic peat
100m; beraing 170			20-66	2+	0	0	2	0	2	2		+									Soft brown peat
			60-200	2	0	0	2	0	+	1	+			3							Organic sand layer
			Impenetrable subtstrate/bedrock?																		
NAP/04/100			0-20	3	0	0	3	NA	3			1									Dark Brown organic peat
100m; beraing 170			20-50	2+	0	0	2	0	2	2		+									Soft brown peat
			50-120	2	0	0	2	0	+	1	+			3							Organic sand layer
			Impenetrable subtstrate/bedrock?																		

Ballyhoorisky Point (Donegal) (55 deg. 15.34' N; 007 deg. 45.1' W)

Very shallow (<50cm) sediement sequences. No isolation basins present

APPENDIX V

Inventory of Radiocarbon Dates

Sample Data	Measured Radiocarbon Age	13C/12C Ratio	Conventional Radiocarbon Age(*)
Ballinacurra Creek (west)			
Beta - 203536	6330 +/- 50 BP	-27.5 o/oo	6290 +/- 50 BP
SAMPLE : SHANNO501			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 5350 to 5210 (Cal BP 7300 to 7160) AND Cal BC 5170 to 5080 (Cal BP 7120 to 7030)			
Coonagh (south)			
Beta - 203537	8110 +/- 50 BP	-27.5 o/oo	8070 +/- 50 BP
SAMPLE : SHANNO502			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 7140 to 7020 (Cal BP 9100 to 8970) AND Cal BC 6960 to 6920 (Cal BP 8910 to 8870) Cal BC 6880 to 6840 (Cal BP 8830 to 8790)			
Ballinacurra Creek (west)			
Beta - 203538	6540 +/- 50 BP	-27.6 o/oo	6500 +/- 50 BP
SAMPLE : SHANNO503			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 5520 to 5360 (Cal BP 7480 to 7310)			
Ballinacurra Creek (west)			
Beta - 203539	5560 +/- 50 BP	-26.7 o/oo	5530 +/- 50 BP
SAMPLE : SHANNO504			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 4460 to 4320 (Cal BP 6410 to 6270)			
Ballinacurra Creek (west)			
Beta - 203540	4600 +/- 40 BP	-25.5 o/oo	4590 +/- 40 BP
SAMPLE : SHANNO505			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 3500 to 3450 (Cal BP 5450 to 5400) AND Cal BC 3380 to 3330 (Cal BP 5330 to 5280) Cal BC 3220 to 3180 (Cal BP 5170 to 5130) AND Cal BC 3160 to 3130 (Cal BP 5100 to 5080)			
Coonagh (south)			
Beta - 203541	9220 +/- 40 BP	-28.1 o/oo	9170 +/- 40 BP
SAMPLE : SHANNO506			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 8470 to 8280 (Cal BP 10420 to 10230)			
Coonagh (south)			
Beta - 203542	9410 +/- 70 BP	-28.8 o/oo	9350 +/- 70 BP
SAMPLE : SHANNO507			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 8770 to 8430 (Cal BP 10720 to 10380) AND Cal BC 8360 to 8340 (Cal BP 10310 to 10280)			
Coonagh (south)			
Beta - 203543	9620 +/- 40 BP	-27.6 o/oo	9580 +/- 40 BP
SAMPLE : SHANNO508			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 9180 to 8770 (Cal BP 11130 to 10720)			
Coonagh (west)			
Beta - 203544	5660 +/- 50 BP	-25.2 o/oo	5660 +/- 50 BP
SAMPLE : SHANNO509			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 4590 to 4360 (Cal BP 6540 to 6320)			
Meelick Creek			
Beta - 216647	7110 +/- 70 BP	-7.8 o/oo	7390 +/- 70 BP
SAMPLE : SHANNMSO5C			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (foraminifera): none			
2 SIGMA CALIBRATION : Cal BC 6410 to 6080 (Cal BP 8360 to 8030)			

Sample Data	Measured Radiocarbon Age	13C/12C Ratio	Conventional Radiocarbon Age(*)
Coonagh (west)			
Beta - 203545	6410 +/- 50 BP	-25.7 o/oo	6400 +/- 50 BP
SAMPLE : SHANNO510			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (organic sediment): acid washes			
2 SIGMA CALIBRATION : Cal BC 5480 to 5300 (Cal BP 7430 to 7250)			
Coonagh (west)			
Beta - 203546	5680 +/- 40 BP	-25.6 o/oo	5670 +/- 40 BP
SAMPLE : SHANNO511			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (organic sediment): acid washes			
2 SIGMA CALIBRATION : Cal BC 4580 to 4440 (Cal BP 6530 to 6390)			
Meelick Creek			
Beta - 203547	4050 +/- 40 BP	-26.3 o/oo	4030 +/- 40 BP
SAMPLE : SHANNO512			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 2630 to 2470 (Cal BP 4580 to 4420)			
Meelick Creek			
Beta - 203548	5870 +/- 40 BP	-25.5 o/oo	5860 +/- 40 BP
SAMPLE : SHANNO513			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 4800 to 4660 (Cal BP 6750 to 6610) AND Cal BC 4640 to 4620 (Cal BP 6590 to 6570)			
Meelick Creek			
Beta - 203549	3700 +/- 40 BP	-27.5 o/oo	3660 +/- 40 BP
SAMPLE : SHANNO514			
ANALYSIS : AMS-Advance delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 2140 to 1920 (Cal BP 4090 to 3870)			
Rossadilisk			
Beta - 211391	12210 +/- 40 BP	-22.5 o/oo	12250 +/- 40 BP
SAMPLE : ROSS0501			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 13360 to 12700 (Cal BP 15310 to 14650) AND Cal BC 12420 to 12150 (Cal BP 14370 to 14100)			
Islandmagrath			
Beta - 211392	3220 +/- 40 BP	-27.9 o/oo	3170 +/- 40 BP
SAMPLE : ISLMAG0501			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 1520 to 1390 (Cal BP 3460 to 3340)			
Islandmagrath			
Beta - 211393	2520 +/- 40 BP	-25.8 o/oo	2510 +/- 40 BP
SAMPLE : ISLMAG0502			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 790 to 500 (Cal BP 2740 to 2450) AND Cal BC 460 to 430 (Cal BP 2410 to 2380)			
Ballinacurra Creek (west)			
Beta - 211394	4430 +/- 50 BP	-27.2 o/oo	4390 +/- 50 BP
SAMPLE : SHANNMS02			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 3310 to 3230 (Cal BP 5260 to 5180) AND Cal BC 3110 to 2900 (Cal BP 5060 to 4850)			
Coonagh (south)			
Beta - 211395	7840 +/- 40 BP	-27.2 o/oo	7800 +/- 40 BP
SAMPLE : SHANNMS03A			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (organic sediment): acid washes			
2 SIGMA CALIBRATION : Cal BC 6680 to 6500 (Cal BP 8630 to 8450)			

Sample Data	Measured Radiocarbon Age	13C/12C Ratio	Conventional Radiocarbon Age(*)
Coonagh (south)			
Beta - 211396	7380 +/- 40 BP	-10.3 o/oo	7620 +/- 40 BP
SAMPLE : SHANNMS03B			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (shell): acid etch			
2 SIGMA CALIBRATION : Cal BC 6490 to 6410 (Cal BP 8440 to 8360)			
Coonagh (west)			
Beta - 211397	5790 +/- 40 BP	-24.8 o/oo	5790 +/- 40 BP
SAMPLE : SHANNMS04			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (organic sediment): acid washes			
2 SIGMA CALIBRATION : Cal BC 4720 to 4530 (Cal BP 6670 to 6480)			
Meelick Creek			
Beta - 211398	3530 +/- 50 BP	-27.8 o/oo	3480 +/- 50 BP
SAMPLE : SHANNMS05A			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (organic sediment): acid washes			
2 SIGMA CALIBRATION : Cal BC 1920 to 1680 (Cal BP 3870 to 3630)			
Meelick Creek			
Beta - 211399	6810 +/- 50 BP	-5.8 o/oo	7120 +/- 50 BP
SAMPLE : SHANNMS05B			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (shell): acid etch			
2 SIGMA CALIBRATION : Cal BC 6060 to 5890 (Cal BP 8010 to 7840)			
Moyasta			
Beta - 216643	8910 +/- 40 BP	-28.5 o/oo	8850 +/- 40 BP
SAMPLE : POULBAY0501			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid washes			
2 SIGMA CALIBRATION : Cal BC 8200 to 7800 (Cal BP 10150 to 9740)			
Blackweir Bridge			
Beta - 216645	8130 +/- 40 BP	-27.5 o/oo	8090 +/- 40 BP
SAMPLE : POULBRIDGE01			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid washes			
2 SIGMA CALIBRATION : Cal BC 7140 to 7040 (Cal BP 9100 to 8990)			
Ballymichael			
Beta - 216639	NA	NA	11430 +/- 40 BP
SAMPLE : BALLY0501			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 11830 to 11730 (Cal BP 13780 to 13680) AND Cal BC 11540 to 11210 (Cal BP 13490 to 13160)			
Comment: the original sample was too small for a 13C/12C ratio measurement. However, a ratio including both natural and laboratory effects was measured during the 14C detection to derive a Conventional Radiocarbon Age, suitable for applicable calendar calibration.			
Rahone			
Beta - 216640	4100 +/- 40 BP	-28.2 o/oo	4050 +/- 40 BP
SAMPLE : CLOON0502			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 2850 to 2820 (Cal BP 4800 to 4770) AND Cal BC 2670 to 2470 (Cal BP 4620 to 4420)			
Rahone			
Beta - 216641	4090 +/- 40 BP	-26.8 o/oo	4060 +/- 40 BP
SAMPLE : CLOON0503			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 2850 to 2820 (Cal BP 4800 to 4770) AND Cal BC 2680 to 2480 (Cal BP 4630 to 4430)			

Sample Data	Measured Radiocarbon Age	13C/12C Ratio	Conventional Radiocarbon Age(*)
Blackweir Bridge			
Beta - 216646	7570 +/- 40 BP	-27.6 o/oo	7530 +/- 40 BP
SAMPLE : POULBRIDGE02			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 6450 to 6360 (Cal BP 8400 to 8300) AND Cal BC 6310 to 6260 (Cal BP 8260 to 8210)			
Rahone			
Beta - 216642	4930 +/- 40 BP	-27.8 o/oo	4880 +/- 40 BP
SAMPLE : CLOON0504			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 3710 to 3640 (Cal BP 5660 to 5580)			
Moyasta			
Beta - 216644	9130 +/- 40 BP	-28.7 o/oo	9070 +/- 40 BP
SAMPLE : POULBAY0502			
ANALYSIS : AMS-Standard delivery			
MATERIAL/PRETREATMENT : (peat): acid/alkali/acid			
2 SIGMA CALIBRATION : Cal BC 8300 to 8240 (Cal BP 10250 to 10190)			

APPENDIX VI

Reference List of British RSL Research Employed in the Glacial Rebound Model Analyses (Chapter 5)

Bedlington DJ. 1993. *Holocene sea-level changes and crustal movements in North Wales and Wirral.* Unpublished PhD thesis, University of Durham: Durham.

