2004

Holocene sea-level changes along the Channel coast of South-west England

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http://hdl.handle.net/10026.1/476

http://dx.doi.org/10.24382/3996
University of Plymouth

HOLOCENE SEA-LEVEL CHANGES ALONG THE CHANNEL COAST OF SOUTH-WEST ENGLAND

By

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A thesis submitted to the University of Plymouth

in partial fulfilment for the degree of

DOCTOR OF PHILOSOPHY

Department of Geography

School of Geography and Geology

May 2004

ABSTRACT

HOLOCENE SEA-LEVEL CHANGES ALONG THE CHANNEL COAST OF SOUTH-WEST ENGLAND

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A Holocene relative sea-level (RSL) history for the south Devon coast is reconstructed through an analysis of twelve cores extracted from the coastal back-barrier environments of Bantham Sands, North Sands, Slapton Sands and Blackpool Sands. Foraminifera preserved in the sediments are assigned an indicative meaning based on the vertical distribution of their modern counterparts from contemporary salt marsh and mudflat environments in the Erme and Salcombe-Kingsbridge estuaries. The contemporary data provide the first inter-tidal (–2.6 to +2.6 m MTL) foraminifera-based transfer function for south-west England from which sea-level changes can be predicted with good precision (RMSEP = ±0.29 m). Sea-level index points (SLIPs) from basal facies provide the long-term (10^3 yr) RSL history. SLIPs derived from non-basal minerogenic and peat sequences are subject to post-depositional consolidation and a vertical correction, using the Paul and Barras (1998) method, was therefore applied to these index points. The vertical displacement of fine minerogenic sediments ranges from <0.1 m above basal facies to >1 m at minerogenic-peat contacts, increasing to >2 m in organic peat facies. The age of each SLIP is obtained by AMS ^{14}C dating of bulk sediment or plant material. The electrical resistivity method of geophysical survey provides additional subsurface mapping information of the back-barrier sediments, and the lithostratigraphic results support the view that a complex barrier-lagoon system existed along the south Devon coastline during the early Holocene. Almost 13 m of RSL rise occurred along the south Devon coastline *ca.* 9000 to 4400 cal years BP. The rate of early Holocene RSL rise is ~5.4 ± 2.1 m/ka and ~1.1 ± 0.2 m/ka during the mid- to late Holocene. The pattern and rate of RSL rise is similar to other sea-level curves produced for the region. Lambeck's (1993a,b, 1995) geophysical model predictions fit the data well during the early Holocene but Peltier's (1998) model is the best overall fit of the robust data used to reconstruct the early to mid-Holocene RSL history. Rates of middle to late Holocene sea-level rise are generally faster than the models predict. The area is still undergoing glacio-isostatic adjustment (GIA) equating to a coastal land subsidence of –1.16 m/ka since *ca.* 4400 cal years BP. The dataset increases the current Holocene SLIP database for the south-west peninsula and southern Britain by 23% and for south Devon by almost fivefold.

Contents **Page**

Copyright Statement	i
Title Page	ii
Abstract	iii
List of Contents	iv
List of Figures	xi
List of Tables	xvi
List of Abbreviations	xviii
Acknowledgements	xix
Author's Declaration	xx
Appreciation	xxii
Dedication	xxiii

List of Contents **Page**

CHAPTER 1 **INTRODUCTION AND SCIENTIFIC BACKGROUND**

1.1	Introduction	1
1.2	Aims	2
1.3	Objectives	5
1.4	Previous Holocene sea-level studies in south-west Britain	6
1.4.1	The sea-level index point database for the region	6
1.4.2	Holocene sea-level histories from the literature	12
1.4.3	Models of barrier and back-barrier coastal evolution	22
1.5	Geophysical modelling	25
1.5.1	Lambeck's geophysical model	27
1.5.2	Peltier's geophysical model	29
1.5.3	Testing the models with published data from south Devon	31
1.6	Unresolved issues	33
1.6.1	Model differences and limitations	33
1.6.2	Radiocarbon dating	34
1.6.3	Autocompaction of sediments	36
1.6.4	Indicative meaning	37
1.6.5	Surveying errors	39
1.6.6	Selection of a local tidal datum and changes in tidal range	40
1.7	Study sites and rationale for selection	42

1.8 Hypotheses 44
1.9 Wider scope 45
1.10 Summary 46

CHAPTER 2 **RESEARCH METHODS**
2.1 Introduction 48
2.2 Field techniques 48
2.2.1 Coring 48
2.2.2 Surveying and datum selection 50
2.2.3 Electrical resistivity surveying 51
2.3 Laboratory techniques 52
2.3.1 Lithostratigraphical analysis 52
2.3.2 Foraminiferal analysis 54
2.3.3 Carbon analysis 56
2.3.4 AMS ^{14}C radiocarbon dating 57
2.3.5 Pollen analysis 59
2.3.6 Geotechnical correction for sediment autocompaction 60
2.3.7 Determination of the sea-level index point 61
2.4 Summary 62

CHAPTER 3 **TRANSFER FUNCTION**
3.1 Introduction 64
3.2 Site descriptions 67
3.2.1 The Erme estuary 67
3.2.2 The Salcombe-Kingsbridge estuary 68
3.2.3 Tidal regime 70
3.3 Methods 70
3.3.1 Field techniques 70
3.3.2 Laboratory techniques 71
3.4 Results 71
3.4.1 Modern analogue of dead foraminifera 71
3.4.2 Quantitative analyses 74
3.5 Discussion and interpretation 78
3.6 Summary 79

CHAPTER 4 **BANTHAM SANDS**

4.1 Site description 81
4.1.1 Geomorphology 82
4.1.2 Geology 83
4.1.3 Tidal regime 85
4.2 Lithostratigraphy 85
4.2.1 Core BA-97-1 85
4.2.2 Core BA-97-2 86
4.2.3 Core correlation 89
4.3 Electrical resistivity survey 90
4.3.1 Shore-parallel resistivity survey (Ban 1) 90
4.3.2 Shore-normal resistivity survey (Ban 2) 91
4.4 Biostratigraphy 93
4.4.1 Core BA-97-1 94
4.4.2 Core BA-97-2 97
4.5 Chronology 100
4.6 Palaeoenvironmental interpretation 102
4.7 Summary 103

CHAPTER 5 **NORTH SANDS**

5.1 Site description 105
5.1.1 Geomorphology 105
5.1.2 Geology 108
5.1.3 Tidal regime 109
5.2 Lithostratigraphy 109
5.2.1 Core NS-97-1 109
5.2.2 Core NS-97-2 110
5.2.3 Core NS-97-3 110
5.2.4 Core NS-97-4 115
5.2.5 Core correlation 115
5.3 Electrical resistivity survey 116
5.3.1 Shore-parallel resistivity survey (Nor 1) 117
5.3.2 Shore-normal resistivity survey (Nor 2) 117
5.4 Biostratigraphy 119
5.4.1 Core NS-97-1 120
5.4.2 Core NS-97-2 125

5.4.3 Core NS-97-3 129
5.4.4 Core NS-97-4 132
5.5 Carbon analyses 132
5.5.1 Total and organic carbon content of core NS-97-3 134
5.6 Geotechnical correction for sediment autocompaction 134
5.6.1 Core NS-97-2 135
5.6.2 Core NS-97-3 138
5.6.3 Summary and limitations of geotechnical correction 142
5.7 Chronology 142
5.7.1 North Sands sea-level index points 142
5.7.2 Pollen analyses of core NS-97-3 149
5.7.3 Summary of North Sands chronology 156
5.8 Palaeoenvironmental interpretation 157
5.9 Summary 159

CHAPTER 6 **SLAPTON SANDS**
6.1 Site description 161
6.1.1 Geomorphology 161
6.1.2 Geology 165
6.1.3 Tidal regime 166
6.2 Lithostratigraphy 168
6.2.1 Core SS-99-1 168
6.2.2 Core SS-99-2 168
6.2.3 Core SS-01-3 169
6.2.4 Core correlation 173
6.3 Electrical resistivity survey 174
6.3.1 Shore-parallel resistivity survey (Sla 1) 174
6.3.2 Shore-parallel resistivity survey (Sla 2) 174
6.4 Biostratigraphy 178
6.4.1 Core SS-99-1 178
6.4.2 Core SS-99-2 181
6.5 Chronology 184
6.6 Palaeoenvironmental interpretation 187
6.7 Summary 189

CHAPTER 7 **BLACKPOOL SANDS**

7.1 Site description 191

7.1.1 Geomorphology 191

7.1.2 Geology 194

7.1.3 Tidal regime 196

7.2 Lithostratigraphy 196

7.2.1 Core BS-97-1 200

7.2.2 Core BS-97-2 200

7.2.3 Core BS-97-3 201

7.2.4 Core correlation 201

7.3 Electrical resistivity survey 202

7.3.1 Shore-parallel resistivity survey (Bla 1) 203

7.3.2 Shore-parallel resistivity survey (Bla 2) 204

7.3.3 Shore-normal resistivity survey (Bla 3) 205

7.3.4 Shore-diagonal resistivity survey (Bla 4) 206

7.4 Biostratigraphy 207

7.4.1 Core BS-97-1 207

7.4.2 Core BS-97-2 208

7.4.3 Core BS-97-3 212

7.5 Geotechnical correction for sediment autocompaction 212

7.5.1 Core BS-97-2 213

7.5.2 Core BS-97-3 216

7.5.3 Summary and limitations of geotechnical correction 220

7.6 Chronology 221

7.7 Palaeoenvironmental interpretation 229

7.8 Summary 232

CHAPTER 8 **DISCUSSION**

8.1 Holocene relative sea-level change along the south coast of Devon 233

8.1.1 The early Holocene RSL history of south Devon 233

8.1.2 The mid-Holocene RSL history of south Devon 240

8.1.3 The late Holocene RSL history of south Devon 245

8.2 Testing the geophysical model predictions for south Devon 246

8.3 Crustal movement along the south Devon coastline 258

8.4 The Holocene coastal evolution of south Devon 263

8.5 Methodological issues 271

8.5.1 Autocompaction of sediments 271
8.5.2 Indicative meaning 273
8.5.3 Electrical resistivity surveying 275
8.5.4 Sea-level indicators 277
8.5.5 Identifying salt-marsh facies 279

CHAPTER 9 **CONCLUSIONS AND FUTURE WORK**

9.1 Original aims 282
9.2 Original hypotheses 282
9.3 Holocene sea-level changes in south Devon 283
9.4 Testing the geophysical model predictions for south Devon 284
9.5 Crustal movement in south Devon 284
9.6 The Holocene coastal evolution of south Devon 285
9.7 The electrical resistivity of back-barrier sediments 286
9.8 Autocompaction of back-barrier sediments 286
9.9 A new foraminiferal training set for sea-level reconstruction in south-west England 287
9.10 The problems with back-barrier sediments 287
9.10.1 Sea-level indicators 287
9.10.2 Identifying and radiocarbon dating fossil salt-marsh facies 288
9.11 Future work 289

References* 290

Appendices (CD-ROM) 330

APPENDIX 1 LITHOSTRATIGRAPHIC RECORDS

1a Sediment description of fossil coastal sediments using Troels-Smith (1955)
1b Key to core logs (adapted from Troels-Smith, 1955)
1c Site core logs (1 m sections)
1d Photographic record of core logs (entire cores and sub-sections)

APPENDIX 2 FOSSIL FORAMINIFERA RECORDS

2a Raw and percentage counts for all foraminiferal assemblages

*Including references cited in text contained in the appendices (CD-ROM)

APPENDIX 3 CARBON RESULTS

3a Carbon percentage values for core NS-97-3

APPENDIX 4 POLLEN RESULTS

4a Raw and percentage counts for pollen assemblages from core NS-97-3

APPENDIX 5 GEOTECHNICAL DATA

5a Geotechnical theory and standard methods

5b Laboratory results

5c Model results

APPENDIX 6 GEOPHYSICAL MODELLING DATA

6a Numbers for geophysical modelling runs by Lambeck (pers. comm. 1997, 2000) and Peltier (pers. comm. 2000, 2002)

APPENDIX 7 MODERN ANALOGUE DATA

7a Raw and percentage counts of live and dead foraminifera

List of Figures **Page**

CHAPTER 1 **INTRODUCTION AND SCIENTIFIC BACKGROUND**

1.1 Holocene RSL data from south Devon available prior to this study 3

1.2 Predicted Holocene RSL histories for Slapton Sands, south Devon 4

1.3 Estimated current rates of late Holocene crustal movement in Great Britain 5

1.4 Holocene RSL data points for south-west Britain 9

1.5 Annual tide-gauge records between 1915 and 2002 from Newlyn, south Cornwall 11

1.6 Holocene sea-level histories from south-west Britain compiled in the 1970s 13

1.7 Holocene sea level curves from south-west Britain 16

1.8 Holocene sea-level change in west Cornwall 19

1.9 A transgressive lagoon-barrier coastline by Kraft and Chrzastowski (1985) 24

1.10 Earth modelling by Walcott (1970) 26

1.11 Geophysical models versus the SLIP database 32

1.12 Location of study sites along the south Devon coastline 43

1.13 Comparing maps of late Holocene crustal movement in the British Isles 46

CHAPTER 2 **RESEARCH METHODS**

2.1 Percussion drilling rig in operation at Slapton Sands 49

2.2 Principal system layout for roll-along resistivity surveying using four electrode cables 51

2.3 Core cutting technique 53

2.4 Lithostratigraphical components key 54

2.5 The geotechnical basis of the Paul and Barras (1998) model 61

2.6 Colour-coded key for SLIPs and models shown in this thesis 62

CHAPTER 3 **TRANSFER FUNCTION**

3.1 Erme estuary and Frogmore Creek mudflat sites with transect locations 65

3.2 Cross-sections from the Erme estuary and Frogmore Creek 69

3.3 Modern surface analogue of dead foraminifera 73

3.4 Observed and predicted values of indicative meaning (m MTL) 77

CHAPTER 4 **BANTHAM SANDS**

4.1 Aerial view of Bantham Sands 82

4.2	Bantham Sands back-barrier system	83
4.3	The solid geology and Quaternary deposits of Bantham Sands and the surrounding country	84
4.4	Lithostratigraphy of core BA-97-1	87
4.5	Lithostratigraphy of core BA-97-2	88
4.6	Bantham Sands core correlation	89
4.7	Shore-parallel resistivity profile (Ban 1)	91
4.8	Shore-normal resistivity profile (Ban 2)	92
4.9	Enlarged view of shore-normal resistivity profile (Ban 2)	93
4.10	Foraminifera stratigraphy (%) of core BA-97-1: basal (lower contact) section	95
4.11	Foraminifera stratigraphy (%) of core BA-97-1: in-core (upper contact) section	96
4.12	Foraminifera stratigraphy (%) of core BA-97-2: basal (lower contact) section	98
4.13	Foraminifera stratigraphy (%) of core BA-97-2: in-core (upper contact) section	99
4.14	Bantham Sands SLIPs	101
4.15	Bantham Sands SLIPs and geophysical modelling results	101

CHAPTER 5 **NORTH SANDS**

5.1	Aerial view of North Sands	106
5.2	North Sands back-barrier system	107
5.3	The solid geology and Quaternary deposits of North Sands and the surrounding country	108
5.4	Lithostratigraphy of core NS-97-1	111
5.5	Lithostratigraphy of core NS-97-2	112
5.6	Lithostratigraphy of core NS-97-3	113
5.7	Lithostratigraphy of core NS-97-4	114
5.8	North Sands core correlation	116
5.9	Shore-parallel resistivity profile (Nor 1)	118
5.10	Shore-normal resistivity profile (Nor 2)	119
5.11	Foraminifera stratigraphy (%) of core NS-97-1: basal (lower contact) section	121
5.12	Profile of the upper contact in core NS-97-1	122
5.13	Foraminifera stratigraphy (%) of core NS-97-1:	

	in-core (upper contact) minerogenic section	123
5.14	Foraminifera stratigraphy (%) of core NS-97-1: in-core (upper contact) peat section	124
5.15	Foraminifera stratigraphy (%) of core NS-97-2: basal (lower contact) section	126
5.16	Profile of the upper contact in core NS-97-2	127
5.17	Foraminifera stratigraphy (%) of core NS-97-2: in-core (upper contact) section	128
5.18	Foraminifera stratigraphy (%) of core NS-97-3: basal (lower contact) to in-core (upper contact) section	130
5.19	Foraminifera stratigraphy (%) of core NS-97-4: basal (lower contact) to in-core (upper contact) section	131
5.20	Total and organic carbon content (%) of core NS-97-3	133
5.21	Input geotechnical data from core NS-97-2	135
5.22	Additional geotechnical data from core NS-97-2	137
5.23	Total geotechnical correction for core NS-97-2	138
5.24	Input geotechnical data from core NS-97-3	139
5.25	Additional geotechnical data from core NS-97-3	140
5.26	Total geotechnical correction for core NS-97-3	141
5.27	North Sands SLIPs	147
5.28	Age/depth plot of index points from core NS-97-3	148
5.29	Pollen stratigraphy (%) of core NS-97-3: basal (lower contact) to in-core (upper contact) section	151
5.30	Corrected age/depth plot of index points from core NS-97-3	154
5.31	North Sands SLIPs and geophysical modelling results	156

CHAPTER 6 **SLAPTON SANDS**

6.1	Aerial view of Slapton Sands	163
6.2	Slapton Sands back-barrier system	164
6.3	The solid geology and Quaternary deposits of Slapton Sands and the surrounding country	165
6.4	Lithostratigraphy of core SS-99-1	170
6.5	Lithostratigraphy of core SS-99-2	171
6.6	Lithostratigraphy of core SS-01-3	172
6.7	Slapton Sands core correlation	173
6.8	Shore-parallel resistivity profile (Sla 1)	175

6.9	Shore-parallel resistivity profile (Sla 2)	176
6.10	Enlarged view of Sla 2 alongside core SS-99-1	177
6.11	Enlarged view of Sla 2 alongside core SS-99-2	177
6.12	Foraminifera stratigraphy (%) of core SS-99-1: basal (lower contact) section	179
6.13	Foraminifera stratigraphy (%) of core SS-99-1: in-core (upper contact) section	180
6.14	Foraminifera stratigraphy (%) of core SS-99-2: basal (lower contact) section	182
6.15	Foraminifera stratigraphy (%) of core SS-99-2: in-core (upper contact) section	183
6.16	Slapton Sands SLIPs	186
6.17	Slapton Sands SLIPs and geophysical modelling results	187

CHAPTER 7 **BLACKPOOL SANDS**

7.1	Aerial view of Blackpool Sands	192
7.2	Blackpool Sands back-barrier system	193
7.3	The solid geology and Quaternary deposits of Blackpool Sands and the surrounding country	195
7.4	Lithostratigraphy of core BS-97-1	197
7.5	Lithostratigraphy of core BS-97-2	198
7.6	Lithostratigraphy of core BS-97-3	199
7.7	Blackpool Sands core correlation	202
7.8	Shore-parallel resistivity profile (Bla 1)	203
7.9	Shore-parallel resistivity profile (Bla 2)	204
7.10	Shore-normal resistivity profile (Bla 3)	205
7.11	Shore-diagonal resistivity profile (Bla 4)	206
7.12	Foraminifera stratigraphy (%) of core BS-97-1: basal (lower contact) to in-core (upper contact) section	209
7.13	Foraminifera stratigraphy (%) of core BS-97-2: basal (lower contact) to in-core (upper contact) section	210
7.14	Foraminifera stratigraphy (%) of core BS-97-3: basal (lower contact) to in-core (upper contact) section	211
7.15	Input geotechnical data from core BS-97-2	213
7.16	Additional geotechnical data from core BS-97-2	214
7.17	Total geotechnical correction for core BS-97-2	216

7.18 Input geotechnical data from core BS-97-3 217
7.19 Additional geotechnical data from core BS-97-3 218
7.20 Total geotechnical correction for core BS-97-3 220
7.21 Blackpool Sands SLIPs 225
7.22 Age/depth plot of index points from core BS-97-1 226
7.23 Blackpool Sands SLIPs and geophysical modelling results 229

CHAPTER 8 **DISCUSSION**

8.1 Holocene RSL history for south Devon 234
8.2 Holocene RSL history for south-west Britain 235
8.3 Holocene RSL history for south Devon and Cornwall: I 238
8.4 Holocene RSL history for south Devon and Cornwall: II 239
8.5 Summary of Peltier's (1998) RSL model predictions for south Devon 247
8.6 Eustatic sea-level rise (ICE-4G model history) 248
8.7 Post-glacial RSL responses to additional ICE-4G (VM2+90km) components 249
8.8 Ice-equivalent eustatic sea-level rise scenarios employed by Lambeck (1990) 250
8.9 South Devon's Holocene RSL history versus geophysical modelling results 254
8.10 ICE-4G (VM1 and 2): Comparisons of predicted and observed sea-level data from the Fenlands, East Anglia and the Bristol Channel (Peltier, 1998) 256
8.11 GB-3: Comparisons of predicted and observed sea-level data from the Fenlands, East Anglia and Bridgwater, Bristol Bay (Lambeck *et al.*, 1996) 257
8.12 The late Holocene estimate of crustal subsidence for south Devon 261
8.13 Late Holocene crustal movement in Great Britain plus new data from south Devon 263
8.14 A transgressive model of landward barrier movement for south Devon 267
8.15 The resistivity ranges of back-barrier sediments from south Devon 276

List of Tables **Page**

CHAPTER 1 **INTRODUCTION AND SCIENTIFIC BACKGROUND**

1.1 The Holocene SLIP database for south-west Britain 7

1.2 Rates of Holocene sea-level rise for south-west Britain 10

1.3 Accuracy of Ordnance Survey benchmark levelling 40

CHAPTER 3 **TRANSFER FUNCTION**

3.1 Tidal heights for field sites interpolated from nearest tidal stations 70

3.2 DCCA on training sets of dead, live and total foraminifera 74

3.3 Performance of the regression models 76

CHAPTER 4 **BANTHAM SANDS**

4.1 Radiocarbon dates and associated sample information of archaeological finds at Bantham 81

4.2 Tidal heights for Bantham Sands interpolated from nearest tidal stations 85

4.3 Summary of the electrical resistivity of sediments at Bantham Sands 93

4.4 Radiocarbon dates and associated foraminiferal counts 100

CHAPTER 5 **NORTH SANDS**

5.1 Tidal heights for Salcombe Harbour 109

5.2 Summary of the electrical resistivity of sediments at North Sands 119

5.3 Radiocarbon dates and associated foraminiferal counts 143

5.4 Non-parametric (Mann-Whitney Test) analyses of biofacies in core NS-97-3 155

5.5 Corrected SLIP heights for proposed double sample section in core NS-97-3 157

CHAPTER 6 **SLAPTON SANDS**

6.1 Radiocarbon dates and associated sample information from the Slapton coastline 162

6.2 Tidal heights for Slapton Sands interpolated from the nearest tidal stations 167

6.3 Summary of the electrical resistivity of sediments at Slapton Sands 178

6.4 Radiocarbon dates and associated foraminiferal counts 185

CHAPTER 7 **BLACKPOOL SANDS**

7.1 Sea-level index point data from the submerged forest on Blackpool Sands 193

7.2 Tidal heights for Blackpool Sands interpolated from the nearest tidal stations 196

7.3 Summary of the electrical resistivity of sediments at Blackpool Sands 207

7.4 Radiocarbon dates and associated foraminiferal counts 222

CHAPTER 8 **DISCUSSION**

8.1 Holocene RSL rates for south Devon 235

8.2 Palaeoenvironmental information associated with SLIPs from Cornwall and Devon 242

8.3 Sea-level data obtained from Tresco, Isles of Scilly 246

8.4 Modelled Holocene RSL rates for south Devon 252

8.5 Estimates of crustal subsidence along the south Devon coastline 260

8.6 Summary of the median electrical resistivity values at all sites 276

List of Abbreviations

Acronyms used in this thesis are listed here in alphabetical order:

AMS (^{14}C)	Accelerated Mass Spectrometry (Carbon-14)
BP	Before Present
CANOCO	Canonical Community Ordination
CAR	Correspondence Analysis Regression
Cc	Coefficient of Consolidation
CLIMAP	Climate and Environment Monitoring with GPS Atmospheric Profiling
CONISS	Constrained Incremental Sum of Squares Cluster Analysis
DCCA	Detrended Canonical Correspondence Analysis
EDM	Electronic Distance Measurer
EIA	Environmental Impact Assessment
GIA	Glacio-Isostatic-Adjustment
GPS	Global Positioning System
HAT	Highest Astronomical Tide (Tidal station: Devonport, Plymouth)
HCl (acid)	Hydrochloric Acid
LGM	Last/Late Glacial Maximum
LL	Liquid Limits
MHW	Mean High Water
MHWOT	Mean High Water of Ordinary Tides
MHWST	Mean High Water of Spring Tides
MLWST	Mean Low Water of Spring Tides
MSL	Mean Sea Level
MTL	Mean Tide Level
NERC	Natural Environment Research Council
OD	Ordnance Datum (Newlyn)
OSGB	Ordnance Survey of Great Britain
OS(GR)	Ordnance Survey (Grid Reference)
PCR	Principal Components Regression
PLS	Partial Least Squares
RMSEP	Root Mean Square Error Prediction
RSL	Relative Sea Level
SD	Standard Deviation
SE	Standard Error
SLI(s)	Sea-Level Indicator(s)
SLIP(s)	Sea-Level Index Point(s)
SPECMAP	Spectral Mapping (using Standard Spectral Coordinate Mapping Systems)
SWLI(s)	Standard Water Level Index(es)
TGView	TILIA*GRAPH Viewer
TSPPlus	Troels-Smith Plotting Programme
WA	Weighted Averaging
WA-PLS	Weighted Averaging – Partial Least Squares
WA-Tol	Weighted Averaging with Tolerance Downweighting

Computer programmes (not acronyms) used in this project to record, display or manipulate data are listed here in alphabetical order:

CALIB 4.2	For the calibration of ^{14}C (radiocarbon) ages
CALIBRATE	For the graphical display and statistical analysis of (multivariate) species/environment data
TILIA	To record stratigraphic data
TILIA*GRAPH	To display stratigraphic data

Acknowledgements

This research project was completed (on 8th September 2003) with (HEFCE) funding from The Faculty of Science and Department of Geographical Sciences, University of Plymouth. The Royal Society are thanked for providing funding for 11 cores and the Natural Environment Research Council Radiocarbon Laboratory, in particular Dr Charlotte Bryant, are thanked for providing 31 radiocarbon ages (allocation nos. 835.1299 and 883.0800 in February 2000 and February 2001 respectively) which were fundamental to this project.

My interest in sea-level change, marine micropalaeontology and the physical sciences in general came from an intellectually stimulating experience as an undergraduate at the University of Plymouth. My current supervisors, Dr Roland Gehrels and Professor Dan Charman, played a significant part in that education. Their sincere belief in my ability to achieve the required standard for completion of this Ph.D. has been similarly inspirational. I also thank Bob Turner and Val Howey of Chippenham Technical College for their encouragement during and after my GCE A Level Geography education.

Professor Dick Peltier, Dr Rosemarie Drummond, Professor Kurt Lambeck and Dr Tony Purcell are thanked for providing me with their geophysical model RSL predictions for the south Devon sites. Thanks to Professor Mike Paul for geotechnical modelling assistance and Dr Katherine Selby for her work on pollen analysis. Thanks to Professor Ian Shennan and Dr Ben Horton for providing me with the Holocene SLIPs and associated data for south-west Britain from the National Database, and Sarah White for sharing her Erme salt-marsh foraminifera data with me.

There are many people to thank for fieldwork assistance, in particular my supervisors Dr Roland Gehrels and Professor Dan Charman. Dr Graeme Taylor and Ian King are thanked for helping me with the electrical resistivity surveys, Dr Jason Kirby for his help in collecting modern analogue samples and Will Marshall for assisting with the site surveys. Thanks to Mr J. S. Aylett of Michelmore Hughes Chartered Surveyors for access permission to core Bantham Sands and South Hams District Council for allowing us to core at North Sands. It was a pleasure to work with all at the Slapton Ley Field Centre, especially Chris Riley who liased with us a great deal throughout the duration of the project. Thanks also to Sir Geoffrey Newman for access to Blackpool Sands. Thanks to Anthony Mildmay-White for access permission to the Erme estuary salt marsh and South Hams District Council for allowing me to work on the Frogmore Creek mudflats.

The list is similarly long for those who assisted in the laboratory and technical support of this project. Science Officers Ann Kelly, Richard Hartley, Kevin Solman, Pat Bloomfield and Bob Head are the best laboratory (and field) support a project student could hope for. Andy Elmes, Pauline Framingham and David Antwis are thanked for their excellent IT support and Tim Absalom, Brian Rogers, Gareth Johnston, Maz Harzuki and Jamie Quinn for their equally first rate cartographic assistance. Dr Allan Jones is thanked for his tuition on ArcInfo, ArcMap and MapManager and Martin Cox for his help in transforming OSGB to latitude and longitude data. Thanks to David Griffiths from Media Services for his photographic work and Brian Moran from Technical Services for constructing the core cutting tool and presentation box. I also thank Professor David Pinder for his support throughout this project, Dr Rewi Newnham for his excellent work during my transfer from M.Phil. to Ph.D., and Dr Antony Long and Professor Neil Roberts for the Ph.D. *viva voce*.

No project of this nature could be achieved without moral support along the way and for that I thank my mother and father, Elaine and Douglas, my best friends Darren Massey, Peter Ian Copeland and Paul Lewis. When all around me was falling apart during my *annus horribilis* you were always there for me, and for that I am eternally grateful.

Author's Declaration

At no time during the registration for the degree of Doctor of Philosophy has the author been registered for any other University award.

This study was (HEFCE) financed by the Department of Geographical Sciences (50%) and the Faculty of Science (50%), University of Plymouth, plus additional field costs from The Royal Society and awards for radiocarbon dates from the Natural Environment Research Council.

The following conferences and courses were attended:

November 1998	Slapton Research Seminar, Slapton, Devon, UK.
April 1999	Postgraduate Palaeoecology Conference, University of Newcastle upon Tyne, UK.
June 1999	Quaternary Research Association Open Lecture by Professor Kurt Lambeck, University of Wales, Bangor, North Wales, UK.
June 1999	Application of Benthic Foraminifera and Thecamoebians for Detecting Environmental Effects, University of Bologna, Italy.
November 1999	Slapton Research Seminar, Slapton, Devon, UK.
January 2000	Quaternary Research Association Annual Discussion Meeting, University of Southampton, Hampshire, UK.
February 2000	Introduction to Diatom Micropalaeontology, University College London, UK.
November 2000	Slapton Research Seminar, Slapton, Devon, UK.
December 2000	IGCP Project 437, UK Working Group Meeting, The Institute of Materials, London, UK.
September 2001	The 3rd International Conference of The International Geological Correlation Programme (IGCP), Project No.437, Durham and Fort William, Scotland, UK.
November 2002	Environment and Archaeology in Devon, University of Exeter, Devon, UK.
November 2002	Slapton Research Seminar, Slapton, Devon, UK.

Presentations were made at the following conferences:

April 1999	Postgraduate Palaeoecology Conference, University of Newcastle upon Tyne, UK.
May 1999	Departmental Lecture Series, University of Plymouth, Devon, UK.
November 1999	Slapton Research Seminar, Slapton, Devon, UK.
December 2000	IGCP Project 437, UK Working Group Meeting, The Institute of Materials, London, UK.
November 2000	Slapton Research Seminar, Slapton, Devon, UK.
March 2001	Departmental Lecture Series, University of Plymouth, Devon, UK.
September 2001	The 3rd International Conference of The International Geological Correlation Programme (IGCP), Project No.437, Durham and Fort William, Scotland, UK.
November 2002	Environment and Archaeology in Devon, University of Exeter, Devon, UK.
November 2002	Slapton Research Seminar, Slapton, Devon, UK.

News Releases

April 2000	Sea-Level Changes in the South West, University of Plymouth, Devon, UK.
April 2000	Sea-Level Study, Plymouth Evening Herald, Devon, UK.
April 2000	Tracking Sea-Level Changes, Sunday Independent, Bristol & Somerset, UK.
June 2000	Sea-Level Changes, UPfront, University of Plymouth, Devon, UK.
June 2000	Sea-Level Changes in Devon, Pirate FM Radio News, UK.
June 2000	Storm Warning, ITV (Carlton) Documentary, UK.
March 2004	Sea-level Changes in the South West Region and the Isles of Scilly - contribution to the forthcoming BBC Natural History Series (October 2004), BBC South West Region.

Publications:

Gehrels, W.R., and Massey, A.C. 2004. Slapton Line Coast Protection: Impact of Sea-Level Rise. *In*: Scott Wilson Consultants (eds). In prep. *A379 Slapton Sands Main Study.* A report prepared for English Nature. Devon County Council, Exeter, Devon, UK.

Kirby, J.R., Waller, M.P., and Massey, A.C. In prep. Holocene coastal evolution of the eastern English Channel: new data from the Rye area to test geophysical models (for *Marine Geology*).

Massey, A.C. 2001. Slapton Sands Borehole Report. *In*: W. S. Atkins Consultants Ltd. (eds). *A379 Slapton Sands Interim Reinstatement and Environmental Statement, May 2001.* A report prepared for English Nature. Devon County Council, Exeter, Devon, UK.

Massey, A.C., and Gehrels, W.R. In prep. The infaunal habitat of salt-marsh foraminifera in south-west Britain (for *Journal of Micropalaeontology*).

Massey, A.C., Gehrels, W.R., Charman, D.J., and White, S.V. In prep. An inter-tidal foraminifera-based transfer function for reconstructing Holocene sea-level change in south-west England (for *Journal of Foraminiferal Research*).

Massey, A.C., Paul, M.A., Gehrels, W.R., and Charman, D.J. In prep. Autocompaction in Holocene coastal back-barrier sediments, south Devon, UK (for *Marine Geology*).

Massey, A.C., Peltier, W.R., Gehrels, W.R., Charman, D.J., Selby, K.A., and Paul, M.A. In prep. Testing geophysical models of postglacial isostatic adjustment with new Holocene relative sea-level data from south Devon, south-west England (for *Quaternary Science Reviews*).

Massey, A.C., Taylor, G.K., Gehrels, W.R., and Charman, D.J. In prep. The electrical resistivity of coastal back-barrier sediments in south Devon, UK (for *Journal of Coastal Research*).

Signed...... A C Massey ......

Date...... 5/5/04 ......

Appreciation

Seven days after successful completion of my Ph.D. viva voce on 9th October 2003 I was taken seriously ill at home. I will be forever indebted to the staff at Bath Royal United Hospital Intensive Care Unit and Marlborough Ward. Without their professionalism and courage I would not be celebrating this great personal achievement.

Dedication

This Ph.D. thesis is dedicated to my mother and father, Elaine and Douglas Massey, my brother and best friend Darren Massey and in fond memory of our dearest friend Henry James August Le Coz.

Chapter 1

Introduction and Scientific Background

1.1 Introduction

Holocene sea level research in the British Isles has gathered pace since Godwin (1940) published the first known sea-level curve for the epoch in England (Tooley, 1978). The chronology for his 34 sea-level index points (SLIPs) was calculated prior to the inception of radiocarbon dating in the late 1940s by Libby and co-workers (Libby, 1955), and was based on pollen and archaeological data. Eighteen years later, Godwin *et al.* (1958) stated: "the development of radiocarbon dating has now placed in our hands the means of securing absolute dating of the submerged peat beds and former strand lines which are the evidence of submergence". Although we have long since known that radiocarbon dates are not absolute, due to atmospheric fluctuations of ^{14}C (Olsson, 1970; Smith and Pilcher, 1973) the statement was quite profound. Post-glacial sea level studies with ^{14}C dates went on to propagate at a significant rate in the British Isles and elsewhere around the globe (e.g., Jelgersma, 1961; Redfield and Rubin, 1962; Greensmith and Tucker, 1976; Devoy, 1977; Shennan, 1980, 1986a,b; Fairbanks, 1989; Gehrels, 1994, 1999; Shennan *et al.*, 2000).