Birks HJB. 1982. Mid-flandrian forest history of Roudsea Wood National Nature Reserve, Cumbria. *New Phytologist* **90**: 339–354.

Healy MG. 1995. The lithostratigraphy and biostratigraphy of a Holocene coastal sediment sequence in Marazion Marsh, west Cornwall, U.K. with reference to relative sea-level movements. *Marine Geology* **124**: 237–252.

Heyworth A, Kidson C. 1982. Sea-level changes in southwest England and Wales, *Proceedings of the Geologists' Association* **93**: 91–111.

Huddart D. 1992. Coastal environmental changes and morphostratigraphy in southwest Lancashire, England, *Proceedings of the Geologists' Association* 103: 217–236.

Kidson C, Heyworth A. 1982. Sea-level changes in southwest England and Wales. *Proceedings of the Geologists' Association* **93**: 91–112.

Roberts DH, Chiverrell RC, Innes JB, Horton BP, Brooks AJ, Thomas GSP, Turner S, Gonzalez S. 2006. Holocene sea levels, Last Glacial Maximum deglacial environments and geophysical models in the northern Irish Sea Basin, UK. *Marine Geology* **231**: 113-128.

Shennan I, Hamilton S, Hillier C, Hunter A, Woodall R, Bradley S, Milne G, Brooks A, Bassett S. 2006b. Relative sea-level observations in western Scotland since the last glacial maximum for testing models of glacial isostatic land movements and ice-sheet reconstructions. *Journal of Quaternary Science* **21**(6): 601–613.

Tooley MJ. 1977. *The Isle of Man, Lancashire coast and Lake District, Field Guide A4, X INQUA Congress.* Geo. Abstracts Ltd: Norwich.

Tooley MJ. 1978a. *Sea-Level Changes in North-West England During the Flandrian Stage.* Clarendon Press: Oxford.

Tooley MJ. 1978b. Interpretation of Holocene sea-level changes. *Geologiska Foreningens I Stockholm Forhandlingar* **100**: 203–212.

Tooley MJ. 1985. Sea-level changes and coastal morphology in North-West England. In *The geomorphology of north-west England*, Johnson RH (ed). Manchester University Press: Manchester; 94–121.

Zong Y, Tooley M. 1996. Holocene sea-level changes and crustal movements in Morecambe Bay, Northwest England. *Journal Quaternary Science* **11**: 43–58.

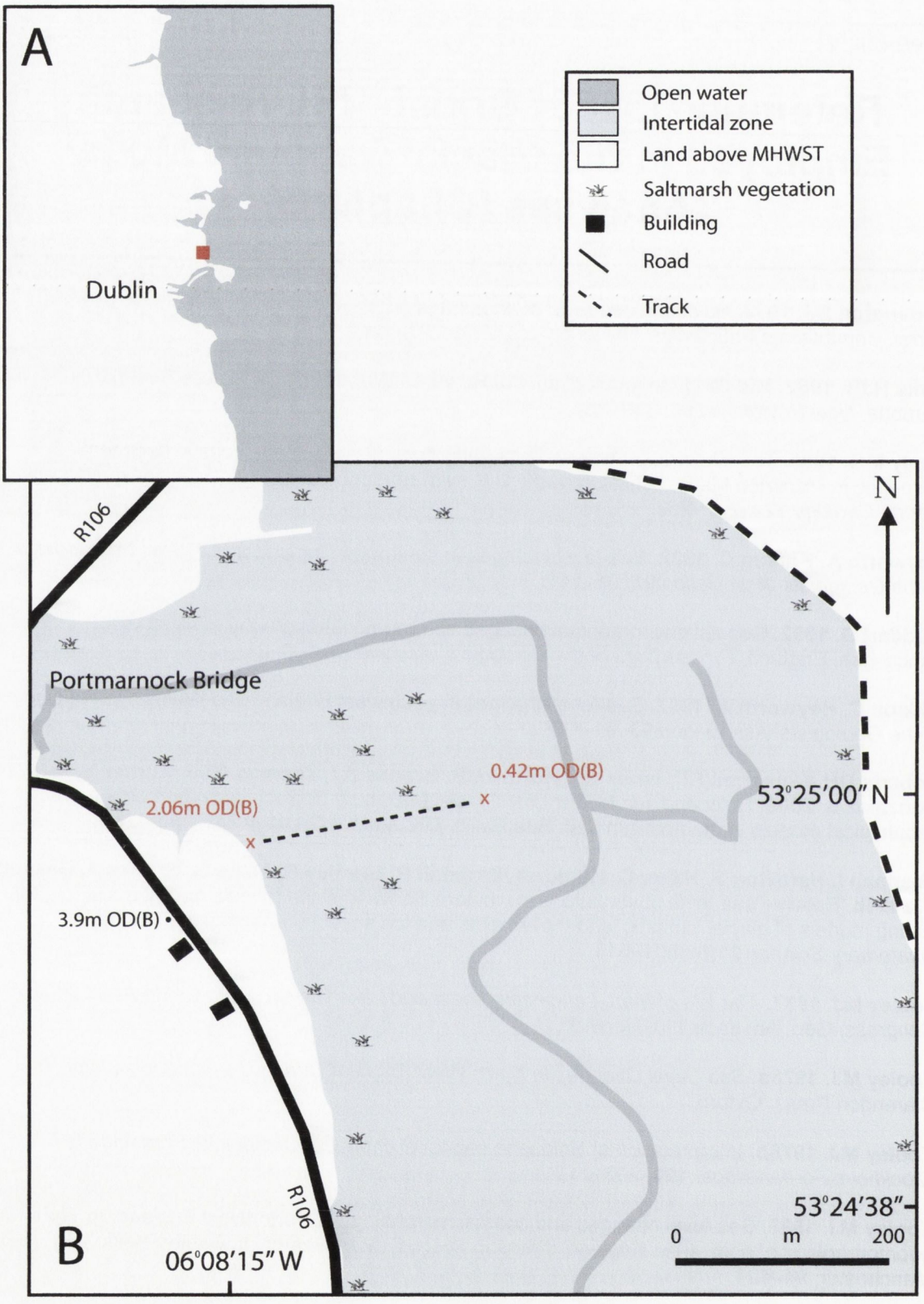


Figure A7.1 Site map showing **A** the location of Portmarnock salt-marsh in a regional context; **B** the location of the surface foraminiferal transect at Portmarnock

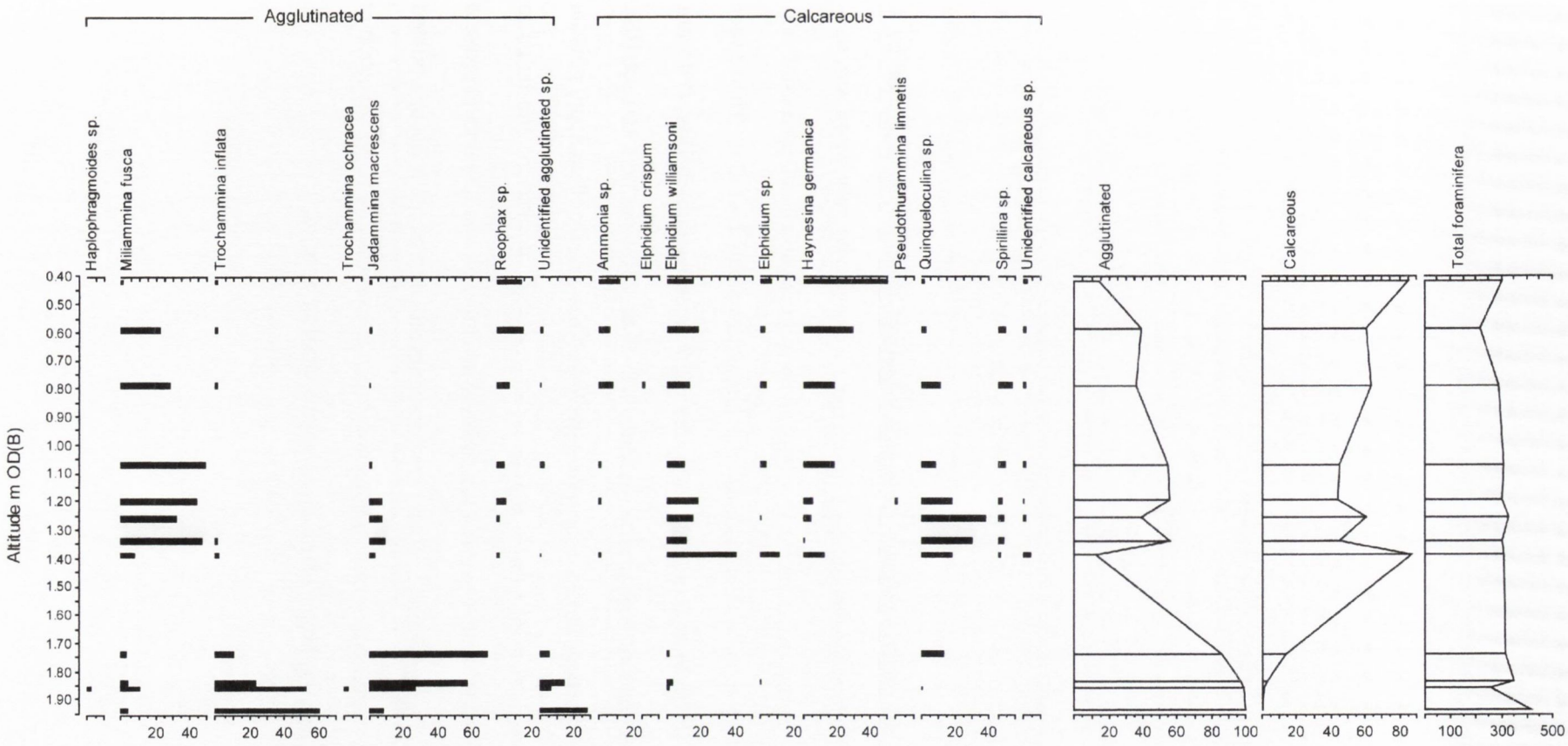


Figure A7.2 Foraminiferal death assemblages from the salt-marsh at Portmarnock. Only samples with counts greater than 40 individuals and species which reach 5% of the total sum are included

THE DEVELOPMENT OF A SEA-LEVEL DATABASE FOR IRELAND

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ABSTRACT

Despite a corpus of qualitative relative sea-level (RSL) data from Ireland, there exists a comparative paucity of precise and reliable information concerning changes since the Last Glacial Maximum (LGM). Much of the existing data are disparate in nature, having been produced by a range of studies employing different methodologies and often not specifically concerned with sea-level reconstruction. Consequently, there are large spatial and temporal gaps in our understanding of RSL changes, whilst other data are ambiguous and prone to misinterpretation. This paper presents a new database of sea-level information extracted from existing publications. Data are screened and classified into one of four groups according to their utility as indicators of RSL change. This objective classification, based upon well-established sea-level methodology, permits researchers to readily identify the most reliable trends in the available information, and helps to discriminate potentially erroneous data. The database is freely available on-line and should prove a valuable resource to researchers working in coastal environments around Ireland. It will be updated as new information becomes available, ensuring maximum value is extracted from the collection of new sea-level data and highlighting areas where more work is urgently needed.