The advent of radiocarbon dating amplified the quest for a single global eustatic post-glacial sea-level curve (Godwin *et al.*, 1958; Fairbridge, 1961; Jelgersma, 1961, 1966; Shepard, 1963; Mörner, 1969). These works were followed by attempts to "define the history of local or regional sea level" (Faure, 1980). Reasons for the shift in focus, from the global to local scale, are well documented in the literature (Smith and Dawson, 1983; van de Plassche, 1986; Tooley and Shennan, 1987; Devoy, 1987). However, these discussions were already preceded by a number of attempts at unravelling regional and local scale Holocene sea-level changes, particularly in the British Isles (e.g., Greensmith and Tucker, 1973; Tooley, 1974; Devoy, 1979). Churchill (1965a,b) provided even earlier evidence for regional and local scale sea and land-level movements in south-west Britain. Eight years later Kidson and Heyworth (1973) combined their data with Churchill's (1965a,b) and others (e.g., Godwin *et al.*, 1958; Godwin, 1960, 1964; Godwin and Willis, 1961) into a "eustatic sea-level curve" for the Bristol Channel. These authors were amongst some of the first providers of ^{14}C dated SLIPs for the region. Similarly, Clarke (1970) was one of the first researchers to investigate Holocene sea-level changes, supported by ^{14}C dating, in south Devon. Yet despite the work of Clarke (1970), Hawkins (1971, 1979), Hails (1975a,b) and Morey (1976, 1983a), research on the nature of Holocene relative sea-level changes in south Devon remains limited, largely due to the low number of SLIPs that have been produced.

1.2 Aims

This project seeks to improve the current record of Holocene sea- and land-level changes along the Channel coast of south-west England. Its aims are threefold:

1) The primary aim of this thesis is to establish a record of middle and early Holocene relative sea-level (RSL) change for the south Devon coast using a variety of palaeoenvironmental techniques. There is a genuine scientific requirement for the provision of a local post-glacial sea-level history of south Devon. Shennan (1983) and Waller and Long (2003) commented on the paucity of Holocene sea-level data available for the south west of England. However, very few "detailed local studies of small homogenous areas" have since been carried out in the region. Much of the work (e.g., Hawkins, 1971) is discussed at the regional scale (Shennan, 1983). Currently, only a few isolated ^{14}C dates are locally available and to illustrate this problem, the SLIP database, held at the University of Durham (Environmental Research Centre, 2003) contains only six screened SLIPs from Devon of which five are from the south Devon coastline (Figure 1.1 and Table 1.1).

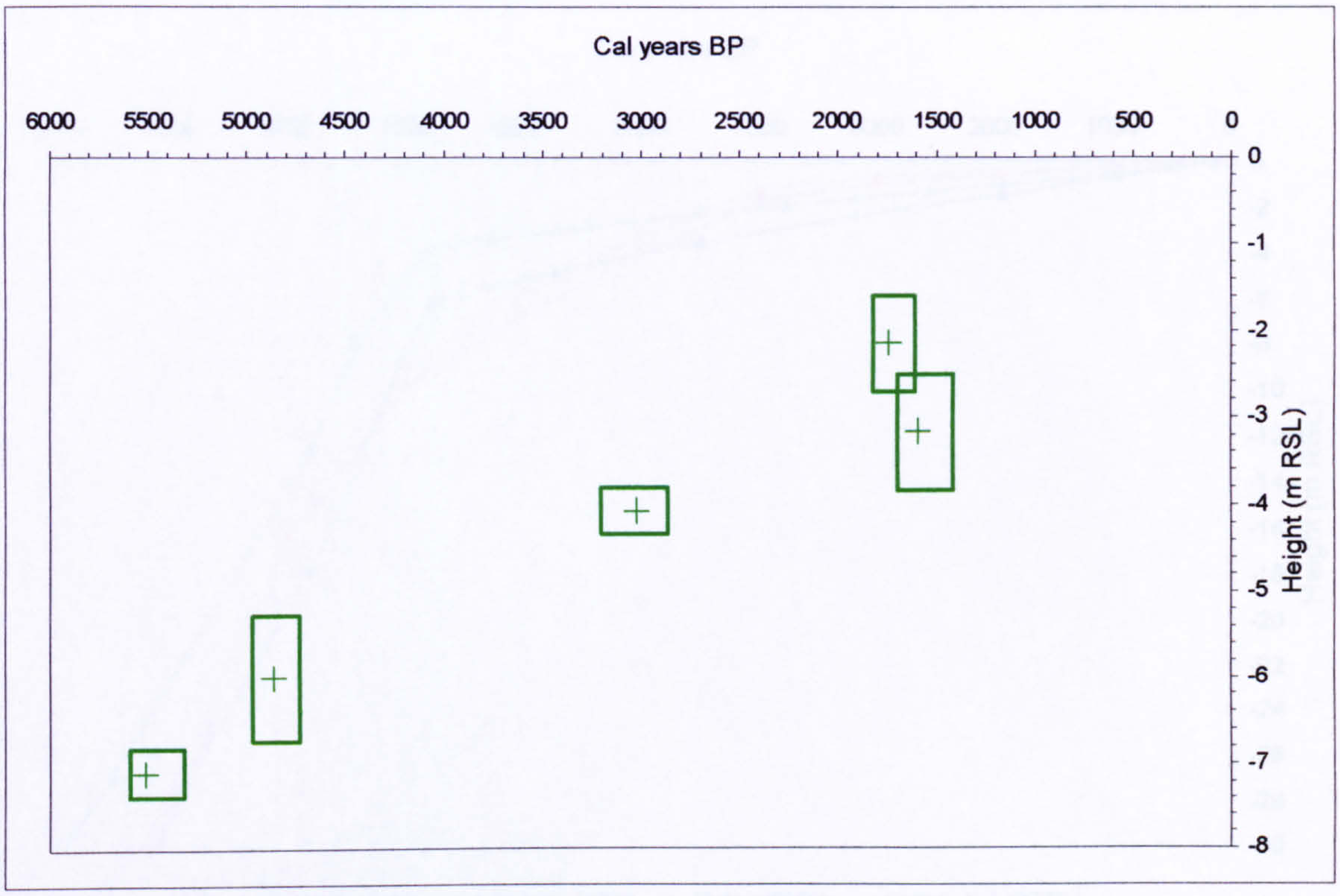


Figure 1.1. Holocene RSL data from south Devon available prior to this study. Index numbers correspond with Table 1.1. Crosses indicate the median 2σ calibrated ^{14}C age and mid-point of MTL range. Source: Environmental Research Centre, University of Durham (2003).

2) The secondary aim of this thesis is to test current geophysical models of ice retreat and Earth rheology against south Devon's Holocene RSL history. Geophysicists routinely compare "actual" sea-level histories with predicted sea-level responses produced by their models (e.g., Lambeck, 1993a,b, 1995; Peltier, 1998; Peltier *et al.*, 2002). Although isolated data points from authors such as Heyworth and Kidson (1982) and Healy (1995) have been used by these geophysicists, there is presently no evidence of a single local Holocene sea-level curve from the south west of England having been used to test current Earth models. The problem faced by the modellers is that, as well as the scarcity of data, that which is available is not of sufficient quality to test the models. The indicative meaning of the available SLIPs, for example, has not been quantified. Also, basal SLIPs are not available (see section 1.6.3). Post-glacial sea-level predictions for the south Devon coastline supplied by Lambeck (pers. comm. 1997, 2000) and Peltier (pers. comm. 2000, 2002) are tested in this project (Figure 1.2).

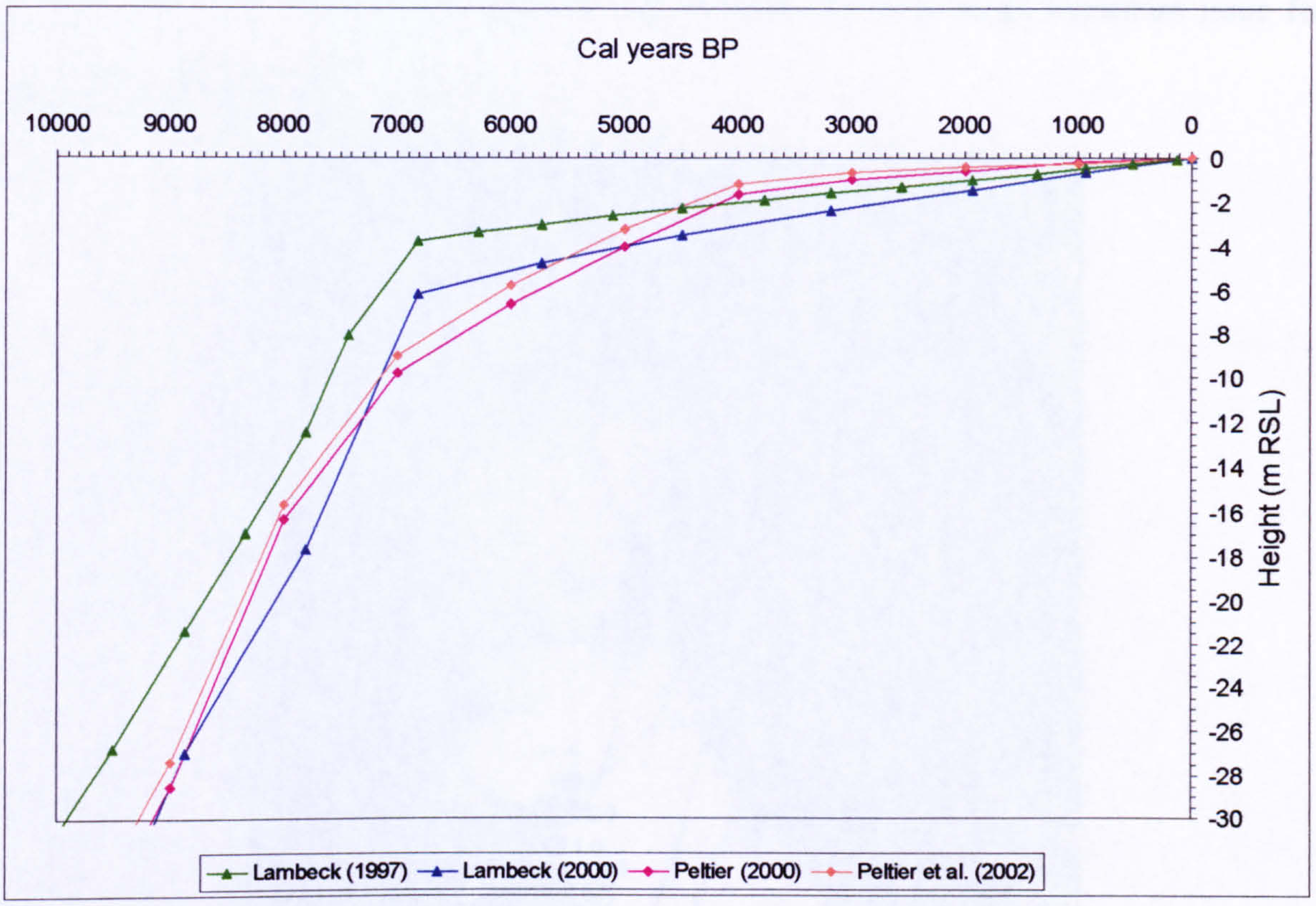


Figure 1.2. Predicted Holocene RSL histories for Slapton Sands, south Devon (SX 825 432). Based on the geophysical models of Lambeck (1993a,b, 1995), Peltier (1998) and modifications by Peltier *et al.* (2002).

3) The tertiary aim of this thesis is to calculate rates of crustal motion along the south Devon coastline. Shennan and Horton (2002, Figure 1.3) estimated rates of crustal subsidence in south-west England to be between 1.0 and 1.2 mm yr^{-1} narrowing considerably the previous estimate of 0.1 to 1.4 mm yr^{-1} (Shennan, 1989). However, a sea-level history based on multiple SLIPs is not available for any coastal location in the region, SLIPs are isolated and scattered. The wider issue has been addressed by Graff (1981) who suggested that future predictions of the return period of extreme sea-level events may be affected by uncertainties in rates of crustal motion as little as ~1 mm yr^{-1}. For example, a 10 cm predicted additional rise in sea level over the next 100 years may make a 1 in 100 year extreme event a 1 in 10 year event. It is important to note that a level of uncertainty for the region of ~1 mm yr^{-1} is highly significant given that the entire range for the British Isles is ~3 mm yr^{-1} (Figure 1.3). Although the earlier range of uncertainty (1.3 mm yr^{-1}, Shennan, 1989) has been reduced to 0.2 mm yr^{-1} (Shennan and Horton, 2002) the limited

dataset used to produce these figures clearly earmark this aim as an important issue for scientific consideration.

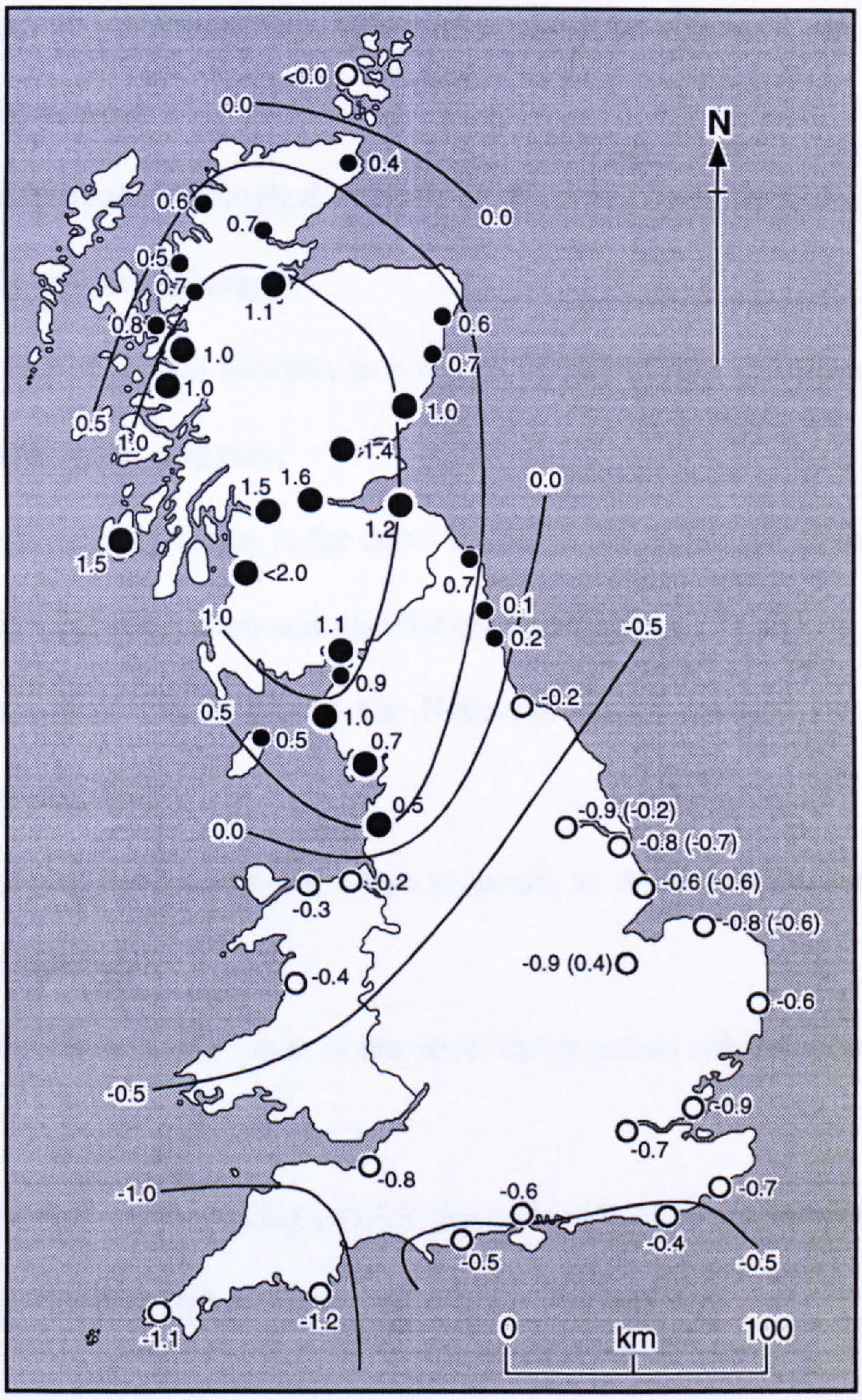


Figure 1.3. Estimated current rates (mm yr^{-1}) of late Holocene crustal movement in Great Britain. Positive values indicate relative land uplift or sea-level fall, negative values are relative land subsidence or sea-level rise. Figures in parentheses take into account modelled tidal range changes during the Holocene (from Shennan and Horton, 2002).

1.3 Objectives

To accomplish the stated aims a Holocene RSL history for south Devon, consisting of a series of carefully screened SLIPs, is produced (see chapters 4 to 8). The following objectives explain how this thesis will set out to achieve the aims. They are:

1) to describe the lithostratigraphy of 12 cores from four neighbouring back-barrier systems along the south Devon coastline;
2) to obtain additional stratigraphical information using the electrical resistivity technique of geophysical survey;
3) to carry out micropalaeontological analysis of the cores using foraminifera as a tool for reconstructing sea-level changes;
4) to assess the use of carbon analysis as a means of identifying salt-marsh facies prior to micropalaeontological analyses;
5) to interpret sedimentary facies in the cores based on the collection of modern analogues sampled from local salt-marsh and mudflat environments;
6) to identify sea-level trends during the Holocene epoch assisted by the radiocarbon dating of core samples;
7) to apply palynological techniques, when required, as a relative chronology in addition to radiocarbon dating;
8) to apply a geotechnical correction to sea-level index points subject to autocompaction.

The scientific rationale and methodology for these objectives is described elsewhere in this and the following chapters (see sections 1.6, 2.2, 2.3, 3.2 and 3.3).

1.4 Previous Holocene sea level studies in south-west Britain

1.4.1 The sea-level index point database for the region

The majority of Holocene sea-level data currently available for south-west Britain has been obtained from the English and Welsh coastlines along the Bristol and English Channels (Table 1.1) and include 104 ^{14}C dated SLIPs recorded in the SLIP database (Environmental Research Centre, 2003). Of these, 78 are listed as originating from research carried out in the Bristol Channel area, including the coastlines of North Devon, North Cornwall, Glamorgan and Pembrokeshire. The southern coastline of Cornwall, Devon and Dorset accounts for only 15 of the data points for south-west Britain. The remaining 11 index points are from the Hampshire coastline in southern England. The low number of SLIPs derived from the English Channel coastline of south-west England clearly pinpoints this locality as one for further research.

Index no.	Area	Site	Laboratory code	Latitude (N)	Longitude (W)	C14 Age	Error (1 Ss. dev.)	Median age (Cal yrs BP)	Cal +	Cal -	RSL (m)	Vertical error (m)
1	BC	Godney Island Somerset Levels	GU3247	5110118	00243464	2560	50	2623	147	154	-2.38	0.21
2	BC	Godney Island Somerset Levels	GU3246	5110118	00243464	2590	50	2724	63	241	-2.38	0.21
3	BC	Tealham Somerset	Q126	5112180	00251320	5620	120	6420	299	239	-6.16	0.25
4	BC	Burnham Somerset	Q134	5113440	00258090	6262	130	7152	272	347	-10.77	0.54
5	BC	Goldcliff 2	CAR778	5131600	00253380	5660	80	6452	180	152	-5.96	0.21
6	BC	Goldcliff 2	CAR776	5131600	00253380	5480	80	6270	158	269	-5.03	0.70
7	BC	Goldcliff 1	CAR657	5132030	00253250	5530	90	6329	174	326	-5.81	0.21
8	BC	Weston-super-mare Somerset	IGS41	5121080	00256230	4530	105	5173	290	304	-5.93	0.32
9	BC	Goldcliff 1	CAR656	5132030	00253250	5360	80	6132	159	189	-6.28	0.21
10	BC	Goldcliff 1	CAR655	5132030	00253250	5440	80	6225	172	228	-6.22	0.21
11	BC	Goldcliff 1	CAR654	5132030	00253250	5090	80	5825	165	174	-6.08	0.21
12	BC	Goldcliff 1	CAR645	5132030	00253250	3440	70	3704	164	225	-4.82	0.80
13	BC	Goldcliff 1	CAR644	5132030	00253250	3130	70	3350	193	234	-4.53	0.21
14	BC	Porlock Forest Bed	Be86775	5113070	00336590	7070	50	7886	83	126	-12.8	0.21
15	BC	Elmore	Be81686	5149416	00219355	2360	60	2425	286	242	-2.73	0.21
16	BC	Goldcliff 1	CAR658	5132030	00253250	5850	80	6658	195	203	-6.66	0.21
17	BC	Bridgwater Somerset	I2690	5111430	00303330	7360	140	8168	235	291	-14.52	0.22
18	BC	Kenn Moor Avon	IGS26	5125230	00248560	6100	100	6972	263	237	-8.24	0.23
19	BC	Avonmouth	IGS27	5132480	00239170	3110	100	3312	242	309	-2.74	0.47
20	BC	Avonmouth	IGS28	5132480	00239170	3905	100	4330	454	342	-3.35	0.25
21	BC	Kingston Seymour	I4846	5123540	00251450	3690	110	4036	364	317	-5	0.47
22	BC	Porlock Bay Somerset	SRR440	5113100	00337060	5120	60	5849	141	126	-9.8	0.41
23	BC	Kingston Seymour	I4844	5123540	00251450	5600	110	6398	260	215	-6.15	0.47
24	BC	Portbury	I4842	5128390	00242360	4240	105	4759	452	318	-3.3	0.41
25	BC	Highbridge Somerset	I4403	5113030	00257300	8360	140	9337	219	331	-27.07	0.82
26	BC	Highbridge Somerset	I4402	5113100	00257240	8480	140	9474	415	439	-25.87	0.80
27	BC	Bridgwater Somerset	I3713	5113290	00259250	7320	120	8130	235	251	-14.22	0.82
28	BC	Stolford	I3397	5112220	00306450	5330	120	6107	287	341	-6.77	0.43
29	BC	Stolford Somerset	NPL146	5112220	00306450	3460	90	3730	229	253	-6.37	0.07
30	BC	Stolford	I3395	5112040	00307250	4790	120	5514	369	436	-3.47	0.42
31	BC	Tealham Moor Somerset	Q120	5112180	00251320	5412	130	6186	258	271	-6.16	0.25
32	BC	Bridgwater Somerset	I2689	5111430	00303330	6890	120	7734	211	223	-12.09	0.23
33	BC	Bridgwater Somerset	I2688	5113030	00306150	7060	160	7879	294	284	-13.335	0.83
34	BC	Clevedon Somerset	IGS34	5125430	00252500	5815	115	6618	268	298	-7.2	0.85
35	BC	Clevedon Somerset	IGS35	5125400	00252100	5360	120	6132	266	224	-6.07	0.85
36	BC	Kenn Pier Somerset	IGS36	5125250	00248530	3510	100	3790	291	305	-3.61	0.50
37	BC	Kenn Pier Somerset	IGS39	5125250	00248530	6100	120	6974	277	295	-8.24	0.32
38	BC	Weston-super-mare Somerset	IGS40	5121080	00256230	3675	100	4014	328	298	-4.2	0.32

Table 1.1. The Holocene SLIP database for south-west Britain. Source: Environmental Research Centre, University of Durham (2003). Data are from the coastlines of South Wales (SWAL), Bristol Channel (BC), Gloucestershire (BC), Avon (BC), Somerset (BC), Cornwall (SW3), Devon (SW2), Dorset (SW1) and Hampshire (SC). Latitude and longitude is abbreviated, e.g., 5110118 00243464 = 51°10′.118N 002°43′.464W.

Index no.	Area	Site	Laboratory code	Latitude (N)	Longitude (W)	C14 Age	Error (1 Ss. dev.)	Median age (Cal yrs BP)	Cal +	Cal -	RSL (m)	Vertical error (m)
39	BC	Highbridge Somerset	IGS53	5113040	00257330	8365	100	9359	174	268	-27.37	0.81
40	BC	Porlock Forest Bed	Be81655	5113070	00336590	7730	50	8502	85	87	-13.47	0.40
41	BC	Stolford Somerset	NPL147	5112300	00306340	5380	95	6149	162	217	-6.77	0.12
42	BC	Bridgwater Somerset	NPL148	5112350	00306220	6230	95	7121	288	313	-7.57	0.22
43	BC	Stolford	I3396	5112220	00306450	5250	140	6031	267	317	-5.27	0.42
44	BC	Tarnock Somerset	Be142352	5116105	00253095	3190	70	3418	207	173	-4.25	0.23
45	BC	Llanwern Monmouth	Q691	5135130	00254340	2660	110	2775	223	411	-3.11	0.27
46	BC	Porlock Forest Bed	Be81654	5113070	00336590	6160	70	7057	185	249	-8.57	0.40
47	BC	Goldcliff 1	CAR659	5132030	00253250	5959	80	6792	209	224	-6.72	0.21
48	BC	Godney Island Somerset Levels	Q2458	5110118	00243464	2860	50	2983	176	135	-2.38	0.22
49	BC	Caldicot Pill Oscar	B79887	5134600	00243000	6360	70	7298	128	204	-7.52	0.22
50	BC	Nyland Hill 4 Somerset	Be101740	5114524	00246230	3370	60	3609	211	144	-1.92	0.21
51	BC	Nyland Hill 11 Somerset	Be101741	5114492	00246492	3380	60	3621	200	152	-3.38	0.21
52	BC	Nyland Hill 12 Somerset	Be101742	5114508	00246204	3250	80	3484	200	212	-1.99	0.21
53	BC	Godney Island Somerset Levels	Q2459	5110118	00243464	2550	50	2608	156	224	-1.97	0.22
54	BC	North Yeo Farm Somerset	Be142351	5117113	00255145	4640	60	5398	183	331	-6	0.23
55	BC	Caldicot Pill, 333	B79886	5134600	00243000	6660	80	7535	125	108	-9.86	0.21
56	BC	Tarnock Somerset	Be142353	5116105	00253095	5370	50	6145	134	149	-6.06	0.23
57	BC	Burnham-on-sea	WK5297	5114397	00300107	5590	70	6378	152	171	-9.07	0.21
58	BC	Burnham-on-sea	WK5298	5114397	00300107	6340	70	7275	146	241	-9.28	0.21
59	BC	Porlock Forest Bed	Be61544	5113040	00336580	6870	90	7712	211	141	-11.67	0.80
60	BC	Porlock Forest Bed	Be81653	5113040	00336580	5070	60	5817	105	156	-6.46	0.70
61	BC	Slimbridge	Be80696	5143125	00223013	3110	50	3330	119	157	-2.28	0.21
62	BC	Longney - R Severn	Be80693	5148367	00220273	2340	60	2373	335	219	-3.48	0.21
63	BC	North Yeo Farm Somerset	Be142350	5117113	00255145	3600	70	3908	179	207	-4.87	0.23
64	BC	Shapwick Heath Somerset	Q423	5109150	00248290	5510	120	6300	250	309	-5.87	0.32
65	BC	Porlock Marsh	Be61543	5113010	00336250	5140	100	5887	284	230	-5.28	0.22
66	BC	Burnham on Sea	WK5300	5114397	00300107	4790	70	5518	128	194	-5.59	0.21
67	BC	Porlock Marsh	Be61542	5113010	00336250	5250	180	6025	370	405	-5.99	0.22
68	BC	Wick Farm Somerset	Be142355	5117062	00258358	5210	80	5982	206	233	-6.61	0.23
69	BC	Wick Farm Somerset	Be142354	5117062	00258358	3500	70	3774	195	188	-5.29	0.23
70	BC	Burnham-on-sea	WK5299	5114397	00300107	5370	70	6142	147	185	-6.01	0.21
71	SC	Stansore Point	HV17323	5047080	00120370	5600	180	6400	381	447	-5.29	0.22
72	SC	Stansore Point	HV17327	5047080	00120370	2480	75	2559	172	198	-1.68	0.22
73	SC	Stansore Point	HV17326	5047080	00120370	2350	110	2420	311	287	-2.04	0.22
74	SC	Stansore Point	HV17325	5047080	00120370	3570	105	3871	278	284	-2.66	0.22
75	SC	Stansore Point	HV17324	5047080	00120370	5320	200	6092	396	440	-5.14	0.21
76	SC	Fawley Hants	Q832	5049270	00121400	3563	96	3860	283	266	-2.75	0.27
77	SC	Hythe Marshes 2	Be93196	5051274	00122549	1250	50	1185	91	121	-0.69	0.21
78	SC	Fawley Hants	Q831	5049270	00121400	3689	120	4036	366	320	-4.1	0.30
79	SC	Stansore Point	HV17322	5047080	00120370	5565	130	6363	294	363	-5.98	0.41
80	SC	Fawley Hants	Q834	5049270	00121400	6368	124	7281	273	326	-8.88	0.47
81	SC	Hythe Marshes 1	Be93198	5051048	00122066	5320	60	6094	179	158	-5.02	0.21
82	SW1	Chesil Beach Dorset	I3429	5037410	00233310	4980	120	5735	249	265	-4.5	0.22
83	SW1	Poole Harbour Arne-5	Be87925	5041246	00201318	3130	60	3353	117	178	-1.64	0.21
84	SW1	Dorset	I5307	5035180	00229120	6290	120	7190	236	292	-6.845	0.87
85	SW1	Chesil Beach Dorset	I3431	5037380	00233310	5410	115	6185	217	248	-5.71	0.22
86	SW2	Westward Hol Devon	Q672	5102370	00414140	6585	130	7480	190	219	-6.42	0.55
87	SW2	Beesands Devon	SRR165	5015240	00339150	4767	45	5517	79	191	-7.085	0.25
88	SW2	North Hallsands Devon	SRR317	5014150	00339280	1680	50	1591	118	167	-3.165	0.65
89	SW2	Beesands Devon	SRR164	5015240	00339150	4302	45	4868	162	135	-5.985	0.72

Table 1.1 continued. The Holocene SLIP database for south-west Britain. Source: Environmental Research Centre, University of Durham (2003). See caption on page 7 for explanation.

Index no.	Area	Site	Laboratory code	Latitude (N)	Longitude (W)	C14 Age	Error (1 Ss. dev.)	Median age (Cal yrs BP)	Cal +	Cal -	RSL (m)	Vertical error (m)
90	SW2	Slapton Ley Devon	SRR492	5016580	00339080	1810	40	1748	112	134	-2.155	0.55
91	SW2	Slapton Ley Devon	SRR493	5016580	00339080	2890	50	3028	174	150	-4.055	0.23
92	SW3	Marazion Marsh	Q2777	5007540	00528500	5050	80	5801	127	185	-8.06	0.21
93	SW3	Marazion Marsh	Q2780	5007410	00528590	4380	55	4960	304	123	-7.06	0.21
94	SW3	Marazion Marsh	Q2776	5007540	00528500	5100	65	5830	154	165	-6.81	0.21
95	SW3	Marazion Marsh	Q2775	5007440	00528540	1210	40	1134	125	128	-1.75	0.70
96	SW3	Marazion Marsh	Q2774	5007440	00528540	4395	60	4986	293	143	-7.63	0.21
97	SW3	Marazion Marsh	Q2778	5007400	00529040	1610	40	1495	108	89	-1.67	0.70
98	SW3	Trewornan	Q2781	5032329	00449433	6460	80	7372	181	123	-12.7	0.40
99	SWAL	Margam Glamorgan	Q265	5134050	00345380	3402	108	3658	242	264	-9.2	0.78
100	SWAL	Port Talbot Glamorgan	Q662	5135410	00347250	8990	170	10075	423	510	-25.66	0.26
101	SWAL	Margam Glamorgan	Q275	5134050	00345380	6184	143	7067	348	333	-10.6	0.78
102	SWAL	Freshwater West Pembrokeshire	Q530	5139300	00503560	5960	120	6800	356	307	-5.09	0.84
103	SWAL	Margam Glamorgan	Q274	5134050	00345380	5605	126	6405	316	340	-10.1	0.85
104	SWAL	Port Talbot Glamorgan	Q663	5135410	00347250	8970	160	10049	428	451	-25.6	0.26

Table 1.1 continued. The Holocene SLIP database for south-west Britain. Source: Environmental Research Centre, University of Durham (2003). See caption on page 7 for explanation.

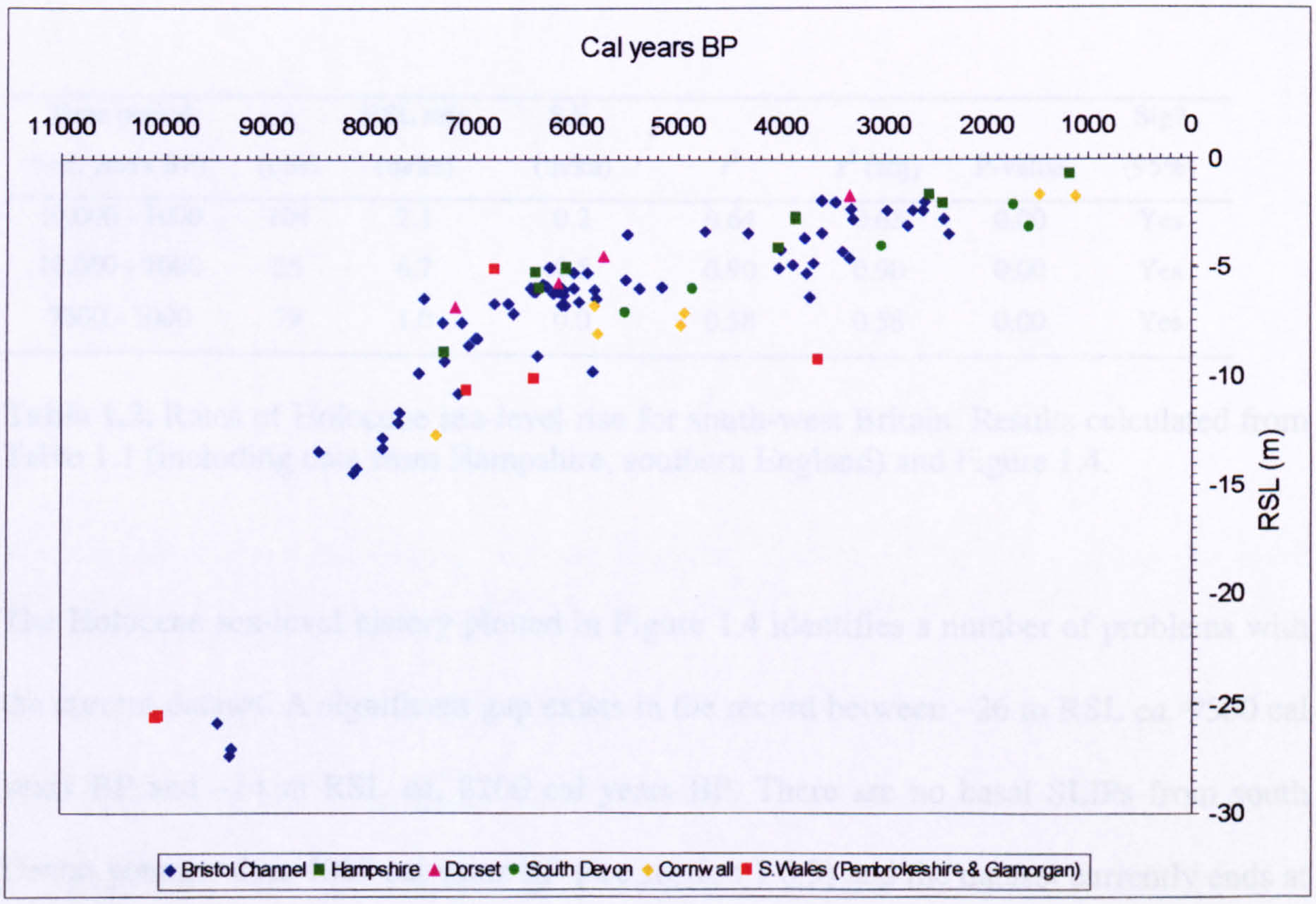


Figure 1.4. Holocene RSL data points for south-west Britain. Data plotted from Table 1.1. Source: Environmental Research Centre, University of Durham (2003).