INTRODUCTION

The delicate balance between post-glacial changes in land and ocean level has helped shape Ireland's coastline. The pattern of relative sea-level (RSL) changes produced by this interplay contains information concerning both the timing and magnitude of eustatic sea-level change, and the spatial variability associated with glacio-isostatic adjustment (GIA). Consequently, a firm understanding of RSL changes around Ireland underpins a variety of research, ranging from studies examining ice sheet history, Earth rheology, climate and sea level, to those concerned with coastal evolution, palaeogeography and archaeology.

The importance of understanding RSL changes around Ireland is reflected by a number of attempts to collate published sea-level data. Many of these catalogue data from geographically restricted areas, reflecting the particular focus of the research in question (e.g. Carter 1982; Penney 1983; Sinnott 1999). A notable exception to this is the paper by Carter *et al.* (1989) which presents a series of age-altitude plots from around the Irish coast. This remains the most comprehensive published record of RSL change in Ireland and now requires updating to reflect the results produced by studies during the last seventeen years. This need is highlighted by researchers seeking to further our understanding of British and Irish ice sheet history and associated GIA patterns (e.g. Peltier *et al.* 2002; Shennan *et al.* 2002, 2006).

This paper presents a new database which provides an updated catalogue of published Irish sea-level data. Data are screened and classified into four groups on the basis of their utility as indicators of RSL change. The rationale behind this classification is explained with reference to internationally recognised sea-level research methodologies. After discussing the limitations inherent with these data, the current temporal and spatial distributions of published information are presented, before concluding with some examples of updated RSL curves.

THE NATURE OF SEA-LEVEL DATA

Over thirty years of international research, co-ordinated under the auspices of a series of International Geoscience Programme (IGCP) projects, has produced a well-defined methodology for developing records of relative sea-level change from sedimentary coasts (Edwards 2005).

Central to this is the use of sea-level index points which fix the past altitude of sea level in time and space (Tooley 1978; Preuss 1979; van de Plassche 1986). Detailed consideration of sea-level index points and their associated error terms are given in a series of publications (e.g. Kidson and Heyworth 1979; Devoy 1982; Heyworth and Kidson 1982; Shennan 1982; 1986). In brief, for a sample to be established as a sea-level index point, it must possess information regarding its location (latitude and longitude), its altitude (relative to a levelling datum), its age (commonly inferred from radiocarbon dating), and its vertical relationship to a contemporaneous tide level (termed the indicative meaning). The latter is important when accounting for the differing vertical distributions of coastal sub-environments and associated sea-level indicators. Sea-level index points are commonly derived from lithostratigraphic contacts between terrestrial and marine sediments, with supporting microfossil data (e.g. foraminifera, diatoms) being used to delimit the onset or removal of brackish/marine conditions.

Where a dated sample does not possess an indicative meaning (e.g. a freshwater peat deposit) it cannot be used to reconstruct the former altitude of RSL. However, given knowledge of its depositional environment it can be used as a limit of possible RSL, and is thus referred to as a limiting date. For example, in the case of a freshwater peat deposit, the sample must have formed above the upper limit of marine influence indicating that, at the time of accumulation, local RSL must have resided at a level somewhere below the sample altitude.

Sea-level index points are routinely plotted as points on age–altitude diagrams with associated vertical and temporal error terms. These points fix the former altitude of RSL in time, but do not provide any information on the nature of sea-level change between points. The resolution at which past RSL changes can be reconstructed is therefore a function of the number and distribution of data points and the relative magnitudes of their associated error terms. Since RSL records are influenced by vertical land movements, it is important that sea-level data are only combined from geographically restricted areas in order to avoid the influence of differential crustal movements (Tooley 1978). In turn, the extraction of coherent sets of sea-level data can provide information on differential crustal movements between regions that may be used to model the GIA process (e.g. Shennan 1989; Shennan and Horton 2002).

A SEA-LEVEL DATABASE FOR IRELAND

The database described here presents sea-level data with information spanning over 20 fields, covering fundamental variables such as location, altitude and age, coupled with important supporting data such as tidal information, type of dated material and the indicative meaning of the sample (see *Primary index points* for details). These data have been extracted from published literature and a full set of source references is given in Appendix 2. The database takes the form of a Microsoft Excel™ spreadsheet to maximise accessibility and compatibility, and is available from the authors or may be downloaded directly from the School of Natural Sciences, Trinity College Dublin (http://www.naturalscience.tcd.ie/SL_Database.php).

Data classes

The contributing sea-level data from around the Irish coast are disparate in nature, being derived as they are from material produced by a range of authors employing differing methodologies and often not specifically targeting sea-level reconstruction. As a consequence the database discriminates four classes of sea-level data on the basis of the range of fields for which information is available, and the associated precision of these data. These four classes represent the different quality of existing sea-level data and reflect the extent to which the samples meet the stipulated requirements of a sea level index point.

Primary index points. These represent the highest quality sea-level data currently available from Ireland, and possess information on location, age, altitude and indicative meaning. Importantly, the relationship between the dated sea-level indicator and the environment in which it formed is quantified and error estimates for age and altitude are included. At the time of writing, there were approximately 50 published primary index points, and these were largely derived from relatively recent studies in the south of Ireland (Sinnott 1999; Devoy *et al.,* in press).

Secondary index points. These offer lower-quality information on the former position of sea level than that offered by primary index points. They are derived from dated sea-level indicators, but one or more of the core variables is unquantified or associated with significant uncertainty. This may include: an absence of accurate levelling data; poor chronological control; limited accompanying microfossil analysis (ambiguous environment); an unclear relationship between the dated material and sea level; association with an erosive contact or stratigraphic unconformity. In addition, sea-

level data sourced from archaeological evidence are included in this category since their indicative meanings are often poorly constrained (Long and Roberts 1997).

As a consequence of the uncertainties associated with these data, only tentative inferences on the position of former sea level can be made on the basis of secondary index points alone.

Limiting dates (Type I). These are derived from samples of known age and associated with a known environment that has no quantifiable relationship with sea level. As a consequence, it is only possible to infer whether sea level was above or below a certain altitude at a given time. These limiting dates are derived from material such as freshwater peat and *in situ* tree stumps (e.g. Carter 1982).

Limiting dates (Type II). These differ from Type I dates in that they are derived from material whose source environment is unclear or contested. Once again, these data only indicate whether sea level occupied a position above or below a certain altitude. However, the (possible) allochthonous nature of the dated material advises against attaching great significance to the age of the deposit, a point emphasized by numerous sea-level researchers (e.g. Kidson 1982). Instead, the date is seen as a 'maximum age' for the deposit since the dated material may have been deposited then reworked several times before finally occupying its current position in the stratigraphic column.

An example of Type II data are radiocarbon dates derived from wood that is not demonstrably *in situ* (e.g. recumbent trunks in estuarine silt). In addition, a number of accelerator mass spectrometry dates derived from foraminifera contained with 'glaciomarine muds' are also included since, whilst potentially important sources of sea-level data, their origins are the source of much controversy (McCarroll 2001).

Database fields

The database fields are closely related to the core attributes of a sea-level index point outlined above. Given the variable quality of the different data classes, not all entries possess data for the entire set of fields.

Geographical location. This information is provided in the form of site name, grid reference and latitude/longitude in decimal degrees. All primary index points have location information that is accurate to within 1km. For the purposes of displaying information, data are also assigned to a broad geographical area, where each area exhibits largely homogeneous RSL histories. At the time of writing, the database was sub-divided into 21 areas (Fig. 4.3), although this may change as more data become available.

Age. Most entries in the database have age information provided by radiocarbon dating. The database contains the sample laboratory code, radiocarbon age and calibrated calendar date calculated using CALIB 5.0.1 (Stuiver *et al.* 2005). The following options were chosen: a laboratory multiplier effect of 1; 95% confidence limits and their dataset IntCal04 (which is confined to 0–26ka cal BP). This dataset is recommended for most non-marine samples and is based on dendrochronologically dated tree-ring samples which cover the period from 0–12.4ka cal BP. For the time interval 12.4–26ka cal BP, data from marine records (corals and foraminifera) are converted to the atmospheric equivalent with a site-specific marine reservoir correction to provide terrestrial calibration.

In instances where marine samples (such as shells and foraminifera) have been dated, the dataset Marine04 was employed. The marine calibration dataset incorporates a time-dependent global ocean reservoir correction of about 400 years but to accommodate local effects, the difference Delta R in reservoir age of the local region of interest and the model ocean should ideally be determined (Stuiver and Reimer 2004). Where possible, this was achieved through reference to Harkness (1983).

Whilst some of the radiocarbon dates contained within the database are corroborated via pollen chonostratigraphic data, the vast majority have no such supporting evidence. Consequently, erroneous age estimates due to contamination may remain undetected. It is also the case that if erosion has occurred between a terrestrial–marine contact in the stratigraphic column, any age estimate for the date of transition between the two environments may overestimate the true age of the event. To guard against this, detailed microfossil analysis can be employed to demonstrate a continuous transition across the stratigraphic contact. This has been carried out for all samples contained within the primary index point tier. It should be noted that for the case of limiting dates,

an erosive contact is not important since no attempt is being made to fix the time of a transition between a marine and terrestrial environment.