The Holocene RSL history for south-west Britain reveals a distinctive trend (Figure 1.4). Sea level rose sharply from 10,000 cal years BP to an inflection at *ca.* 7000 BP (this thesis uses calendar years BP, unless years are explicitly stated in ^{14}C years), then continued rising at a more relaxed pace to the present day. This reduction is in part due to a significant decline in the volume of meltwater input to the world's oceans and is not a local signal (Lambeck, 1995; Peltier, 1998). For example, similar RSL histories have been observed in the Fenlands, England and further afield along the U.S. east coast (Peltier, 1998). These sea-level trends are reflected in the geophysical modelling results shown in Figure 1.2. The rate of early Holocene sea-level rise is about 7 m/ka and reduces to almost 1 m/ka during the mid- to late Holocene. The overall rate of Holocene sea-level rise in south-west Britain is about 2 m/ka but this thesis distinguishes between early and mid- to late Holocene rates due to the apparent inflection in the data *ca.* 7000 cal years BP (Figure 1.4 and Table 1.2).

Time period (cal. years BP)	n (obs)	RSL rate (m/ka)	S.E. (m/ka)	r^2	r^2 (adj)	P-value	Sig? (95%)
10,000 - 1000	104	2.1	0.2	0.64	0.63	0.00	Yes
10,000 - 7000	25	6.7	0.5	0.90	0.90	0.00	Yes
7000 - 1000	79	1.0	0.0	0.58	0.58	0.00	Yes

Table 1.2. Rates of Holocene sea-level rise for south-west Britain. Results calculated from Table 1.1 (including data from Hampshire, southern England) and Figure 1.4.

The Holocene sea-level history plotted in Figure 1.4 identifies a number of problems with the current dataset. A significant gap exists in the record between –26 m RSL *ca.* 9500 cal years BP and –14 m RSL *ca.* 8200 cal years BP. There are no basal SLIPs from south Devon younger than 4000 cal years BP (see section 1.6.3) and the dataset currently ends at *ca.* 1100 cal years BP. Crucially, the vertical range of index points reaches >7 metres between *ca.* 7500 and 3500 cal years BP. In particular, the SLIP from Margam, South Wales (*ca.* 3650 cal years BP at –9.2 m RSL, Lab. code Q265, Table 1.1) plots at least 3m below data of a similar age from the Bristol Channel (Figure 1.4). This SLIP is from compacted material (Heyworth and Kidson, 1982) and doubts are raised about its original height of deposition. This trend in the spread of early to mid-Holocene SLIPs is also reflected in geophysical modelling results where a vertical disparity of approximately >6 metres can be observed between models *ca.* 7000 to 6800 cal years BP (Figure 1.2). Further research is required to produce a continuous record of Holocene sea-level change, reduce vertical errors and improve geophysical modelling predictions for the region.

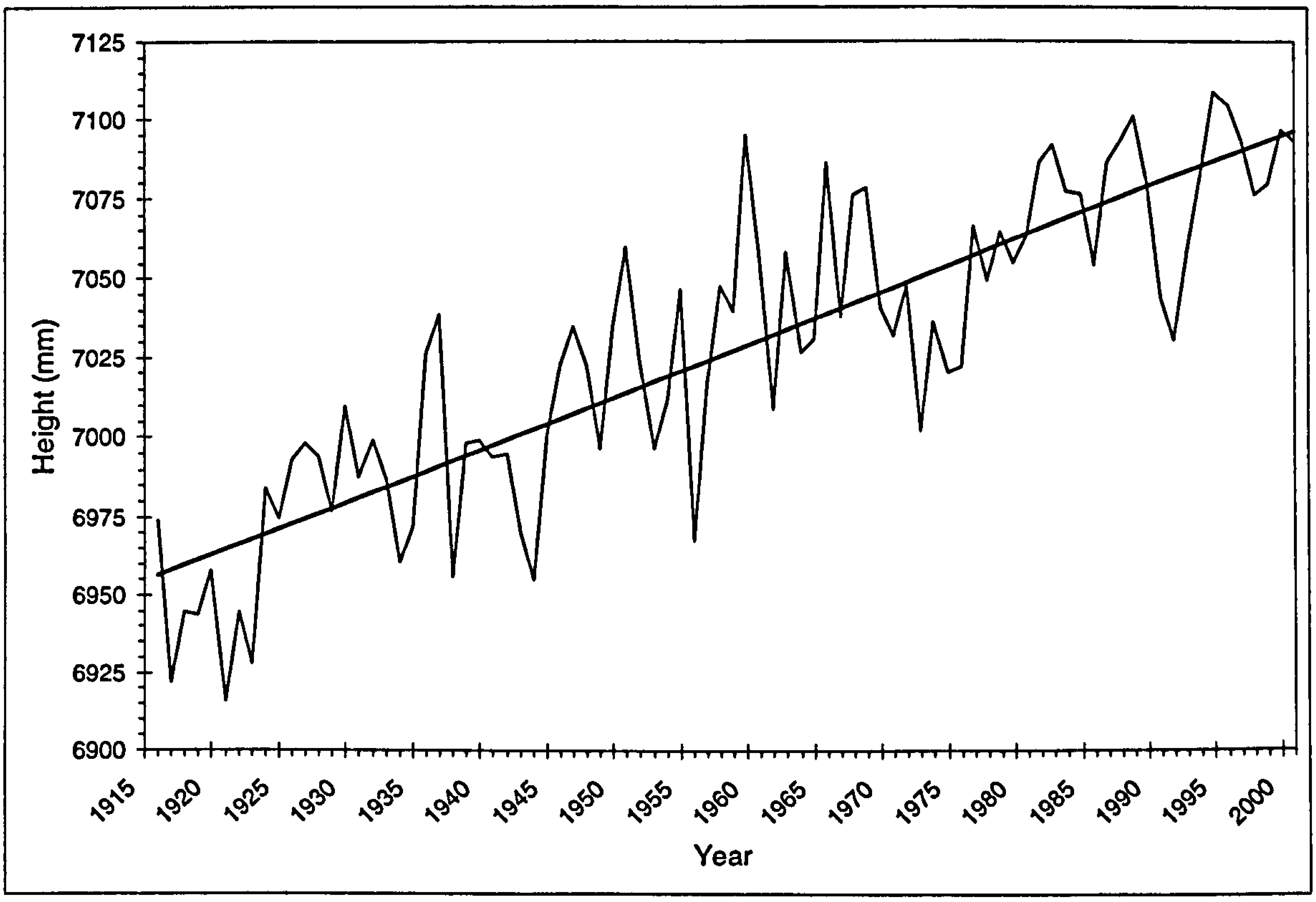


Figure 1.5. Annual tide-gauge records between 1915 and 2002 from Newlyn, south Cornwall. Source: http://www.pol.ac.uk/psmsl/pubi/rlr.annual.data/170161.rlrdata. (2003). Trendline = 1.64 mm yr^{-1}.

Tide gauges provide continuous records of RSL changes, albeit on a much shorter (hourly to 10^2 yr) time-scale. For example, the stations at Sheerness, Aberdeen and North Shields have been recording since 1832, 1862 and 1895 respectively (Woodworth, 1987). The Newlyn tide-gauge station in south Cornwall has been in operation since 1915 and provides the longest record for south-west Britain (Figure 1.5). It has been exceptionally well maintained during the last 87 years, due to its importance for the Ordnance Datum, and is uniquely sensitive to sea-level changes in the Atlantic and across the English

Channel (Woodworth, 1987). During the last 78 years the Newlyn tide gauge has recorded a 1.64 ± 0.12 mm yr^{-1} rise in mean sea level ($r^2 = 0.73$, $p = 0.00$, sig. at 95%). This is 0.66 mm yr^{-1} faster than data from the mid- to late Holocene (Table 1.2) and could be associated with climatic warming (Wigley and Raper, 1987, 1992; UKCIP, 2004) or a range of geophysical factors. The link between trends in RSL and crustal movement, as explored by geophysicists (Walcott, 1975; Lambeck, 1995; Peltier, 1998), has also been investigated by the use of tide-gauge records (Fleming, 1982; Emery and Aubrey, 1985, 1991; Shennan, 1989; Shennan and Woodworth, 1992).

1.4.2 Holocene sea-level histories from the literature

The number of publications pertaining to Holocene land- and sea-level movements in south-west Britain is surprisingly limited (Waller and Long, 2003), given that important and related scientific issues remain unresolved (see section 1.6). For example, some authors have remarked that all or part of the region has been subsiding since the last Pleistocene deglaciation (Hawkins, 1971; Emery and Aubrey, 1985; Woodworth, 1987; Allen and Rae, 1988; Shennan, 1989; Allen 1991; Shennan and Woodworth, 1992) whilst others have argued for recent isostatic and tectonic stability (Churchill, 1965a,b; Kidson and Heyworth, 1973, 1976, 1978; Heyworth and Kidson, 1982). However, an assessment of the data that is presently available should leave the reader in no doubt that post-glacial sea-level histories for south-west Britain reveal an upward trend (Allen, 1990). A comparison of four published Holocene sea-level curves from around the region successfully illustrates this point (Figure 1.6). Unfortunately, many of these curves are not considered precise reconstructions of sea-level change (Shennan, 1983; Haslett *et al.*, 1998) and cannot therefore reliably elucidate opposing theories of subsidence and stability in the region (Shennan, 1983, 1989). This is primarily related to vertical errors such as inaccuracies in indicative meaning, the lack of correction for autocompaction on non-basal index points, and problems associated with survey and tidal datums.

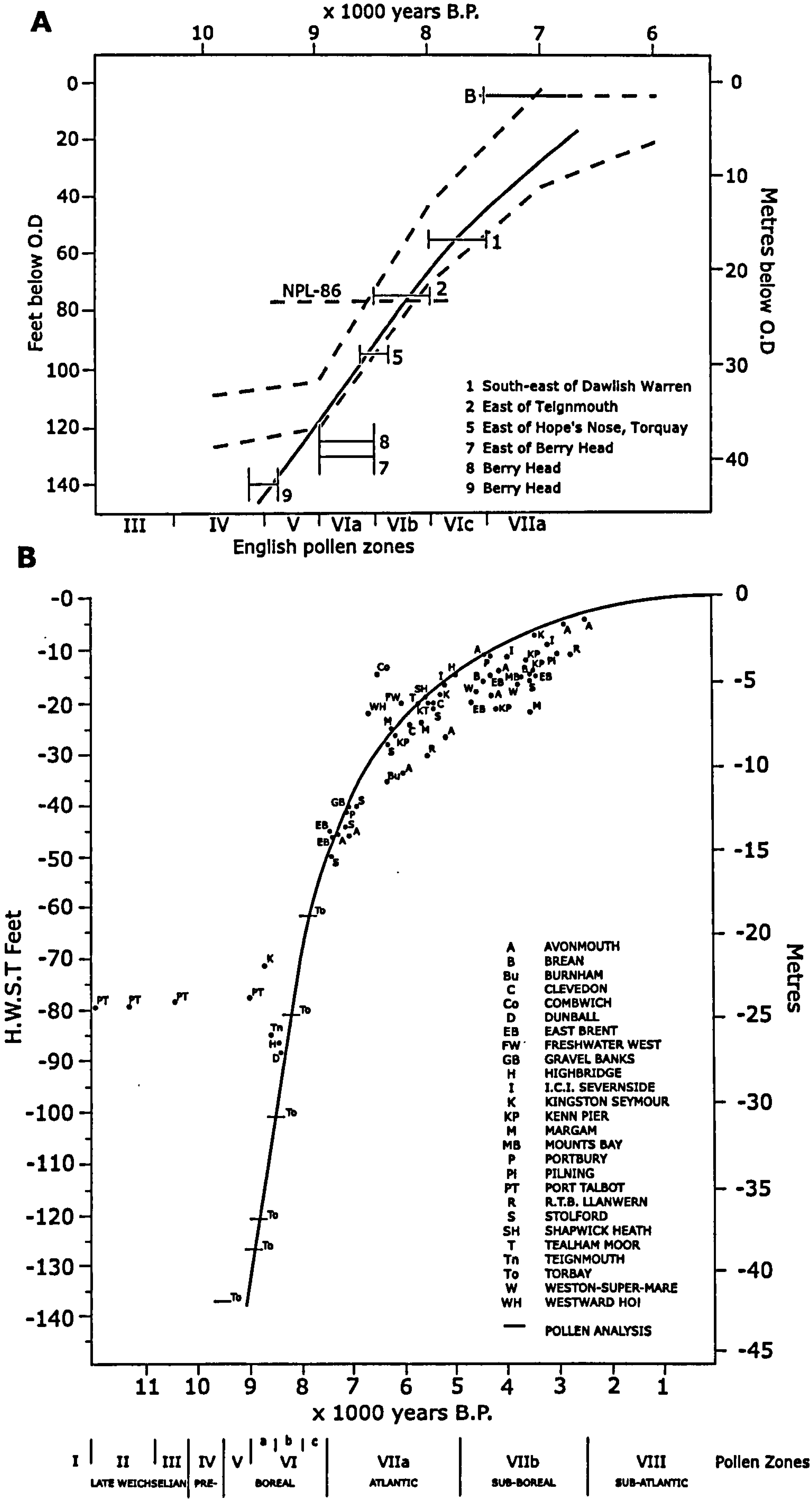


Figure 1.6. Holocene sea-level histories from south-west Britain compiled in the 1970s. **A.** Clarke, 1970: south-east Devon (dotted line is Clarke's comparison with Schofield, 1964); **B.** Hawkins, 1979.

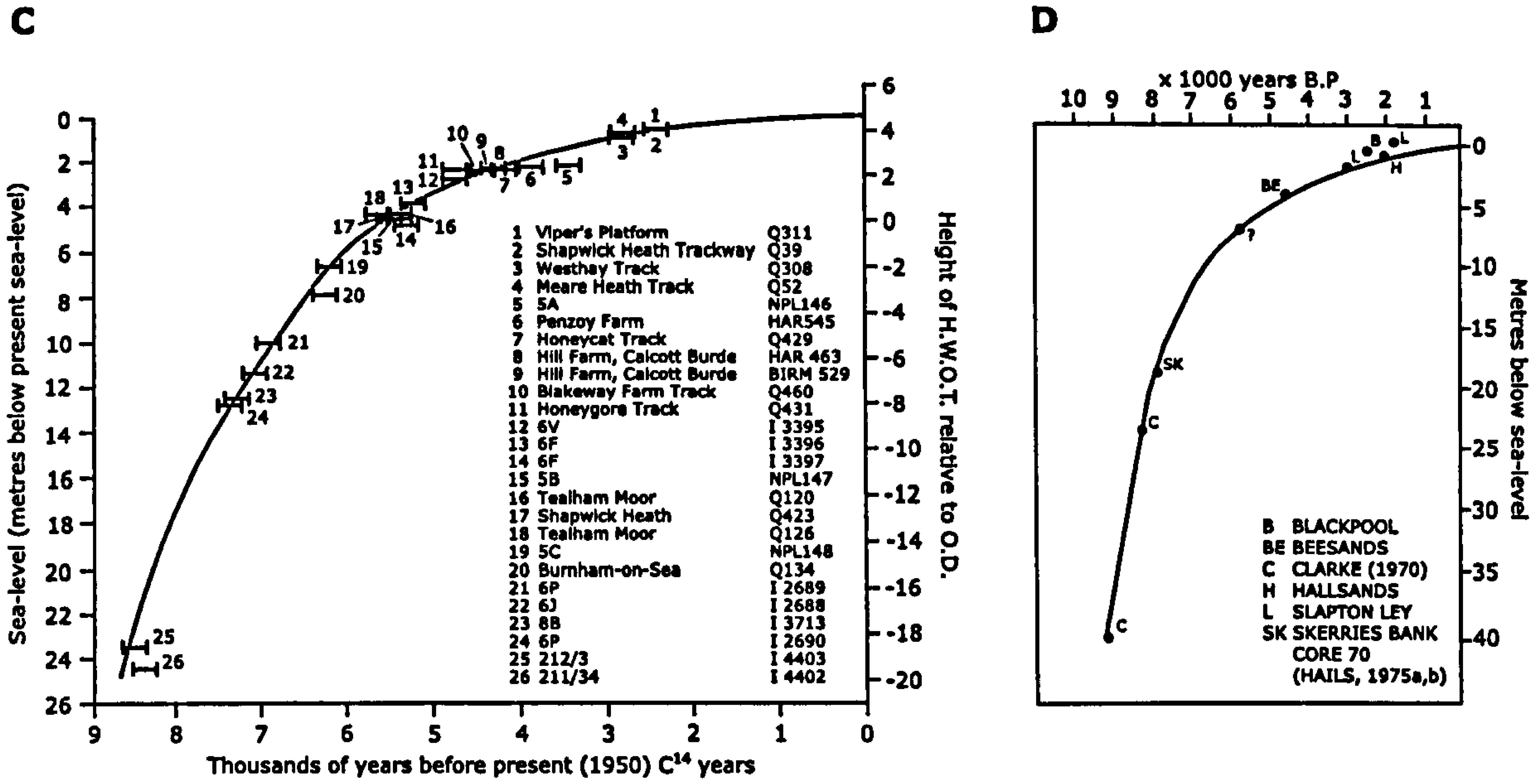


Figure 1.6 continued. **C.** Kidson and Heyworth, 1973, 1976: Bristol Channel; **D.** Morey, 1976.

Sea-level histories constructed from data points questionably linked to the actual freshwater-marine interface are common in the literature of the 1960s to early 1980s (Shennan, 1983; Haslett *et al.*, 1997; Haslett *et al.*, 1998). Sea-level curves were often constructed from index points positioned at some altitude above Highest Astronomical Tide (HAT) and not from a reference tidal elevation. For example, submerged forests (see Hawkins 1971 re. Horner 1816) and ancient peat beds are common along the coastline of the south-west peninsula and have provided researchers with the opportunity to study coastal evolutionary trends and sea-level changes (Churchill, 1965a,b; Kidson and Heyworth, 1978; Heyworth and Kidson, 1982). However, as Haslett *et al.* (1998) have pointed out, data from such horizons may represent the rate of peat accumulation not synonymous with sea level (Kidson and Heyworth, 1976).

An advanced understanding of what constitutes a reliable link between a SLIP and actual sea level has been developed by a number of authors (van de Plassche, 1982; Shennan, 1986b; van de Plassche, 1986; Shennan, *et al.*, 1995; Gehrels, 1999; Gehrels *et al.*, 2001; Edwards, 2001). However, it should be said that previous researchers were generally aware of these limitations. For example, Hawkins (1979) alludes to the problems in dating horizons, considering the effects of tectonism and isostasy, and dealing with sediment consolidation and the relationship of the accumulating environment to sea level. Nevertheless, many Holocene sea-level change studies carried out in south-west Britain between 1960 and the mid-1980s have been done so without the benefit of recent advancements in the science (e.g., Churchill, 1965a,b; Clarke, 1970; Hails, 1975a,b; Hawkins, 1971, 1979; Heyworth and Kidson, 1982; Kidson and Heyworth, 1973, 1976, 1978; Lees, 1975; Morey, 1976, 1983a).