Altitude. More than one levelling datum has been employed in research around the coastline of Ireland. In the database, all sample altitudes refer to Ordnance Datum (Belfast). Conversions were as follows: Irish Ordnance Datum (Poolbeg) lies 2.7m below Ordnance Datum (Belfast) which in turn lies approximately 0.03m below Ordnance Datum (Malin Head) (Admiralty Tide Tables 1997).

Indicative meaning. The indicative meaning of a sample describes its vertical position relative to the tidal frame at the time of its formation. Where available, the database provides information on the inferred indicative meaning, the nature of the evidence from which this is derived, and the local tidal parameters used in reconstruction.

TYPE OF CONTACT	REFERENCE WATER LEVEL	INDICATIVE RANGE
1: Base of basal peat	MTL to MHWST	?80cm
	MHWST+20cm	60cm
2: Fenwood/carr below marine clastics		40cm
+	MHWST-20cm	
3: Reedswamp below marine clastics		20cm
4: Carr above marine clastics	MHWST+80cm	60cm
-above Phragmites or saltmarsh	M^1-10cm to M^1-20cm	20cm
	MHWST	20cm
5: Reedswamp above marine clastics	MHWST+60cm	40cm
	M^1-10 to M^1-20cm	≥20cm
- above saltmarsh	M^1-20cm	20cm
6: Intercalated peat	M^1-10 to MHWST-20cm	70cm
7: Organic clay silt (salt marsh) below marine clastics	MHWST-20cm	40cm
8: Within basal peat (middle of layer)	MHWST to MTL	>80cm
	infer from stratigraphy	70cm
9: Saltmarsh peat above organic silt	MHWST	50cm
10: Fish trap (Archaeological evidence)	MTL	50cm

Table 1 Reference water levels (and associated indicative ranges) of sea level indicators used in the database. Adapted from Shennan, (1986). (M^1 = (HAT+MHWST)/2)

Indicative meanings comprise the portion of the tidal frame to which a sea-level indicator is related (termed the reference water level), and the modern vertical range over which the sea-level indicator is found (termed the indicative range). Table 1 provides a list of commonly employed sea-level indicators with their associated indicative meanings, adapted from Shennan (1986). For primary index points, all indicative meanings are derived from a combination of lithostratigraphic and biostratigraphic data. For other data classes, the formation of freshwater peat is conservatively assigned to elevations above mean high water of spring tides (MHWST), whilst mean high water of neap tides (MHWNT) is used as the boundary between the formation of organic-rich intertidal deposits (e.g. salt-marsh) from inter-tidal to sub-tidal minerogenic sediments containing shells.

Sea-level tendency. Established sea-level literature often makes reference to sea-level tendencies (Shennan 1980, 1982, 1983; Tooley 1982). A positive sea-level tendency is assigned to a sea-level indicator where it indicates an increase in marine influence (and vice versa for a negative sea-level tendency) irrespective of its cause. It is important to recognise that sea-level tendencies are not synonymous with rises or falls in RSL and changes in tendency may occur in the absence of sea level movement. The database provides tendency information and its mode of establishment, although in general, there are currently insufficient data points to perform meaningful tendency analysis (Shennan *et al.* 1983).

Data limitations

Users of the sea-level database must be aware of some fundamental limitations in the data it contains. Firstly, the database is compiled from published material and assumes that the information included within these publications is accurate and reliable. Full bibliographic details are provided in Appendix 2, and the user is advised to refer to the original publications for further details of any data they select to employ.

More specific limitations are associated with the differing classes and fields of data, reflecting wider limitations associated with existing sea-level methodology. In many cases, the influence of these additional variables cannot be precisely quantified and, as a result, the uncertainties associated with age or altitude may be larger than those indicated by the error terms. Below, the key points to be aware of are highlighted, along with reference to publications which deal with these issues in more detail.

Location, data distribution and regions. The quality of a local record of relative sea-level change is dependent upon the accuracy, precision and distribution (in time and space) of sea-level data. This is highly variable around the coast of Ireland with the result that sea-level changes in certain parts of the country are still associated with considerable uncertainty (Fig. 4.5).

The combination of sea-level index points from different sites into 'regional curves' assumes that inter-site differences in RSL change are not substantial. This assumption becomes increasingly prone to error as the areas become larger, or the individual records are analysed at higher resolutions. It will also be more suspect in areas with complex hydrographic regimes and coastal geometries since significant inter-site variability in tidal parameters (and hence indicative meanings) may be produced. The current 21 geographical areas reflect broadly similar modelled RSL histories derived from recent attempts to model the GIA process (Brooks *et al.*, submitted; (appendix 8). These regions are therefore suitable for input to geophysical models with similar, comparatively coarse, spatial resolutions. Conversely, care should be taken in extrapolating the general results from these broad regions to detailed, site specific studies.

Inferred age. The reliable interpretation of age data rests upon the assumption that the estimated age is correct, and the dated sea-level indicator reflects the timing of the sea level change of interest. This may not be the case if the sample has been contaminated by older or younger carbon, such as by rootlet penetration or inwashed material. In particular, it is important that the stratigraphic contact between terrestrial and marine sediments is not erosional or associated with a hiatus, since this will result in the overestimation of the change in marine influence. Recently, Waller *et al.* (2006) have questioned the reliability of SLIs based upon (apparently continuous) transgressive contacts. Their findings suggest that radiocarbon dates from the upper surface of peat layers should in most instances only be regarded as limiting ages for the deposition of the overlying clastic sediments, since peat growth often appears to slow down or cease well in advance of marine inundation. Other issues concerning the age of sea-level index points and dating material for sea-level reconstructions are given in van de Plassche (1986) and Edwards (2004).

Altitude. Errors in altitude can be introduced during sample collection as a consequence of limitations in the precision of the surveying equipment used, uncertainties inherent within the benchmark network, and errors associated with the measurement of lithostratigraphic contacts in the field (Shennan 1982, 1986). An additional potential source of error comes from uncertainties associated with establishing the indicative meaning of the sample (see below).

In the database, a standardised vertical error term after Shennan (1982, 1986) is derived from the following equation:

$$\sqrt{(a^2 + b^2 + c^2)} \quad \textbf{(Eq. 1)}$$

(a = Indicative range, b = error from boundary ID, c = depth measurement error)

Where a published estimate of total altitudinal error exceeds that calculated from Eq. 1, the former is adopted. Some entries in the lower quality tiers lack precise levelling data, having been related, for example, to 'around MHWST'. In these instances, a standard 'indicative' error term of ±1m is assigned. In the case of dated samples where the indicative meaning is ambiguous (e.g. fish traps) or where the environment of deposition is contested (limiting dates series II), no attempt has been made to allocate a vertical error term. Consequently, the absence of an error term indicates an error of unquantified magnitude rather than the absence of any vertical error.

In addition to the altitude errors described above, a further unquantified error may be introduced by post-depositional compaction of the sediment column, either under its own weight or as a consequence of subsequent loading by water or an overlying sediment burden. Whilst the potential influence of compaction on Holocene RSL records is long established and widely acknowledged (e.g. Jelgersma 1961; Terzaghi and Peck 1967; Greensmith and Tucker 1971a, 1971b, 1973; Tooley 1978; Heyworth and Kidson 1982; Shennan 1986; Allen 1996; Haslett *et al.* 1998; Shennan *et al.* 2000b; Edwards 2006), it has commonly been set aside due to the lack of a formal means for correcting for its influence (Allen 2000; Shennan and Horton 2002). Compaction serves to lower the reconstructed altitude of former RSLs and will be greatest where index points are established

from thick intercalated sedimentary sequences. Index points should therefore be interpreted in light of the stratigraphic data supplied in the database.

Indicative meaning. Each primary index point contained within the database has an associated indicative meaning which has been constructed with reference to contemporary tidal characteristics. Since the majority of data points in the database are collected some distance from an established tide gauge, site-specific tidal parameters have to be extrapolated or interpolated from the closest available records. This will introduce an error term which may only be accurately corrected by on-site data-logging of tidal conditions.

The indicative meaning method inherently assumes that significant changes in tidal range have not occurred at the study site. In reality, many regions will have experienced alterations in tidal parameters which will introduce error into reconstructions (e.g. Austin 1991; Shennan *et al.* 2000b). These errors are likely to be greatest in macro-tidal regions such as the Shannon Estuary and Dundalk Bay, or in areas that have experienced large changes in coastal geometry over time.

Area	**Number of data points**			
	Primary index points	*Secondary index points*	*Limiting dates type I*	*Limiting dates type II*
(1) Dublin		1	3	3
(2) N. Wexford	1			
(3) S. Wexford	11	1	3	1
(*) Celtic Sea				2
(4) E. Cork	10	1		
(5) W. Cork	13	1	3	1
(6) Kerry	1		10	
(7) S. Clare		1	6	
(8) Mid. Shannon		3		
(9) Inner Shannon			1	8
(10) Galway			1	
(11) Connemara			17	
(12) N. Mayo				8
(13) Sligo			1	
(14) W. Donegal	6		10	1
(15) N. Donegal	3		3	3
(16) L. Swilly			1	7
(17) Derry			4	6
(18) N. Antrim			5	4
(19) N. Down			10	5
(20) S. Down		2	4	4
(21) Dundalk			4	12
	45	**10**	**86**	**65**

Table 2 frequency distribution of Irish RSL data in each geographical region (as of 2/08/2006)

SEA-LEVEL CHANGES AROUND IRELAND: AN UPDATE

Data distribution

The spatial distribution of sea-level data from each data class is presented and summarised in Figure 4.5 and Table 2 respectively. A clear spatial bias in the quality and quantity of data is apparent. The highest quality sea-level data (primary index points) are concentrated along the southern Irish coast. This is in marked contrast to much of the eastern and western coastlines which have fewer reliable sea-level index points. This regional bias primarily reflects the availability of sedimentary sequences suitable for establishing sea-level index points. In comparison with the UK, surprisingly few intercalated sequences of terrestrial and marine sediments have been reported from the east and west coast of Ireland, and many of these are localised and of limited thickness (e.g. back-barrier deposits). Similarly, unlike a number of studies from Scotland (Shennan *et al.* 1993, 1995a, b, 2000a), there is a lack of isolation basin data from the north of Ireland. It is likely that this also reflects the relative scarcity of suitable sites, since modelling results suggest potential sites will be restricted to specific stretches of the Ulster coastline (Lambeck 1996; Lambeck and Purcell 2001).

Figure 4.6 shows the temporal distribution of different classes of sea-level data in 1000 year bins. From this figure it is clear that, in addition to the spatial variability outlined above, there is also a marked bias in the ages of available data with most of the sea-level index points dating to the mid or late Holocene. This reflects the fact that southern Ireland has experienced rising relative sea-level over the past 10 000 years and the rapid rate of early Holocene eustatic sea-level rise means early Holocene coastal sequences will now be found at a considerable depth or in offshore contexts.

Example data and their implications

Figure 4.7 shows example age–altitude plots from four regions distributed around the coast of Ireland. These plots show the variability in data quality in space and time and reveal the general patterns of relative sea-level change for these regions. As noted by previous workers, a generally coherent and consistent pattern of change is revealed by these data, showing the contrast between long-term RSL rise in southern Ireland with the more complicated, non-monotonic changes in north-east Ireland (e.g. Carter 1982; Carter *et al.* 1989). However, inspection of the classes of data

used reveal that, for many areas, the pattern of RSL change is only broadly delimited by limiting dates, and does not fix the altitude of past sea level in time and space. This is particularly the case for the RSL curves for the north of Ireland compiled by Carter (1982), which have frequently been referred to in papers considering aspects of coastal palaeo-environmental change (e.g. Lambeck 1996). Since these hypothesised 'sea-level curves' are primarily constructed from limiting dates, the magnitude and timing of specific features (e.g. the Holocene high stand or early Holocene low stand) cannot be accurately constrained and are potentially erroneous. Indeed, this is highlighted by the recent work of Orford *et al.* (2003) who consider the evolution of prograding beach ridges in Dundrum Bay (Co. Down). Their data indicated that the regressive phase of the Holocene highstand is likely to have lasted longer than Carter envisaged (Carter 1982).

CONCLUSIONS

The database presented here is a first attempt at compiling a standardized sea-level database for Ireland, which is consistent with established protocols for the evaluation of sea-level information. The available data are of varied quality, yet limitations associated with this are mitigated by applying a rigorous appraisal of data errors and the consequent adoption of a quality tier system. The database is now available at http://www.naturalscience.tcd.ie/SL_Database.php and should represent a valuable tool to a range of user groups including coastal scientists, archaeologists and geophysicists interested in understanding patterns of GIA in Ireland.

However, users of the database should be aware of the limitations (in precision and accuracy) of the data points it contains, and should interpret these in light of established sea-level methodology. For example, of the 200 records in the database, less than 25% occupy the primary SLI tier and, as a consequence, a large proportion of the existing data must be considered as 'limiting' in nature. In the past, attempts have been made to treat these limiting data as index points, which in some instances has resulted in misleading conclusions being reached. Arguably the biggest weakness in the dataset at present is the spatial and temporal bias of the index points. The highest quality data are almost exclusively found along the south coast and are of mid/late Holocene age. Forthcoming data from Delaney *et al.* (in press) (Co. Kerry), Edwards and Brooks (The Shannon) and Harmon (Co. Donegal and Co. Down) will go someway to rectifying this imbalance. However, further targeted field research is urgently required, and this will need to utilise alternative methodologies in order to investigate a wider range of environments (e.g. sub-tidal or shallow marine contexts).

Several attempts have been made to model RSL change and GIA around the Irish coast and elsewhere (e.g. Lambeck 1996; Lambeck and Purcell 2001). However, some authors have argued for caution in the adoption of these models, highlighting contradictions with field data (e.g. McCabe 1997; Smith 2005). It is, therefore, vital that the collection of primary data continues in association with ongoing model development. It is hoped that this database will both stimulate and facilitate this process of exploration, testing and revision.

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POST-GLACIAL RELATIVE SEA-LEVEL OBSERVATIONS FROM IRELAND AND THEIR ROLE IN GLACIAL REBOUND MODELLING

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ABSTRACT

The British Isles have been the focus of a number of recent modelling studies owing to the existence of a high quality sea-level data set for this region and the suitability of these data for constraining shallow Earth viscosity structure, local to regional ice sheet histories and the magnitude/timing of global melt water signals. Until recently, the paucity of both glaciological and relative sea-level (RSL) data from Ireland has meant that the majority of these glacial isostatic adjustment (GIA) modelling studies of the British Isles region have tended to concentrate on reconstructing ice cover over Britain. However, the recent development of a sea-level database for Ireland along with emergence of new glaciological data on the spatial extent, thickness and deglacial chronology of the Irish ice sheet means it is now possible to revisit this region of the British Isles. Here, we employ these new data to constrain the evolution of the Irish Ice Sheet. We find that in order to reconcile differences between model predictions and RSL evidence, a thick, spatially extensive ice sheet of c. 600-700m over much of north and central Ireland is required at the LGM with very rapid deglaciation after 21 000 cal. BP.

1. INTRODUCTION

Recently, a number of researchers have modelled observations of glacio-isostatic adjustment (GIA) from the British Isles (e.g. Lambeck, 1991; Lambeck, 1993a; Lambeck, 1993b; Lambeck, 1995; Lambeck *et al.*, 1996; Peltier *et al.*, 2002; Shennan *et al.*, 2002; Shennan *et al.*, 2000c; Shennan *et al.*, 2000d; Shennan *et al.*, 2006). These efforts have been facilitated by the availability of long time-series records of relative sea-level (RSL) change in Britain which, through the process of geophysical modelling, can be distilled to infer information regarding former ice extent, thickness and melting history, as well as viscosity structure within the mantle. However, in the past, the fragmentary nature of the Irish RSL record, allied with a paucity of observational constraints on the Irish Ice Sheet, has deterred researchers from using these data to constrain either the ice history or sub-surface viscosity structure for this region. Notable exceptions to this are the works by Lambeck (1996) and Lambeck & Purcell (2001), which utilize existing records of RSL to constrain the vertical limits of the Irish ice sheet.

Since these publications, there have been a number of pertinent developments that mean it is now timely to re-evaluate the contribution Irish observational data can make to modelling GIA in the British Isles. Firstly, the body of available Irish sea-level data has grown since the work of Lambeck and co-workers. This is augmented by new, high resolution RSL records dating back to c. 16 000 BP, from sites immediately adjacent to Ireland (Shennan *et al.*, 2006), which provide a powerful and unique constraint on model reconstructions of the Irish Ice Sheet.

Secondly, Brooks & Edwards (2006) have developed a sea-level database for Ireland in which observations of RSL have been screened and classified according to established protocol for the analysis of sea-level data (e.g. Kidson & Heyworth, 1979; Heyworth & Kidson, 1982; Shennan, 1982; 1986; Shennan, 1989; Shennan & Horton, 2002). This objective classification permits users of the database to readily identify the reliability of the available RSL information, and helps to discriminate potentially erroneous data. A similar database for the UK (Shennan & Horton, 2002), has been integral to previous GIA modelling studies from Britain (e.g. Lambeck, 1995; Shennan *et al.*, 2002; Shennan *et al.*, 2006).

Thirdly, new data on the spatial extent (e.g. O'Cofaigh & Evans, 2001; Hiemstra *et al.*, 2006; Serjup *et al.*, 2005) and thickness (e.g. Rae *et al.*, 2004; Bowen *et al.*, 2002) of the Irish ice sheet since

the last glacial maximum (LGM) are now available. Furthermore, a number of Accelerator Mass Spectrometry (AMS) radiocarbon dates derived from marine muds have, for the first time, delivered chronological control on the post-LGM deglacial phase (e.g. McCabe & Haynes, 1996; McCabe & Clark, 1998, 2003; McCabe *et al.*, 2005). In addition, field evidence now indicates that ice from the British Isles coalesced with Fennoscandian ice during the late Devensian (e.g. Serjup *et al.*, 1994, 2005; Svendsen *et al.*, 2004; Carr *et al.*, 2006), producing a more laterally extensive ice sheet than envisaged by Lambeck (1996).

Finally, Shennan *et al.*, (2006) have recently developed a revised GIA model for the British Isles (hereafter referred to as BIM-1), which satisfies key glaciological, seismic and sea-level constraints from both near- and far-field localities. This new model incorporates a more realistic assessment of ice sheet mass (and associated crustal loading) than previous studies by accounting for the effects of topography/bathymetry on ice thickness (Milne *et al.*, 2006). Importantly, whilst this model has an Irish ice sheet component, it is not calibrated using any of the Irish RSL data. Consequently, the new dataset of Irish RSL reconstructions can provide a powerful, independent test of its earth and ice components.

In this paper, we evaluate the RSL predictions of BIM-1 against the independent Irish dataset. We demonstrate that significant misfits exist between observations and predictions for certain portions of the Irish coastline, indicating limitations in one or more components of BIM-1. We investigate the source of these limitations by examining the effects of modifying the Irish ice sheet and earth parameters of this model. In light of these results, we produce a revised model of the Irish ice sheet that satisfies both RSL and glaciological field constraints. Our results also indicate that minor modifications of the earth model are required to reconcile RSL predictions with observations from Ireland and the west coast of Britain.

2. MODELLING RELATIVE SEA-LEVEL CHANGE

The use of geophysical models to simulate RSL change associated with the growth and ablation of mid-to-high latitude ice sheets is well established (e.g., Walcott, 1972; Peltier, 1974; Farrell & Clark, 1976; Peltier & Andrews, 1976; Nakada & Lambeck, 1987; Mitrovica & Peltier, 1991; Johnston, 1993; Milne & Mitrovica, 1996; Milne *et al.*, 1999). In essence, models of GIA consist of three key elements: an ice loading model (to define the global distribution of grounded ice

thickness over time), an earth model (to simulate deformation of the solid earth to surface loading) and an algorithm to compute associated changes in sea level (which, as well as being a key observable, is also an important component of the GIA loading model).

Some authors have expressed concern about the ability of geophysical models to accurately reproduce patterns of RSL change (e.g. McCabe *et al.*, 2005). It has been argued that models of this type are over-reliant on Holocene sea-level observations to constrain predictions of RSL change during the earlier deglacial phase (e.g. McCabe, 1997), whilst others have suggested that such models are unable to capture detailed variations in RSL change during the Holocene period (e.g. Smith, 2005). However, such criticisms are often based upon controvertible indicators of former sea-level which as yet cannot provide unequivocal data with which to test these suggestions.

This study tests and modifies the BIM-1 GIA model employed in Shennan *et al.*, (2006), the key components of which are outlined below.