Shennan (1983) identified Kidson and Heyworth (1973, 1976, 1978, 1979), Kidson (1977) and Heyworth and Kidson (1982) as the most important contributions to sea-level change research in the Bristol Channel during the 1970s and early 1980s. Heyworth and Kidson (1982, their Tables 3 and 4) list 142 "tree-ring dates" and 169 "radiocarbon dates" from Wales and south-west England of which 167 ^{14}C dates from "sea-level deposits" and "prehistoric trackways" throughout south-west Britain are used to reconstruct a Holocene sea-level history for the region (Figure 1.7). However, both Shennan (1983) and Haslett *et al.* (1998) agree that the vast majority of their data do not qualify as SLIPs because they are not taken from peat-clay interfaces. Instead they are derived from freshwater peat deposits, in some cases of raised bog origin (Coles, 1982). It should nevertheless be stated that, like Hawkins (1979) before them, Heyworth and Kidson (1982) noted that the intrinsic error to their curves "operates in one direction only, the true sea level always being below that indicated". Haslett *et al.* (1998) point out that this is not the only problem but that the "underestimation of sediment compaction and failure to establish sea-level tendency" further negate the validity of their SLIPs. Furthermore, the assertion that "the generally unsuitable nature of salt-marsh deposits for accurate sea-level determination" (Heyworth and Kidson, 1982) is now known to be false (Gehrels *et al.*, 1996).

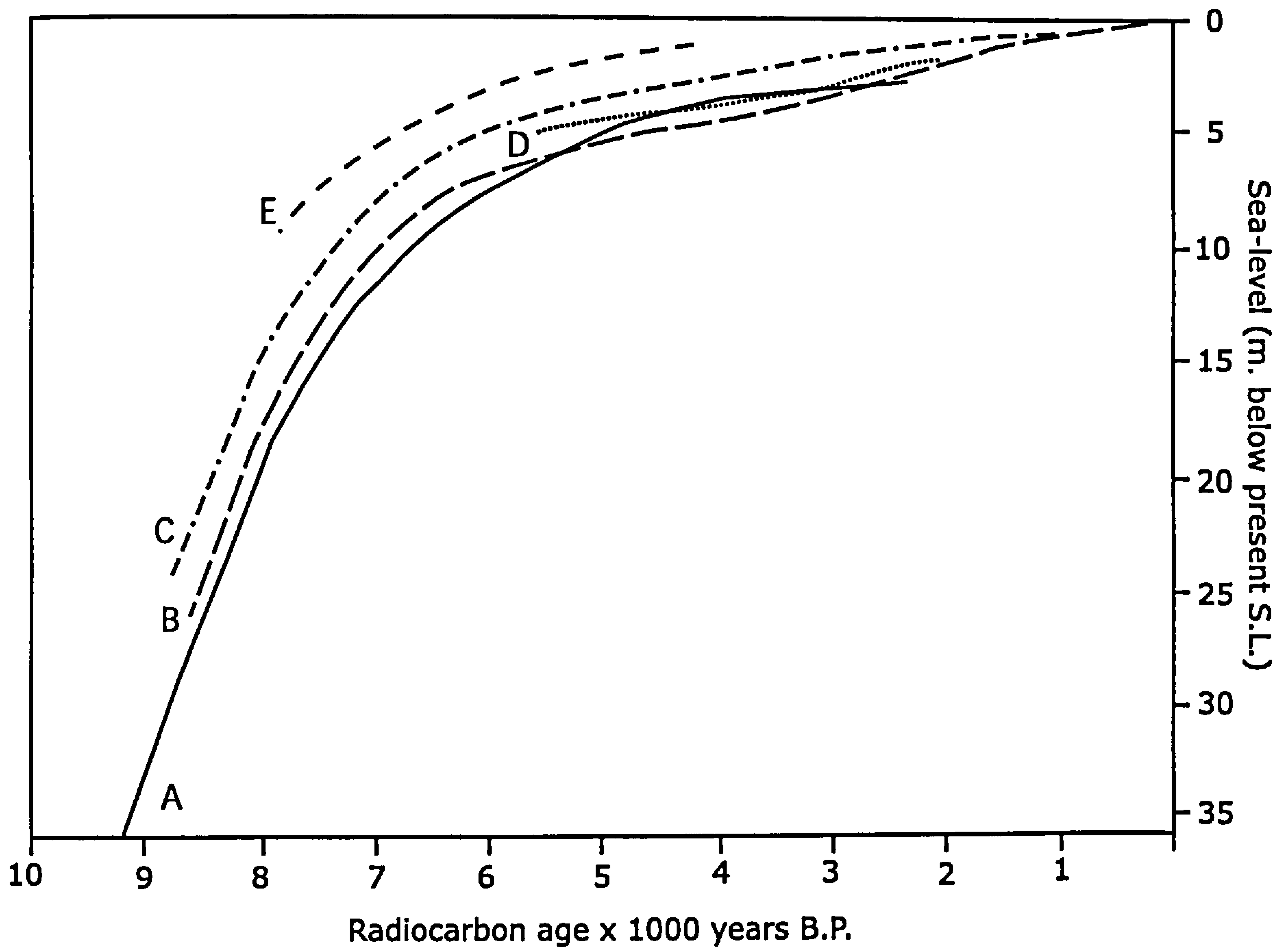


Figure 1.7. Holocene sea level curves from south-west Britain: A. Bristol Channel; B. English Channel (including the Channel Islands); C. Somerset Levels Trackways. D. Cardigan Bay; E. North Wales (re-drawn from Heyworth and Kidson, 1982).

Heyworth and Kidson's (1982) publication was the first collated dataset of SLIPs available for south-west Britain. The RSL curves are similar in trend, steeply rising during the early Holocene with a significant fall in the rate of RSL rise *ca.* 7000 to 6000 cal years BP (Figure 1.7). Although the altitudinal differences between the regional histories can be attributed to differential crustal movement, sediment compaction and tidal range, the authors propose a stable south-west peninsula citing their own data (Kidson and Heyworth, 1973, 1976). They comment on the lack of evidence for Clarke (1970) and Churchill's (1965a,b) early Holocene regional differential isostatic compensation between the Bristol and English Channels. Clarke (1970) suggested that inter-tidal sediments deposited in the early Holocene throughout the Bristol Channel are now some 20 m higher than those identified from the English Channel. Heyworth and Kidson (1982) suggest that this could not have occurred beyond 8000 BP because the rate and timing of sea-level rise is comparable over the whole region. Similarly, Kidson and Heyworth (1973) dismiss the suggestion by Churchill (1965b) that small height differences between peat samples formed around 6500 years ago along the Bristol and English Channel coastlines are not reliable indicators of differential crustal warping. Healy (1995) questions the reliability of their analysis because there is "little conceptual or methodological commonality" in the data from which they construct the regional curves.

Kidson and Heyworth (1973) propose that sedimentation in the Somerset Levels has kept pace with sea-level rise at a height between, what they term "Mean High Water Ordinary Tide (MHWOT)" and MHWST. Heights of prehistoric wooden trackways correlated very closely within this height range and indicated that the inhabitants relocated these features with the changing water table level. Following corrections for gravitational compaction, radiocarbon dates were derived from peat deposits, at the trackway heights, and a regional Holocene sea-level curve was produced (Figure 1.7c). The curve is claimed to be eustatic in nature (Kidson and Heyworth, 1973). They cite Godwin (1943) in support of this argument stating that "valleys in the region would now be drowned were it not for continuous infilling occurring there during the late Holocene". This means that, with no isostatic factor to consider, "almost all peat units in the Somerset Levels are direct indicators of the corresponding water-table" (Kidson and Heyworth, 1973). The tenuous link between the radiocarbon dated samples and their relation with a former sea level is the primary concern with this dataset. The use of current tide tables by the authors to relate ^{14}C samples to a common tidal datum (MHWST) in an effort to reconstruct palaeo-tide cycles is similarly flawed (Healy, 1996).

Heyworth and Kidson's (1982) North Wales curve plots significantly above their sea-level histories for south-west Britain (Figure 1.7e). They conclude that isostatic uplift has significantly influenced the RSL signal from the region during the last 8000 years. However, the 'curve' is based on only two index points from the region. Shennan and Horton (2002) use nine data points, including Heyworth and Kidson's (1982) data, and estimate a 0.3 mm yr^{-1} fall in land level since *ca.* 5000 BP (Figure 1.3). Kidson and Heyworth (1973) suggest their evidence for isostatic and tectonic stability in the south-west peninsula is unaffected by the uplift proposed in North Wales. They further propose that the hinge point for any tilt mechanism in south-west Britain would exist in the Bristol Channel. Shennan and Horton (2002) show that, generally, the north of Britain is uplifting and the south subsiding with a 'hinge point' from Lancashire in the north-west to County Durham in the north-east (Figure 1.3).

Healy's (1993, 1995, 1996) Holocene RSL history from west Cornwall exhibits a similar trend to that of Heyworth and Kidson's (1982) regional curves although the early to mid-Holocene data plot significantly lower (Figure 1.8a,b). Healy's index point 1 is from a Marazion Marsh basal peat resting on the local bedrock but the remaining SLIPs are uncorrected for autocompaction and could explain some of the height differences between Healy's (1995) data and Heyworth and Kidson's (1982) regional curves. Healy (1995) suggests that in addition to the fundamental problems affecting the altitudinal position of the regional curves it is likely that sedimentation processes produced coastal geomorphological features that influenced the relationship between sedimentation and RSL position during the mid-Holocene. This is a reference to the influence that proto-barriers (Morey, 1983a,b) may have had on introducing unique local variations in levels of the coastal water tables of the sites. Jennings *et al.* (1998) investigated further the link between phases of Holocene barrier evolution and changes in the rate of local RSL rise at Porlock in Somerset on the Bristol Channel coast. They agree with the findings of Healy (1993) that barrier morphodynamics can be closely associated with the rate of RSL rise and variations in local sediment supply (see section 1.4.3).

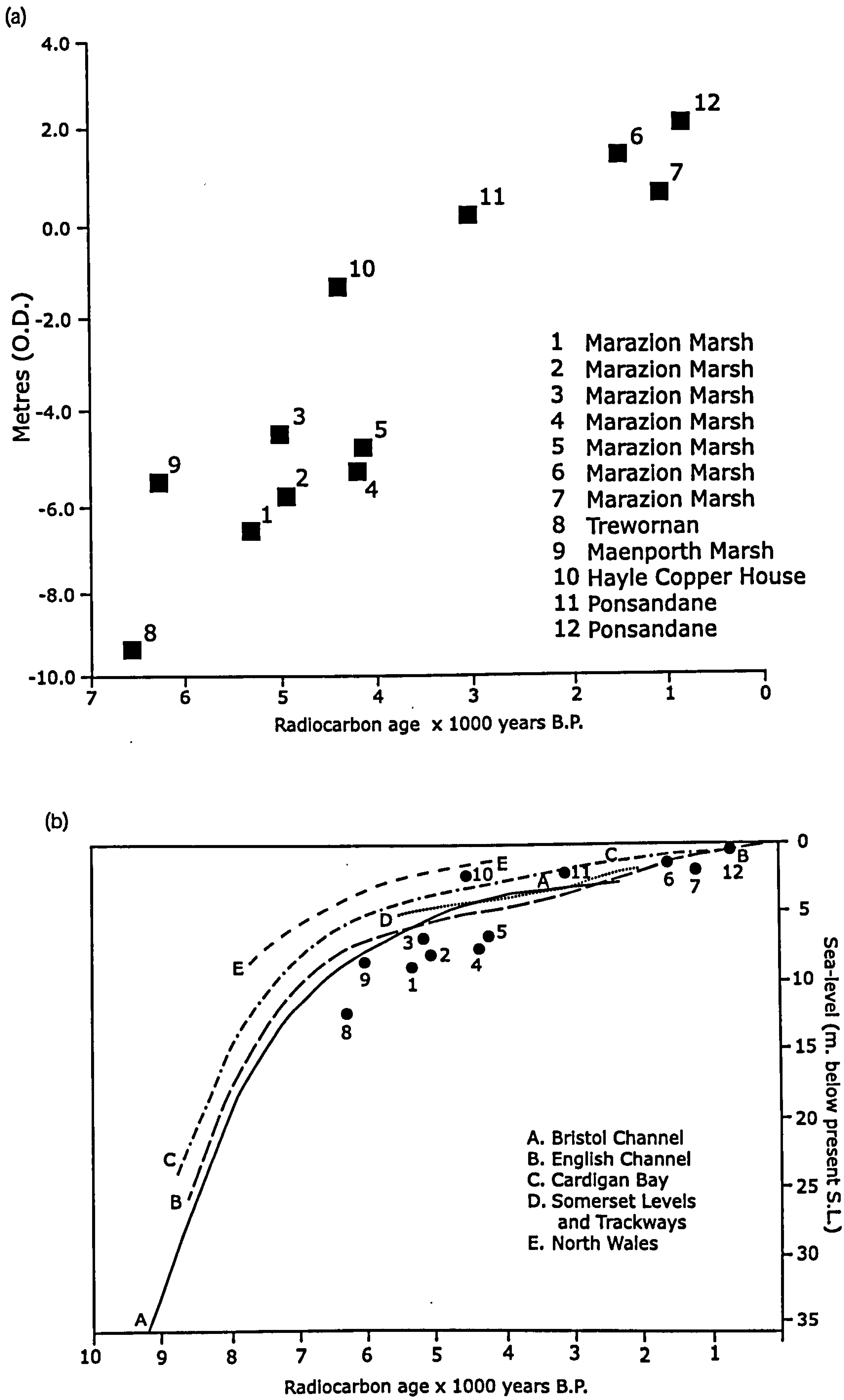


Figure 1.8. Holocene sea-level change in west Cornwall: (a) ^{14}C-based Holocene sea-level history from west Cornwall (after Healy, 1995). (b) altitudinal relationship between Healy's (1993, 1995) and Heyworth and Kidson's (1982) data.

Healy's (1996) lithostratigraphic data reveal similar local sedimentation responses to early Holocene sea-level rise identified by Hails (1975a,b) and Morey (1976, 1983a) in south Devon. Minerogenic sediments and shells overlie an organic-rich basal deposit in cores from Marazion Marsh. The minerogenic deposits are of marine origin and their estimated time of deposition is *ca.* 4500 years ago. Radiocarbon dating of organic-rich layers overlying the marine sediment places the cessation of marine influence at *ca.* 1600 years BP (Healy, 1996). However, the indicative meaning of the fossil record has not been quantified with a local modern analogue and transfer function, and the intercalated samples (e.g., SLIPs 6 and 7, Figure 1.8) are uncorrected for autocompaction. SLIPs are plotted as error ellipses (Healy, 1993) instead of bars or rectangles (e.g., Haslett *et al.*, 1998; Gehrels, 1999). Heyworth and Kidson (1982) discuss the relative merits of using error ellipses or circles in the representation of age and altitude error estimates. They propose the use of ±2 standard deviations to represent the error term of the radiocarbon date. This would mean that, for example, a small dense circle or ellipse would represent a sample with a small error of uncertainty.

Healy (1993) makes other comparisons with his west Cornwall RSL data and that from the literature. Sedimentary sequences from the locality do not contain clear evidence of laterally continuous onlap or offlap sedimentary sequences, such as the intercalated peat beds identified by Shennan (1982) and Shennan *et al.* (1983) in the Fenlands. Similarly, no definable transgressive or regressive contacts, as described by Tooley (1978), Long (1991) and Long and Innes (1992, 1993) for other parts of the UK, can be found in the lithostratigraphy from Cornwall. Healy (1993) concludes that sea-level tendencies alter across diffuse boundaries and not sharp sedimentary contacts, and this may be a site-specific factor related to the amount of available minerogenic sediment and local coastal regimes evolving during the Holocene.

Waller and Long (2003) discuss data from the Isles of Scilly (Ratcliffe and Straker, 1996) in their analyses of Holocene RSL change and coastal evolution in south-west England. Four SLIPs from south Tresco define the trends in RSL during the late Holocene and conform to the RSL history from west Cornwall (Healy, 1993, 1995, 1996) between *ca.* 3000 and 1000 cal years BP (see section 8.1.3 and Figures 8.3a and 8.4a). Waller and Long (2003) suggest that this offers greater credibility to the argument of Ratcliffe and Straker (1996) that the submergence history of the islands was more gradual than previously suggested by Thomas (1985). Furthermore, local tectonism may be discounted as the primary cause of late Holocene RSL rise in the Isles of Scilly. The four late Holocene SLIPs from Ratcliffe and Straker (1996) are not included in the current Holocene SLIP database for south-west Britain. There are only two SLIPs from west Cornwall and three from south Devon *ca.* 3100 – 1100 cal years BP (Figure 1.4 and Table 1.1). The inclusion of the Scilly Isles data by Waller and Long (2003) has therefore highlighted the need for more data from that time period.

Holocene sea-level histories from around south-west Britain generally show a curvilinear pattern of RSL rise with an apparent inflection around *ca.* 7000 to 6000 years ago. Variations in Holocene RSL rates are shown by Heyworth and Kidson (1982, Figure 1.7, curve B) and Jennings *et al.* (1998). The work of Heyworth and Kidson (1982), Healy (1993, 1995, 1996) and others is clearly important in explaining sea-level changes and coastal evolution in the region yet many issues remain unresolved. The accurate interpretation of the lithostratigraphic record and the production of local RSL curves from investigations into the biostratigraphy, including the assessment of autocompaction and the quantification of indicative meaning, are amongst a number of issues in need of clarification (section 1.6). This is especially important when reliable data are required to test rigorously geophysical model predictions of local and regional RSL histories (sections 1.5 and 8.2), and assist in modelling local barrier and back-barrier coastal evolution.

1.4.3 Models of barrier and back-barrier coastal evolution

Coastal sedimentary records have shown that sea-level change is one of the most important processes influencing the evolution of the world's coastlines (Kraft and Chrzastowski, 1985; Carter, 1988). In particular, the development of barrier-lagoon systems has been related to changes in the rate of sea level and sediment supply (Jennings *et al.*, 1998). These systems have world-wide distribution and provide an excellent opportunity for the study of transgressive, stationary and regressive coastlines (Reinson, 1984; Evans *et al.*, 1985; Oertel, 1985). Many of the current models of barrier and back-barrier coastal evolution have been developed along the U.S. Atlantic seaboard (e.g., Kraft, 1971; Belknap and Kraft, 1977, 1981, 1985; Kraft and Chrzastowski, 1985; Niedoroda, 1985; Kraft *et al.*, 1987; Belknap, 1991; Barnhardt *et al.*, 1997). Coastal barrier responses to Holocene sea-level changes around the U.K. coastline (Wilks, 1979; Carter, 1982b; Jennings and Smyth, 1982, 1987, 1990, 1991; Long and Innes, 1995; Long *et al.*, 1998; Evans *et al.*, 2001) include a limited number of studies from south-west Britain (Clarke, 1970; Carr and Blackley, 1973, 1974; Hails, 1975a,b; Morey, 1976, 1980, 1983a,b; Healy, 1993, 1995, 1996). However, very few models of coastal evolution have been proposed for the region (e.g., Morey, 1980, 1983b; Jennings *et al.*, 1998). The model presented by Kraft and Chrzastowski (1985, Figure 1.9) is reviewed here and re-considered, as a further aim and in the light of new findings from this thesis, in section 8.4.

Transgressive coastlines are found at their most developed stages along the Atlantic shores of North America (Kraft *et al.*, 1987). Kraft and Chrzastowski (1985, p.644, Figure 9-10) and Belknap and Kraft (1985, p.256, Figure 12) present a three-stage model showing the landward migration of a lagoon-baymouth barrier at Rehoboth Bay, Delaware. In the model, offshore sand and gravel barriers are driven landward by rising RSL over a coastline with shallow topographic relief. The resultant barrier-island system consists of a variety of sub-environments including a main tidal inlet with flood- and ebb-tidal deltas, inter-tidal marshes and mudflats, and secondary creeks in the back-barrier lagoon (Reinson, 1984). A freshwater lagoon develops following closure of the tidal inlet and continuous deposition in the back-barrier space results in occupation by a freshwater marsh. The model indicates a level of barrier stability during these later stages. However, Jennings and Smyth (1982) suggest that barriers are short-lived features, often breached and reformed at this stage, but Carter (1982b) refutes this stating that they are rarely breached and littoral drift quickly reseals any opening. Furthermore, Kraft (1979) suggests that overwash is the most significant factor influencing sediment transfer across the barrier crest (Morey, 1982b). Eventually, continued net shoreface erosion and local RSL rise transgresses the back-barrier marsh across steeper topographic relief and the barrier is pinned against the highland forming a headland beach (Kraft and Chrzastowski, 1985). Further sea-level rise and net surface erosion results in the coast returning to a barrier-lagoon complex as shallower topographic gradients are transgressed. Sediment extracted by net shoreface erosion is once again deposited offshore or down the coast by longshore drift (Belknap and Kraft, 1981).

The model of Kraft and Chrzastowski (1985, Figure 1.9) ideally illustrates the process of coastal evolution taking place at Slapton Sands today (see chapter 6 and section 8.4). Furthermore, Morey (1982b) suggests that Holocene coastal evolution in Start Bay, south Devon, is similar to the models of Halsey (1979) and Oertel (1979) on the U.S. Atlantic coast. There are also similarities with Long and Innes (1995) who proposed a three-phase model of gravel barrier development on the Dungeness Foreland in Kent, south-east England, comprising initiation, stability and breakdown. They suggest that barrier initiation began *ca.* 6000 – 5000 years BP with gravel supplied from offshore (Jennings and Smyth, 1991), longshore (Nicholls, 1991) or a combination of these and other local sources. The accumulation of back-barrier peat from *ca.* 5000 – 2000 years BP characterises the established phase, and barrier breakdown in the late Holocene is inferred by washover and channel facies found in the marsh stratigraphic sequences (Long and Innes, 1995). It has been proposed that the current phase of barrier stability at Slapton Sands has been punctuated on only two occasions by significant washover events, the latest being *ca.* 1800 years BP (Morey, 1982b).

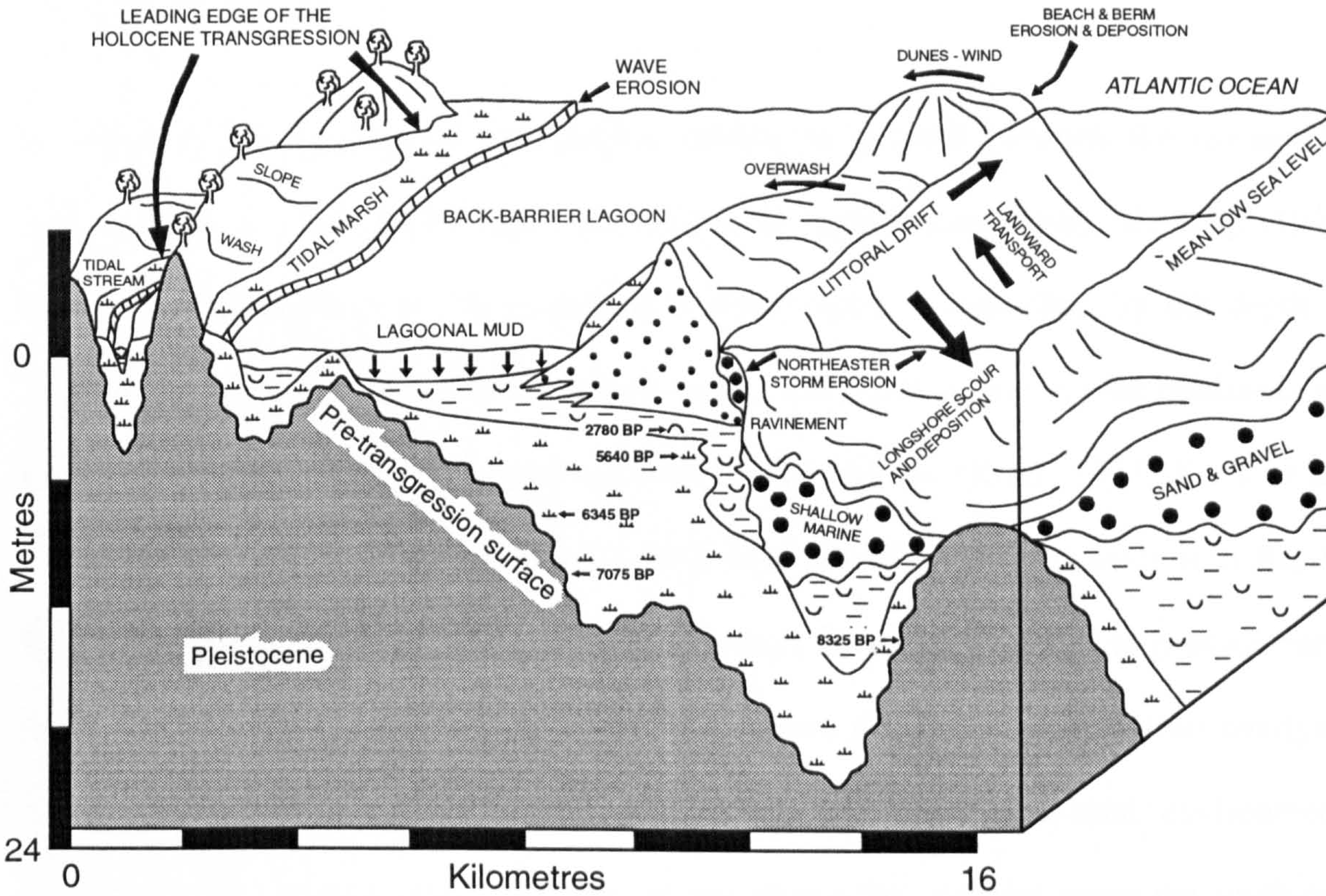


Figure 1.9. A transgressive lagoon-barrier coastline by Kraft and Chrzastowski (1985). Modelled on Rehoboth Bay lagoon on the Atlantic coast of Delaware. Radiocarbon dates are from shells, wood fragments and basal peat.

Kraft and Chrzastowski (1985) suggest that barrier and back-barrier cores examined from Rehoboth Bay show that the typical transgressive sequence, of the type described in Walthers Law (1894, in Middleton, 1973), can be modified by lateral migration of tidal inlets along the barrier. This is likely to have occurred at sites around south Devon during the Holocene. In Morey's (1982b) model the lagoons and estuaries of Start Bay, south Devon, were flushed clear by tidal scour during the early Holocene. The ravinement unconformity formed under rising sea level and barrier-lagoon systems transgressed the Pleistocene surface (Hails, 1975a,b) reworking older sediments in relatively shallow water conditions (Clarke, 1970). As RSL rates lowered during the mid-Holocene the continuous deposition of back-barrier sediments from freshwater sources and littoral drift sealed the barriers and the frequency of barrier over-wash reduced (Morey, 1982b). Back-barrier sediments examined from Slapton Ley infer that closure of the barrier occurred *ca.* 3000 years BP (Morey, 1976, see chapter 6).

In Figure 1.9 the preserved stratigraphic section is situated between the ravinement unconformity and the pre-transgression surface or basal unconformity. Belknap (1991) states that "preservation of the back-barrier stratigraphy is controlled by the depth of palaeovalleys". In the Delaware bay-mouth barrier and headland system the headlands are former interfluves, while the lagoons occupy palaeovalleys (Kraft, 1971; Kraft *et al.*, 1979). Similar deposits, up to 28m thick, have been identified in cores from Start Bay by Hails (1975a,b) and at least six pre-Holocene valleys have been dammed to produce back-barrier lagoons (Morey, 1983a). Lees (1975) has shown that Holocene sediments overlying the pre-transgression surface contain salt marsh and other inter-tidal environments preserved in the basal section. This review has shown that systems along the south-east Devon coastline described by Hails (1975a,b), Lees (1975) and Morey (1976, 1980, 1983a,b) are remarkably similar to that of the Delaware transgressive barrier-lagoon coastline. This thesis will apply the model of coastal evolution by Kraft and Chrzastowski (1985) to the south Devon coastline (section 8.4) on the basis of findings in chapters 4 to 7.

1.5 Geophysical modelling

Geophysical models used in this project predict the effect of glacio-isostasy on sea-level change along the Channel coastline of south-west England. Daly (1934) and Walcott (1970) presented two conceptualised versions of current geophysical or 'Earth' models (e.g., Lambeck, 1990, 1991, 1993a,b, 1995, 1997; Lambeck *et al.*, 1996; Peltier, 1991, 1994, 1996, 1998) based on the 'visco-elastic' and 'plastic' properties of the mantle rheology (Figure 1.10a,b). The schematic model of a visco-elastic Earth shows that flexing of the lithosphere occurs at the ice margin. This flexing is termed a 'forebulge' by Earth modellers and is constrained by parameters such as ice thickness, ice mass, areal extent of the ice sheet, visco-elastic properties of the mantle and lithospheric thickness at the ice margin.

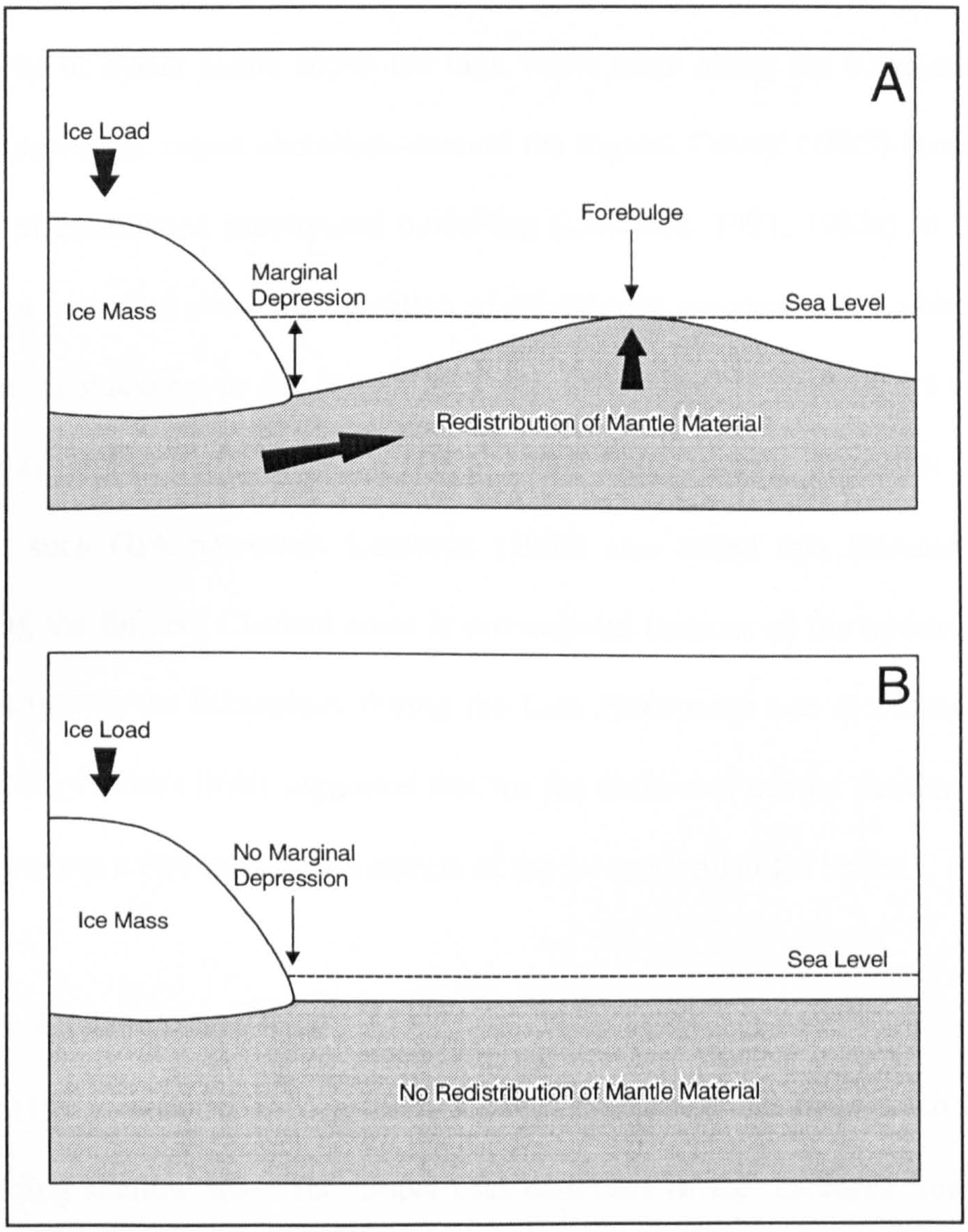


Figure 1.10. Earth modelling by Walcott (1970). Schematic illustration of (A) a 'visco-elastic' Earth showing the forebulge effect beyond the ice margin, and (B) a 'plastic' Earth in which the ice load effect terminates at the ice margin.

Wingfield (1995) suggested that a forebulge migrated northwards as the ice sheet retreated producing north-south aligned coasts west of Brittany and later across the English Channel entrance. Forebulge passage into the northernmost Celtic Sea then produced a west-east aligned coast south of Ireland. Carter's (1982a) pattern of isostatic behaviour in southern Ireland suggests that the land surface has remained relatively stable during the late- and post-glacial. However, Devoy (1983) dismisses this citing a topographical feature identified by Synge (1981) in south-west Ireland, approximately 8500 km^2 in size and probably produced by the growth of an ice-cap, that is difficult to explain without invoking a glacio-isostatic adjustment of the land surface. Low beach ridges in Bantry Bay and accumulations of oyster shells above the high water mark along the Kenmare River are cited as evidence for raised shorelines around the region. Devoy (1995) states that "the application of established geophysical modelling (Lambeck, 1991, 1993a) to the regional data does not appear to show the operation of a forebulge mechanism". Lambeck appears to infer that ice thickness in the British Isles was insufficient to produce a forebulge and Peltier (1998) implies that Lambeck's (1995, 1997) Earth (mantle viscosity) model does not support such GIA processes. Lambeck (1995) also stated that Holocene sea-level change along the English Channel coast is not uniform because of the hydro-isostatically induced changes in the lithosphere during the Late Pleistocene and Holocene period. In contrast to this, Peltier (1998) suggested that the ice mass over central Scotland was thick enough to produce a forebulge at the margin of the former British Ice Sheet.

1.5.1 Lambeck's geophysical model

Lambeck's (1995) model (GB-3) predicts ice retreat across Great Britain and Ireland and the surrounding shallow seas. The model uses estimates of ice thickness from trim-lines and erratics (the formation of nunataks during the Last Glacial Maximum) on some of the higher mountain summits. Empirical relations between ice heights and horizontal ice-sheet dimensions are also included (Lambeck, 1993b). A maximum ice thickness of 1500 m over central Scotland and approximately 500 m over Ireland is estimated. Rapid ice retreat is postulated over the southern Irish Sea resulting in ice-free conditions at the Isle of Man *ca.* 18,000 BP (Lambeck, 1995). In GB-3 the completion of ice retreat occurs by 12,500 BP, but in an earlier model (GB-3a) ice retreat is completed by 13,000 BP. GB-3a predicts only minor differences in sea-level changes due to the small ice volumes at this time.

GB-3 is defined spatially by a series of grids (25 km^2) superimposed on Great Britain and Ireland. Glacial cycles are constructed at time intervals of 1000 years from 22,000 years BP, the timing of maximum glaciation (Huddart *et al.*, 1977; Bowen and Sykes, 1988; Eyles and McCabe, 1989), to 14,000 years BP. The model is then defined by time steps of 500 years into the Holocene period (Lambeck, 1995). Lambeck (1991) suggests that the centre of post-glacial rebound in Scotland would have been much further east had there been a substantial North Sea ice load. The timing of ice-sheet movements during the past 20,000 years is given in the conventional radiocarbon time scale and is also the chronology used for RSL predictions. Lambeck (1995) interpolated SLIPs at 500- or 1000-year intervals and RSL histories are produced for sites around the British coastline. The reliability of modelled predictions of RSL histories is dependent on the density of empirical data. Lambeck (1995) suggests "this grouping (leading to the production of a single sea-level curve) has only been possible for locations beyond the limits of the former ice sheet margin such as southern England".