2.1 Earth Model

The earth model employed in this investigation is a spherical, self-gravitating, compressible, Maxwell visco-elastic body. The elastic and density structure are taken from seismic constraints (Dziewonski & Anderson 1981) and are depth-parameterised with a resolution of between 10 and 25 km. The viscosity structure is more crudely parameterised into three layers which correspond to the lithosphere (in which the viscosity is set to a very high value of 1×10^{43} Pa s), the upper mantle (from the base of the lithosphere to the 670 km seismic discontinuity) and the lower mantle (from 670 km depth to the core-mantle boundary). In this study, we employ the same viscosity parameters that have resolved discrepancies between far-field RSL data and model predictions (Bassett *et al.*, 2005) and delivered a good fit to the observational evidence of post-glacial RSL change in Britain (Shennan *et al.*, 2006). These values are 1) a lithospheric thickness of 71km, 2) an upper mantle viscosity (v_{UM}) of 5×10^{20} Pa s, 3) a lower mantle viscosity (v_{LM}) of 4×10^{22} Pa s.

2.2 Sea-Level Model

To generate sea-level predictions based on an input earth and ice model, we solve the most recent, generalised, form of the sea-level equation (Mitrovica & Milne 2003; Kendall *et al.*, 2005). By applying this most recent version of the theory, the influence of shoreline migration and sea-level changes in areas of ablating marine-based ice are accounted for. The theory we apply also takes into account the influence of GIA-induced perturbations in earth rotation on sea-level change (e.g. Milne & Mitrovica 1996). All predictions were generated using a spherical harmonic truncation of degree and order 256.

2.3 Ice Model

The ice model we adopt is comprised of a relatively low resolution global model with a higher resolution model for the BIIS. The global component of the ice model is the same as that used by Shennan *et al.*, (2006) and follows the analyses of Bassett *et al.*, (2005) which gives a close fit with far-field observations of RSL dating from the time of the LGM to c. 9000 BP. The Holocene component of this model remains to be calibrated and so in the following analyses we adopt the findings of Nakada & Lambeck, (1989) who suggest a late Holocene eustatic melt water contribution of c. 2m between 6000 and 2000 BP.

In order to accurately model GIA observations from Ireland and adjacent regions, a high-resolution spatial and temporal description of the local to regional ice load leading up to and following the LGM is required (Lambeck, 1993a). The space-time evolution of the Irish component of BIM-1 is illustrated in Figure 5.3. Overall advance and retreat phases are based on the findings of Sejrup *et al.*, (1994, 2005), with extensive ice accumulation signalling the end of the Alesund Interstadial at c.33 000 BP and leading to coalescence of the British and Fennoscandian ice sheets. This configuration is interrupted by marked ice sheet retreat at 26 000 and 25 000 BP before rapid expansion of the BIIS around the time of the LGM. Deglaciation begins at 21 000 BP, with the removal of ice from both Ireland and the UK largely complete by 15 000 BP (e.g. Ballantyne, 1997).

3. RECONSTRUCTING RELATIVE SEA-LEVEL CHANGE

Over thirty years of international research, coordinated under the auspices of a series of International Geoscience Programme (IGCP) projects, has produced a well-defined methodology for developing records of relative sea-level change from sedimentary coasts (Edwards, 2005).

Central to this is the use of sea level index points which fix the past altitude of sea level in time and space (Tooley, 1978; Preuss, 1979; Van de Plassche, 1986). Detailed consideration of sea level index points and their associated error terms are given in a series of publications (e.g. Kidson & Heyworth, 1979; Heyworth & Kidson, 1982; Shennan, 1982; 1986; Horton *et al.*, 2000). In brief, for a sample to be established as a sea level index point, it must possess information regarding its location (latitude and longitude), its altitude (relative to a levelling datum), its age (commonly inferred from radiocarbon dating), and its vertical relationship to a contemporaneous tide level (termed the indicative meaning).

Relative sea-level data from Ireland have recently been screened, classified and compiled into a database by Brooks & Edwards (2006) (Appendix Two). All index points have location information (latitude / longitude in decimal degrees) that is accurate to within 1 km. At present, lengthy suites of index points from single sites in Ireland are rare. Instead, SLIs are grouped into 21 distinct geographical regions to permit more meaningful comparison between RSL observations and predictions (see Figure 5.6). Most of the sea-level data have age information provided by radiocarbon dating, calibrated using CALIB 5.0.1 (Stuiver *et al.*, 2005). Some of these dates are also corroborated via pollen chonostratigraphic data. The altitude of all index points is defined relative to OD (Belfast), a mean sea-level datum.

The RSL database classifies samples primarily on the quality of their indicative meaning. All 'Primary' SLIs have a quantified vertical relationship to a former tide level, associated with clearly defined error terms. In situations where the indicative meaning of a sample is less well quantified, data have been grouped into a 'Secondary' tier of index points. Where a dated sample has no discernable relationship with sea-level, or has formed outside of the tidal frame, samples are categorized as 'limiting dates', and provide a maximum (or minimum) limit for sea level at a given time. 'Primary' limiting dates include dated samples which are associated with a known environment (e.g. freshwater peat), whilst 'Secondary' limiting dates are derived from material whose source environment is unclear or contested. A number of AMS radiocarbon dates taken from foraminifera contained within 'glaciomarine muds' are included in this latter category.

Both types of data are useful for testing RSL predictions produced by GIA models, since sea-level curves should pass through SLIs and remain within the constraints imposed by limiting dates. Post-

depositional compaction of the sediment column, either under its own weight or as a consequence of subsequent loading by water or an overlying sediment burden may lower the reconstructed altitude of former RSL. This will be greatest where index points are established from thick intercalated sedimentary sequences and consequently, when comparing predictions with observations from such sequences, best fit RSL predictions should lay towards the top of the SLI scatter (e.g. Shennan *et al.*, 2002, 2006).

4. TESTING THE BIM-1 GIA MODEL FOR THE BRITISH ISLES

The BIM-1 GIA model is used to produce estimated RSL curves for a series of locations around the Irish coast and adjacent areas in western Britain. These curves are plotted in Figure 5.7 alongside the geologically-based RSL reconstructions. The model predictions along Ireland's east coast (Dublin (1) and S. Wexford (2)), south coast (E. Cork (4)) and west coast (Connemara (11)), are in reasonable agreement with the observational dataset. This indicates that, away from the former centre of loading by the Irish ice sheet, BIM-1 reproduces the dominant patterns of change with some skill. Further to the north and west, however, significant discrepancies begin to emerge between the model predictions and the empirical RSL data.

The first of these misfits is the tendency for BIM-1 to underestimate the maximum height of late Devensian RSL along the north Ulster coastline, as indicated by abundant raised shoreline evidence in this region. Limitations associated with dating and quantifying the indicative meaning of raised shorelines, coupled with the lack of primary index points for this time period, prevents a precise quantification of this discrepancy. Nevertheless, these shorelines provide strong evidence of higher sea-levels, both to the west and south of Antrim, than indicated by the model (Stephens and Synge, 1965; Synge and Stephens, 1966; Synge, 1977a, b; Mitchell, 1977). For example, in Derry (17), a late Glacial RSL of up to +25 m (compared to a modelled estimate of c. +13m) is suggested by Stephens & Synge (1965), whilst in North Down (19), Stephens & McCabe (1977) find evidence of raised shorelines at heights of +30 m (compared to a modelled estimate of just over +10m).

The second misfit concerns the presence of a Holocene high-stand (commonly around 6000 BP), where ongoing crustal rebound starts to outpace dwindling sea level rise. The model delivers predictions of Holocene RSL higher than present as far south and west as Sligo (13), as indicated

by the 0 m isobase in figure 5.8. In west Donegal (14), BIM-1 predicts RSL of approximately +2 m at 6000 BP. Whilst there are very few locations in Ireland where the magnitude of the Holocene highstand can be tightly constrained (Brooks & Edwards, 2006), the work of Shaw (1985), Shaw & Carter (1994), Gehrels (2004) and Harmon (2006) in Donegal has enabled the spatial signature of the 0 m Holocene isobase to be reasonably well defined. The index point data provided by the authors above shows that RSL occupied a position around -2m at c.5000 BP in west Donegal (14) and suggests that the 0 m threshold lies somewhere along the north Donegal coastline (i.e. further to the north and east than suggested by BIM-1). On the Isle of Man (24), similar discrepancies exist with the Holocene RSL data. Here, the model predicts values in excess of +5 m whereas the data constrain sea level to have been no more that 2 m above present during this period.

5. EXAMINING THE POSSIBLE CAUSES OF POOR MODEL PERFORMANCE

The RSL data from Ireland demonstrate that BIM-1 performs poorly in areas that have experienced the greatest loading from the Irish ice sheet, but that it can predict RSL change with reasonable skill in those areas that have experienced thinner, or shorter-duration ice cover. In light of this, we perform a series of analyses to examine the Irish ice sheet component of BIM-1 and its influence on the pattern and magnitude of predicted RSL change.

5.1 The role of the Irish Ice Sheet

In order to predict the timing and magnitude of glacial rebound, it is necessary to know the size (thickness and lateral extent) of the ice sheet. In addition, it is important to understand when, how fast and by how much this size has changed through time. This choice of input parameters inevitably means that there is a degree of model solution non-uniqueness due to the trade-off between variables, and this is a recognised limitation inherent in GIA modelling studies (e.g. Mitrovica & Peltier, 1995). However, both the observational RSL data and glaciological field evidence (below) provide a stringent set of independent conditions that must be met, and any adjustment to the nominal BIM-1 ice sheet model are made in light of these constraints.

Ireland (along with Britain) is thought to have been ice free during the Derryvee Interstadial, in agreement with a radiocarbon date of c.34 000 BP derived from organic silts in northwest Ireland (Colhoun *et al.*, 1972). This warm episode was abruptly terminated by rapid ice sheet growth and expansion which lasted until c.27 000 BP. Before the onset of the LGM in this region at 24 000 BP,

a brief retreat phase of the BIIS ensued between 26 000 and 25 000 BP in agreement with the inferences of Serjup *et al.*, (1994).

During the LGM, a laterally extensive ice sheet is believed to have existed over much of Ireland and out onto the adjacent shelf region (e.g. Evans *et al.*, 1980; Warren, 1992; McCabe, 1995; Clark and Meehan, 2001; Serjup *et al.*, 2005). To the south, ice is now considered to have extended beyond the traditional South of Ireland End Moraine (SIEM) limit (O'Cofaigh & Evans, 2001; Evans & O'Cofaigh, 2003; Hegarty, 2004; Ballantyne *et al.*, in press). Recent evidence also indicates ice extending out into the Celtic Sea south of the Scilly Isles (Scourse & Furze, 2001; Hiemstra *et al.*, 2006). In terms of ice thickness, there is a relative paucity of constraints provided by trimline data within Ireland when compared to the UK (e.g. Ballantyne *et al.*, 1998a, b; McCarroll & Ballantyne, 2000). In the south west of Ireland (Macgillycuddy's Reeks), Rae *et al.*, (2004) use trimline evidence to infer an ice sheet thickness of c.370m whilst in the east (Wicklow Mountains), Ballantyne *et al.*, (in press) employ similar evidence to suggest that only peaks above c.725m remained ice free at the LGM. Elsewhere, the evidence is more equivocal: In the north of Ireland, ice is believed to have reached a thickness of at least 750m (after terrain correction) (e.g. Clark & Meehan, 2001; McCabe, 1997) although thicknesses are suggested to be somewhat less further south since the highest peaks of the Knockmealdown mountains (Lewis, 1976), Galtee mountains (McCabe, 1985) and Connemara mountains (Coudé, 1977) are proposed to have been nunataks at the LGM.

In BIM- 1, the onset of deglaciation is said to have began after 21 000 BP with (rapid) ice retreat taking place up until 19 000 BP. By this time, the Irish Midlands are suggested to have been ice free, and the ice margin in the Irish Sea occupied a position to the north of the Isle of Man. Interrupting this ice retreat from the LGM was a large-scale ice re-advance which affected much of the country (e.g. McCabe et al. 1998; McCabe & Clark, 2003; McCabe *et al.*, 2005). This 'Killard Point' or 'Heinrich 1' event involved ice re-advance as far south as the Shannon estuary and the north of the Isle of Man in the Irish Sea Basin (Thomas *et al.*, 2004). This event was brief however, and ice retreat followed shortly after, with the region becoming ice free by 15 000 BP (e.g. McCabe *et al.*, 2005). During the Younger-Dryas event (c.13 000 BP) ice was confined to mountain cirque glaciation (Murray-Gray & Coxon, 1991; Coxon, 1997) and was on a significantly smaller scale than its equivalent in the Scottish Highlands (e.g. Thorp, 1986).

The underestimation of maximal late Devensian RSL by BIM-1 indicates the need for greater crustal depression at this time. This requires either an earlier, or larger LGM ice build up, and/or a more persistent ice sheet during deglaciation (e.g. slower melting). In contrast, the overestimation of the Holocene high-stand around 6000 BP indicates that the model predicts too much glacial rebound by this time. This requires a reduction in LGM ice volume and/or persistence accompanied by more rapid melting during the deglacial phase. The latter therefore places clear constraints on the modifications acceptable to produce the required late Devensian RSLs.

5.1.1 Sensitivity analysis

In order to demonstrate the effects of varying ice volume at the time of the LGM and during the deglacial phase, we begin by presenting two sets of sensitivity analyses conducted using BIM-1 but with differing modifications to the Irish ice sheet component. In the past, researchers working outside the GIA modelling community have sometimes expressed concern at the accuracy of GIA ice model components at the local scale and the impact these inaccuracies may have on predictions of RSL (e.g. McCabe, 1997; McCabe *et al.*, 2005). For this reason, we first examine the influence of local-scale ice movements, such as the re-advance associated with the 'Killard Point' stadial included in the nominal ice model for BIM-1. We do this by considering the contrasting RSL changes that arise if Ireland were completely ice-free after 19 000 BP (Model A). We also consider a second model (Model B) in which the spatial extent of Irish ice to the west and north-west of Ireland at the time of the LGM is reduced, with (onshore) ice limits similar to those proposed by Stephens & Synge (1965), and Bowen *et al.*, (2002).

The results of these two experiments are plotted in Figure 5.7 alongside the original RSL predictions of BIM-1. It is clear from this that in the south and southwest of Ireland (away from the centre of crustal loading), the effects of removing all ice from Ireland after 19 000 BP (Model A) are small. However, in the north of Ireland the changes are more pronounced. The greatest divergence from the nominal BIM-1 ice model occur at c. 15 000 BP with a reduction in RSL predictions of c.10-12m. During the Holocene there is also an appreciable difference between BIM-1 and Model A predictions, with the latter commonly reducing the amplitude of the Holocene highstand by c. 2-3m. The greatest difference between BIM-1 predictions and the RSL predictions from Model B (removal of west coast ice at the LGM) occurs, perhaps unsurprisingly, at sites located in NW

Ireland. However, even in west Donegal (14), where the largest change in RSL is experienced, the effect of total offshore ice removal only contributes to c. 5 - 6m of RSL lowering at 20 000BP. At all sites, the influence on the Holocene GIA signature is not significant.

5:1:2 Improving the fit with empirical RSL evidence

The results of these sensitivity analyses clearly demonstrate the requirement for a very thin ice sheet during the deglacial phase, since significant reductions in the amplitude of the Holocene high stand (and the associated geographical location of the Holocene 0m isobase) cannot be generated by local-scale changes in LGM ice thickness alone (Model B). Whilst the results from Model A significantly improve the fit with Holocene RSL evidence, the position of the Holocene 0m isobase in Donegal is still too far to the west.

We perform a further set of analyses to examine the effects of changing the duration of ice loading, initiating deglaciation shortly after 22 000 BP (e.g. McCabe *et al.*, 2005) as opposed to after 21 000 BP in BIM-1. In the first analysis (Model C), we adopt the same pre-LGM and LGM ice sheet as BIM- 1 and, in accordance with the earlier sensitivity analyses, adopt a very thin Irish ice sheet of approximately half the ice thickness of BIM-1 during the deglacial phase. The second analysis (Model D) is identical to Model C except that, at the time of the LGM, we invoke a more laterally extensive ice sheet of similar morphology to that proposed by Serjup *et al.*, (2005). In this model, offshore ice thicknesses approach c. 650m off NW Donegal with ice extending out to the continental shelf break. Ice thickness in the northern Irish Sea Basin are also increased to a similar thickness in order to reconcile glaciological evidence suggesting that Snaeffel (621m) on the Isle of Man was transgressed by ice at the LGM (e.g. Bowen, 1973; Boulton *et al.*, 1977).

Figure 5.10 shows a comparison between RSL predictions generated by the nominal ice model for BIM-1 and the revised ice models C and D. Once again, away from the centre of ice loading over the north of Ireland, the disparity between RSL predictions from the three models becomes insignificant. However, around the north of Ireland coastline the differences become more apparent. Predictions from Model C deliver a far better fit with the Holocene observational RSL data than BIM- 1 and the Holocene 0m isobase no longer lies to the west of Sligo. A brief comparison between Model C and Model A (all Irish Ice removed from BIM -1 after 19 000 BP)

reveals that after c. 15 000 BP, predictions from the two models are largely identical with differences in the Holocene of no more than a few decimetres.

The RSL predictions produced by Model D (laterally extensive, thicker LGM ice) show even greater agreement with RSL data. The geographical extent of the Holocene high stand in the north is restricted, whilst concomitantly increasing maximal late Devensian RSL predictions to heights above those delivered by Model C. However, even this increase is insufficient to satisfy the geomorphological RSL evidence and limiting data (minimum) provided by dated 'glaciomarine' muds from around the north and eastern Ulster coastline. Neither do predictions from Model D yield a particular good fit with the high resolution sea-level index point data from Islay (22) and Knapdale (23) in southeast Scotland. In fact, in the north of Ireland, maximal late Devensian RSL predictions derived from this ice model are no higher than those produced by BIM- 1 since the increase in RSL predictions induced by thicker ice is offset by the effects of rapid and earlier deglaciation.

5.1.3 A best-fit Irish Ice Model

The analyses conducted in section 5:1:2 demonstrate that whilst changes to the ice sheet in the deglacial phase have greatly improved the fit with the Holocene observational evidence, significant misfits remain between model predictions and geomorphological RSL evidence from the late Devensian. Specifically, the maximal predictions of late Devensian RSL are consistently below the geomorphological evidence of RSL for this time period. Because of this, it would appear that early deglaciation shortly after 22 000 BP is incompatible with the Irish RSL data. In light of these results, we develop a revised Irish ice sheet component for BIM-1 that is constrained by both glaciological and RSL data (Model E, Figure 5.11). This model uses the same pre LGM ice build up, and LGM duration (24 000 – 21 000 BP) as the nominal Irish Ice component of BIM-1. Our best fit model differs from BIM-1 by having thicker ice in the Irish Sea and to the northwest of Ireland at the LGM (similar to Model D). We also have thickened ice across the Wicklow Mountains, Irish Midlands and over Kerry, in accordance with the trimline data of Rae *et al.*, (2004) and Ballantyne *et al.*, (in press). During the deglacial phase (20 000 to 16 000 BP), we adopt an almost identical ice sheet to that described in Model C, including rapid onset of deglaciation at 21 000 BP (figures 5.11 and 5.12).

Despite significant differences between the Irish ice sheet component of the nominal BIM-1 model and our revised 'best fit' ice model (Model E), there are no discernible differences between the RSL predictions of the two models for the majority of sites along Ireland's east, south and west coasts (see figure 5.13). The only exception to this is in Dublin (1), where the significantly greater volume of ice in the Irish Sea Basin associated with Model E results in more crustal uplift and higher predicted Holocene RSLs. Similarly, for most sites along the western coast of Britain, there is good agreement between RSL predictions from the two models.

Model E also produces a greatly improved fit with RSL data from the north of Ireland. The problematic Holocene high stand from Sligo (13) no longer exists and is greatly lowered in west Donegal (14). At all sites along the Ulster coastline, RSL predictions from Model E now yield a far better fit with the limiting data (maximum) in both the Holocene and back into the late Devensian. This is achieved whilst also generating sufficient crustal depression to satisfy most of the raised shoreline data and a lot of the limiting data (minimum) from the AMS radiocarbon dated 'glaciomarine' muds.

It is apparent that at both Islay (22) and Knapdale (23), the fit with the index point data is not as good as with BIM-1. This misfit is caused by our thin Irish ice sheet during the deglacial phase which has enabled increased crustal rebound to occur and a subsequent lowering of RSL. However, local scale adjustments in both Scottish ice thickness and more specifically the rate of ice retreat should provide a solution to these model misfits.

5.2 The Role of the Earth Model

The Earth model used in these analyses (and outlined in section 2:1) is parameterised into three layers which correspond to the lithosphere, the upper mantle and the lower mantle. When considering patterns of GIA in regions once occupied by small ice sheets such as the BIIS, changes in the depth/viscosity of the lithosphere and upper mantle are of greatest significance. Adjustments to the lower mantle would appear less important since even the rebound associated with the much larger Fennoscandian ice-sheet is only weakly sensitive to this model parameter (Mitrovica and Peltier, 1991; Peltier, 1996).

Lithospheric thickness controls the magnitude of crustal depression in response to loading and any changes to this will either uniformly raise or lower predictions of RSL throughout the entire post glacial period. It follows that changes in the depth of the lithosphere will be unable to reconcile differences between predictions from BIM-1 and the RSL evidence since maximal predictions from BIM- 1 are generally too low in the late Devensian and too high in the Holocene. However, upper mantle viscosity is of more relevance here since this parameter controls the rate of crustal response to loading-unloading episodes (e.g. Lambeck, 1993a,b). Estimates for upper mantle viscosity have been consistent between various modelling studies from the British Isles with most workers preferring either a value of 4 x 10^{20} Pa ^{14}Cs (Lambeck, 1993a; Lambeck, 1993b; Lambeck, 1995; Lambeck *et al.*, 1996; Shennan *et al.*, 2000c) or 5 x 10^{20} Pa s (Peltier, 1998; Peltier, 2004; Peltier *et al.*, 2002; Shennan *et al.*, 2006). Reducing the values of upper mantle viscosity will increase the response rate of the crust to loading and unloading episodes. The corollary of this will be a reduction in the rate of crustal deformation during the Holocene and consequently induce a lowering of RSL predictions during this period.

To examine the impact of changes in upper mantle viscosity, we perform a final set of sensitivity analyses on BIM-1 by producing RSL predictions generated with reduced upper mantle viscosity values of 3 x 10^{20} Pa s and 4 x 10^{20} Pa s. These are compared with the original RSL predictions from BIM-1 (v_{UM} 5 x 10^{20} Pa s) and the results presented in figure 5.15. In the south and west of Ireland, where the eustatic contribution to RSL change appears to dominate over the glacio-isostatic component, the reduction in upper mantle viscosity has negligible impact on the RSL predictions. However, toward the centre of loading in the north the impact of these changes becomes far more apparent. At all sites around the Ulster coastline, reduced upper mantle viscosity raises maximal predictions of RSL by, in the case of the v_{UM} 3 x 10^{20} Pa s model, in excess of 10m. Conversely, the magnitude (and geographical range) of the Holocene highstand is reduced: In west Donegal (14) it is lowered by c.2m by adopting the v_{UM} 4 x 10^{20} Pa s model, whilst the v_{UM} 3 x 10^{20} Pa s model has the effect of removing the highstand altogether. In fact, when adopting the low viscosity (v_{UM} 3 x 10^{20} Pa s) Earth model, the greatest magnitude of Holocene highstand in Ireland is no more than +2.5m.

Finally, we consider the effects of running our best fit model (Model E) with a lower upper mantle viscosity Earth model of 4 x 10^{22} Pa s. Reference to figures 5.16 and 5.17 shows that this

combination of a thin Irish ice sheet during the deglacial phase coupled with a reduced upper mantle Earth model does yield a close fit with the empirical RSL data. The Holocene highstand is removed from west Donegal (14) and model predictions are now in good agreement with the index point data from this locality. In north Antrim (18) (where the isostatic component of the RSL signal is at its greatest) the Holocene highstand is c. +5m. This figure is significantly lower than the c. +8m highstand predicted by running BIM- 1 with an earth model of v_{UM} 5 x 10^{20} Pa s (such as it was in the original Shennan *et al.*, analyses). Elsewhere around the Ulster coastline, model predictions satisfy all the constraints placed by the primary limiting date evidence as well as a large number of the secondary limiting dates.

6 DISCUSSION

The analyses undertaken in section 5 outlines the extent to which model predictions of RSL respond to adjustments in both ice and Earth model input parameters. Significant improvements in the fit between BIM- 1 predictions and the Irish RSL data can be achieved through altering the Irish component of this regional ice model yet changes in the ice model alone have not completely reconciled discrepancies with observations of RSL. In particular, we are still left with a Holocene highstand in west Donegal which is unsubstantiated by field evidence. Interestingly, a similar problem relating to the position of the 0m Holocene isobase on the west coast of Britain has been encountered in recent modelling studies of the British data (Shennan *et al.*, 2002; Shennan *et al.*, 2006) and, as in the case of this investigation, a solution could not be found through adjustments in the regional ice model. Model predictions of the geographical location of the 0m isobase can be marginally improved through increasing the eustatic melt water contribution since 7000 BP to a maximum of 3m (e.g. Lambeck, 2002). However, we are still left with predictions of a Holocene high stand in both west Donegal and north Wales which are at odds with available field evidence.

Both Milne *et al.*, (2006) and Shennan *et al.*, (2006) have shown that predictions of RSL in the post glacial are sensitive to ice sheet evolution in the pre-LGM period. For this reason, changes in the ice sheet thickness were considered during the temporary retreat phase (25 000 – 26 000 BP). However, removing this retreat stage only imparted a very minor GIA signal after the LGM. Because no further changes can be made to the Irish ice model without contradicting the empirical glaciological evidence, it would appear that the Earth model employed in the Shennan *et al.*, (2006) analyses is untenable with the Irish RSL evidence.

Significant alterations to BIM- 1 RSL predictions can also be achieved through lowering the upper mantle viscosity of the Earth model. A greatly improved fit between BIM- 1 predictions and the Irish RSL dataset is achieved by adopting an upper mantle viscosity of 4 x 10^{20} Pa s which is a similar value to that used by Lambeck and co workers in their earlier analyses from the region. However, the problematic Holocene high stand remains in west Donegal. Whilst this may be removed by lowering the upper mantle viscosity still further to a value of 3 x 10^{20} Pa s, the overall fit with available RSL data from the north of Ireland becomes extremely poor: The greatest Holocene highstand around the Ulster coastline is no more than 2.5m and the RSL predictions become significantly at odds with the Holocene index point data from south Down (20) and south west Scotland. In order to reconcile the differences between the BIM-1 (v_{UM} 3 x 10^{20} Pa s) and the Holocene RSL data, a substantial increase in crustal loading would be required during a large part of the late Devensian. However, the resultant rise in RSL in the late Devensian time interval would almost certainly be at odds with available geomorphological evidence of RSL from Ireland. Similarly, it is hard to see how such a low viscosity upper mantle Earth model would be compatible with the long time-series records of RSL change from the west coast of Scotland, particularly since trimline evidence provides a strict upper bound for the maximum thickness of ice at the time of the LGM. Because of the above, such a low viscosity upper mantle is extremely unlikely to provide an acceptable solution.

The above demonstrates that a resolution for model misfits cannot be achieved through changes to one or the other of the ice/Earth model. Instead, it is evident that combined revisions to both the existing BIM- 1 ice model and the original Earth model are desirable in order to achieve a solution that satisfies the constraints provided by both the RSL and glaciological evidence from Ireland. Indeed, running our best fit ice model (Model E) with a slightly reduced upper mantle viscosity of 4 x 10^{20} Pa s would appear to provide a very good overall solution. Using this ice-Earth model combination satisfies nearly all the late Devensian RSL evidence from around the Ulster coastline and in Donegal, the predicted 0m isobase now occupies the same geographical location as that reconstructed through field evidence (figure 5.17). On the Isle of Man and in North Wales, the fit with the Holocene data is considerably improved. Importantly, the well constrained 0m Holocene isobase on the east coast of Britain is largely unaffected by the revision to the Earth model.

An important, additional test of this revised Earth-ice model solution is the quality of the fit to the long time-series RSL records from Scotland (see figure 5.16). On first inspection, it appears as though the new reduced viscosity model is incompatible with the index point data from both Islay and Knapdale. However, this needn't hold true: BIM- 1 is constrained by glaciological evidence and is also inferred on the basis of the Earth model employed in the Shennan *et al.*, (2002) investigation. Significantly, the analyses of Shennan *et al.*, (2006) has shown that the 96km lithosphere used in the Shennan *et al.*, (2002) paper is untenable with observational RSL data from Britain. This means that there now exists considerable scope to make local scale adjustments in the British Ice model, specifically in areas which are poorly constrained by trimline data. This will be looked into in a future paper that will consider GIA model fits to the entire UK and Irish RSL data as well as observations of present-day crustal motions derived from Global Positioning System (GPS) measurements.

7 CONCLUSION

Observations of post-glacial RSL around Ireland are limited in both time and space and are generally less complete than records found in Britain. However, we have shown that the newly compiled and quality assessed Irish RSL database (Brooks & Edwards, 2006) does provide an important data set with which to test GIA models for Ireland and the UK. Assuming that the influence of lateral earth structure beneath the British Isles produces a minor impact on predictions of RSL in this region, we conclude that the Irish ice sheet depicted in Shennan *et al.*, (2006) is wholly incompatible with observations of late Devensian and Holocene RSL in Ireland. Employing the same Earth model as Shennan *et al.*, our 'best fit' ice model for Ireland comprises a thick, spatially extensive Irish ice sheet of around 700m over much of north and central Ireland at the LGM. Relatively thick ice out to the continental shelf off the west coast of Scotland and Ireland coupled with thick Irish Sea Ice is required to produce a sufficient fit with observational RSL data from the late Devensian.

We find that predictions of RSL are sensitive to the specific rate and timing of deglaciation prior to the LGM. The Irish RSL data strongly favour very rapid deglaciation after 21 000 BP: Early (pre 21 000 BP) deglaciation produces unacceptably low predictions of maximum late Devensian RSL in the north of Ireland and SW Scotland whilst gradual deglaciation between 20 000 to 16 000 BP deteriorates the fit with observations of Holocene RSL.

This study cannot offer unequivocal evidence for or against rejecting the South of Ireland End Moraine as the southerly limit of late Devensian Irish Ice. However, in order to fit the RSL data, a significant thickness of ice (c.500m) is needed over the Irish Midlands, which are only a short distance to the north of the South of Ireland End Moraine. If the moraine is taken as the true southern limit of LGM ice extent, it follows that the ice sheet would have had a very steep southern margin and this may be glaciologically implausible.

Significantly, all of the models considered in the analyses presented here deliver maximal predictions of late Devensian RSL along Ireland's west coast that are well below present msl. No evidence is found in support of the suggestion by McCabe, (2005); McCabe *et al.*, (1986; 2005) of higher than present late Devensian RSL along this stretch of coastline.

Only a limited fit with available records of RSL can be obtained by altering the Irish ice sheet component of BIM-1. Significant discrepancies with RSL data could not be resolved, particularly those in relation to the geographical position of the 0m Holocene isobase. However, it was possible to solve this problem by making a minor reduction to the value of upper mantle viscosity employed in the Earth model. Despite the significant changes in RSL predictions bought about through making this adjustment to the Earth model, our results still conclusively demonstrate that the Irish component of BIM-1 is not compatible with the RSL data. The primary conclusion from our ice modelling – the need for a thinner Irish ice Sheet soon after the LGM and throughout the deglacial phase - remains a robust requirement of the RSL data regardless of the adopted viscosity model.

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