Lambeck's (1995) model uses histories of the Fennoscandian and Laurentian ice sheets with contributions from distant ice sheets, as well as northern Britain. He proposes two main factors causing sea-level change around the British Isles: (i) post-glacial crustal rebound and (ii) hydro-isostatic contributions from the surrounding shallow seas and the Atlantic Ocean. The glacio-hydro-isostatic effect is a constraint on the model parameters and contributes to the estimation of lithospheric thickness, mantle viscosity, and ice sheet characteristics. The inclusion of the rebound and sea-level change models for the British Isles rule out any likelihood of them being treated in isolation of the other ice sheets. As a result of this inclusion Lambeck (1995) suggested that British ice model thickness is too high by some 10%. Ice thickness over the Scottish Highlands, at a maximum of 1300 – 1400 m, would therefore be insufficient for a forebulge (Lambeck, 1995, 1997).

Observational information used in the model predictions of sea-level heights is subject to uncertainties. Three significant areas where errors can be found are (i) radiocarbon ages of samples, (ii) reductions of peat heights to mean sea level (MSL), and (iii) the problem of autocompaction of peat and sedimentary units. Lambeck (1995, 1997) includes radiocarbon age errors when comparing observed with predicted RSL histories. Lambeck's RSL predictions for south Devon (Figure 1.2) are converted to calendar ages, unless otherwise stated, for comparison with data from Peltier and this project. This could produce errors because of the non-linearity of the timescale with which the data was run (see section 1.5.3). Reducing the peat heights to mean sea-level is achieved by: (i) using (MHWS-MHW)/2 for brackish-water peat units, and (ii) ascribing a value of 1 m for the accuracy of the tidal reduction when $t < 6000$ years BP, and 2 m for earlier epochs. These methods apply to Lambeck's (1997) work on the French Atlantic and Channel coasts. Compaction errors are assumed to be 1 m, which leads to the possibility that SLIPs may plot below their true level ignoring all other factors (Lambeck, 1997).

1.5.2 Peltier's geophysical model

The ICE-1 model of Peltier and Andrews (1976), and the slightly modified ICE-2 model of Wu and Peltier (1983), described the deglaciation history of the Earth's surface based on geomorphological and ^{14}C evidence. Peltier's (1991) ICE-3G model of late Pleistocene deglaciation employs 192 ^{14}C controlled time series of RSL change for ice covered sites from 21,000 BP (the Late Glacial Maximum, LGM). Sea-level predictions are tested with those from 200 RSL series of non-ice-covered sites (Peltier, 1991). Adjustments are then made to the theoretical model of post-glacial rebound, visco-elastic properties of the Earth's crust and mantle, and the altitudinal position and timing of RSL curves. Other adjustments are made to the spatial and temporal pattern of ice retreat i.e., melting history (eustasy) from horizontal (moraines) and vertical (ice thickness) estimates.

Peltier (1994) reconstructed the palaeotopography of glacial-age ice sheets using his refined ICE-4G model and proposed that the LGM ice volume was much lower, by some 35%, than in previous models. In particular, the Fennoscandian ice sheet thickness predictions from the CLIMAP maximum reconstruction (CLIMAP Project Members, 1976) were 1.5 km more than those predicted with the ICE-4G model. The thinner ice sheet predictions of ICE-4G consequently produced a lower eustatic RSL rise of 105.2 m compared to the CLIMAP maximum predictions of 163 m. The differences in ice sheet topographic height between the two glacial reconstructions clearly have significant effects on predictions of post-glacial sea-level change.

Peltier (1998) tested his ICE-4G (VM2) model of post-glacial sea level history on the British Isles. He refers to Great Britain as "probably the most exotic location on Earth from the perspective of glacio-isostatic adjustment (GIA) as it is not only immediately peripheral to the large ice sheet that covered Scandinavia, and therefore subject to the process of '*proglacial forebulge*' collapse, but it was itself also glaciated in the north, with the thickest ice cover found over the Scottish Highlands" (Peltier, 1998, p.609). Refinements to ICE-3G were primarily the average mantle viscosity profiles inferred on the basis of the GIA data (Peltier, 1996). Fennoscandian GIA proceeds faster with his ICE-4G VM2-based model, which decelerates RSL fall for present-day predictions.

Peltier (1998) reconstructed the coastal evolution of north-west Europe centred on the British Isles using the ICE-4G model parameters. Sites in south-west Britain from which ^{14}C dated RSL histories were obtained, include the Bristol Channel and the south coast of Devon and covers most of the Holocene. He suggests that rising sea level along the

Channel coast of south-west Britain is controlled by the same process of forebulge collapse observed along the U.S. eastern seaboard (Barnhardt *et al.*, 1995). The spatial extent and collapse of the forebulge is governed by the deglaciation histories of the Scottish and Fennoscandian Ice Sheets (Peltier, 1998). The model constructed by Peltier in the late 1980s to early 1990s has been constantly refined with respect to its various parameters (Peltier *et al.*, 2002; Shennan *et al.*, 2002). The main refinements are primarily mantle viscosity profiles, the weight and thickness of continental ice sheets that covered the Earth's surface at the last glacial maximum, and additions to ^{14}C controlled RSL histories. The latest adjustments include the "broad shelf" effect (Peltier and Drummond, 2002), prehistory and the rotational effect (Peltier, 2002a,b; Peltier *et al.*, 2002, see section 8.2).

1.5.3 Testing the models with published data from south Devon

Modifications to the deglaciation histories and mantle viscosity parameters produce significantly different RSL predictions for south Devon in the case of Lambeck's (1993a,b, 1995) GB-3 model revisions and less so for Peltier's (1998) ICE-4G (VM2) revisions (Figure 1.2, see section 1.6.1). Models overestimate all of the south Devon data post-6000 cal years BP (Figure 1.11a) and comparisons with Lambeck's model predictions using the ^{14}C ages show a similar misfit (Figure 1.11b). Lambeck's models diverge significantly during the early Holocene. For example, Lambeck (2000) lowers his earlier (Lambeck, 1997) RSL predictions *ca.* 7800 cal years BP by more than 5 metres (Figure 1.11a). The provision of early Holocene basal SLIPs from south Devon in this project should help to resolve the model differences (see Figure 8.9). It is clear that the critical time period within which to test all the models is *ca.* 7000 to 6000 cal years BP where they differ by as much as 7 metres (Figures 1.2 and 1.11a,b). This may be due to differences in the models eustatic input, i.e., ice melting histories, and the effect that the GIA process has on RSL history at this time, e.g., Peltier's model predicts high rates of subsidence due to forebulge collapse.

This is discussed further in section 1.6.1 and data from this project will test the models early to mid-Holocene RSL predictions in section 8.2.

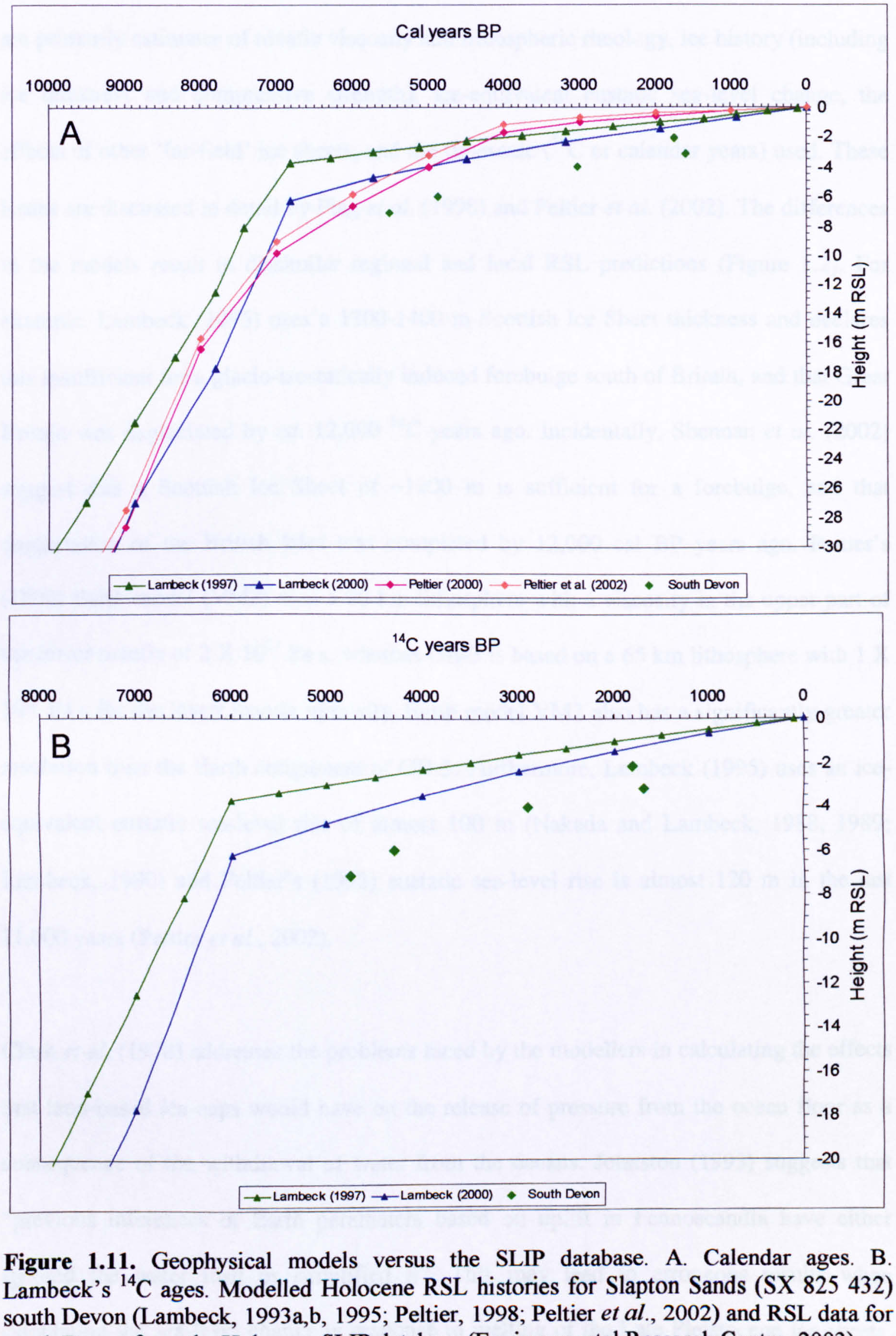


Figure 1.11. Geophysical models versus the SLIP database. A. Calendar ages. B. Lambeck's ^{14}C ages. Modelled Holocene RSL histories for Slapton Sands (SX 825 432) south Devon (Lambeck, 1993a,b, 1995; Peltier, 1998; Peltier *et al.*, 2002) and RSL data for south Devon from the Holocene SLIP database (Environmental Research Centre, 2003).

1.6 Unresolved issues

1.6.1 Model differences and limitations

The differences between ICE-4G (VM2) (Peltier, 1998) and GB-3 (Lambeck, 1995, 1997) are primarily estimates of mantle viscosity and lithospheric rheology, ice history (including ice thickness and compressive strength), ice-equivalent eustatic sea-level change, the effects of other 'far-field' ice sheets, and the timescale (^{14}C or calendar years) used. These issues are discussed in detail by Plag *et al.* (1998) and Peltier *et al.* (2002). The differences in the models result in dissimilar regional and local RSL predictions (Figure 1.2). For example, Lambeck (1995) uses a 1300-1400 m Scottish Ice Sheet thickness and declares this insufficient for a glacio-isostatically induced forebulge south of Britain, and that Great Britain was deglaciated by *ca.* 12,000 ^{14}C years ago. Incidentally, Shennan *et al.* (2002) suggest that a Scottish Ice Sheet of ~1200 m is sufficient for a forebulge, and that deglaciation of the British Isles was completed by 12,000 cal BP years ago. Peltier's (1998) Earth model (VM2) uses a 90 km lithosphere with a viscosity in the upper part of the lower mantle of 2×10^{21} Pa s, whereas GB-3 is based on a 65 km lithosphere with 1×10^{22} Pa s for the lower mantle viscosity. Earth model VM2 also has a significantly greater resolution than the Earth component of GB-3. Furthermore, Lambeck (1995) uses an ice-equivalent eustatic sea-level rise of almost 100 m (Nakada and Lambeck, 1988, 1989; Lambeck, 1990) and Peltier's (1998) eustatic sea-level rise is almost 120 m in the last 21,000 years (Peltier *et al.*, 2002).

Clark *et al.* (1978) addresses the problems faced by the modellers in calculating the effects that land-based ice-caps would have on the release of pressure from the ocean floor as a consequence of the withdrawal of water from the oceans. Johnston (1993) suggests that "previous inferences of Earth parameters based on uplift in Fennoscandia have either ignored the water load or simplified it". This may lead to erroneous results when calculating the sea-level change in response to melting of the Late Pleistocene ice sheets.

Lambeck (1995) explains that Holocene sea-level change along the English Channel coast is not uniform because of the crustal response to the changing ice (glacio-isostasy) and water loads (hydro-isostasy) in the Late Pleistocene and Holocene time. According to Peltier (1998), south-west Britain's RSL history is governed by the process of forebulge collapse due to the deglaciation of the Scottish and Fennoscandian ice-sheets. Faster rates of empirical RSL rise than those predicted by Lambeck's models may suggest that a collapsing forebulge mechanism existed off Britain's south-west coastline.

The RSL history of south-west Britain may be affected by hydro-isostasy in primarily two ways. Firstly, the melting of thicker (than predicted) land-based ice would result in greater amounts of water being added to the oceans. This will in turn cause greater depression of the ocean floor and newly flooded continental margins. Johnston (1993) suggests that, sea-level rise relative to the solid Earth near to the ice sheet (e.g., southern Britain) would be less dramatic due to the increased post-glacial rebound of the surface and greater depression of the sea floor. Secondly, the increased pressure on the ocean floor is likely to push mantle material towards the margins of the ocean basins. These changes in Earth rheology are likely to have an effect on forebulge activity and south-west Britain's RSL history. The collection of reliable SLIP data will improve the predictive powers of the models and help to resolve issues of glacio- and hydro-isostasy, in particular forebulge activity.

1.6.2 Radiocarbon dating

The accurate dating of core samples is crucial in determining reliable rates of sea-level change over time (Heyworth and Kidson, 1982; Gehrels and Belknap, 1993; Healy, 1996; Gehrels *et al.*, 1996; Törnqvist and van Ree, 1998). The choice between Accelerated Mass Spectrometry (AMS) dating and the conventional dating method is important on two main fronts: 1) the ability to target smaller samples, and 2) the identification of a more precise average age through the analysis of individual plant fossil remains. Uncertainties in the age and altitude of a radiocarbon dated sample result in SLIP boxes with considerable horizontal and vertical errors (Törnqvist and van Ree, 1998).

Radiocarbon dating errors can result from the contamination of peat units during the sampling stage (Gehrels *et al.*, 1996). Natural contaminants include those from nearby lake waters, unstable soils and in-wash of local soils and bedrock (Lowe and Walker, 1980; Lowe *et al.*, 1988; Törnqvist *et al.*, 1992). For example, material to be dated may have been transported from elsewhere and may be older than the horizon in which it is found. Some minerogenic facies may also contain very little carbon (Olsson, 1979). Problems at the laboratory stage include prolonged handling and loss of carbon during treatment (Olsson, 1986). Mislabelling could result from the processing of large numbers of samples. Some of these problems may be difficult to trace and could be lost in the analytical stages of the investigation (Gehrels *et al.*, 1996). The AMS ^{14}C dating of individual plant fragments from the centre of core samples can reduce some of the natural and field problems (Törnqvist and van Ree, 1998), and careful handling at the laboratory stage will lessen problems there (Gehrels *et al.*, 1996).

Previous authors working in south-west and southern Britain have calibrated their radiocarbon ages (e.g., Healy, 1993, 1995, 1996; Haslett *et al.*, 1997, 1998; Long *et al.*, 1999; Edwards, 2001). Ages of SLIPs from authors of earlier works (e.g., Churchill, 1965a,b; Hails, 1975a,b; Morey 1976, 1983a) have been calibrated by others (Environmental Research Centre, 2003). The calibration programmes of Stuiver and Pearson (1986), Stuiver *et al.* (1986) or Stuiver and Reimer (1993) have been routinely used but these earlier versions are now updated by Stuiver *et al.* (1998a,b). Calibrating ^{14}C ages resolves the remaining two problems, namely: (1) the known 3% error of the ^{14}C half life (conventionally 5568 yr) and (2) the varying amounts of ^{14}C isotopes in the Earth's atmosphere through time (Gehrels *et al.*, 1996).

The AMS ^{14}C dating of sample material used for SLIPs can improve existing chronologies and strict sampling methods should be employed to reduce contamination and error margins. These methods include the extraction (where possible) of horizontal individual plant fragments from selected horizons. AMS ^{14}C ages in this project are calibrated using Stuiver *et al.* (1998a).

1.6.3 Autocompaction of sediments

There are currently only a few geotechnical models that have been used to estimate the autocompaction of sediments (e.g., Paul and Barras, 1998; Rybcyk *et al.*, 1998; Allen, 1999), an issue that is routinely ignored by sea-level researchers. This is especially true for samples collected within thick unconsolidated sequences of inter-tidal sediments (Allen, 1995, 1999, 2000; Shennan and Horton, 2002). SLIPs obtained from basal peat directly overlying hard substrate are compaction-free and, as the basal sediment is not vertically displaced over time, are likely to improve the reliability of RSL histories (Kiden, 1995; Gehrels, 1999). Many of the SLIPs published by researchers in south-west England (Churchill, 1965a,b; Hails, 1975a,b; Morey, 1976, 1983a; Heyworth and Kidson, 1982; Healy, 1993, 1995, 1996) are from non-basal sediments and are likely to have been subjected to some autocompaction.

Attempts have been made to deal with the post-depositional compression of a dated sample in the stratigraphic sequence, and its relationship to contemporary sea level (Godwin *et al.*, 1958; Heyworth and Kidson, 1982). However, Haslett *et al.* (1998) suggest that these earlier estimates largely underestimate the effect of sediment consolidation on their samples, resulting in sea-level histories plotting lower than at the time of deposition. Paul and Barras (1998) state that "elevation errors are especially significant when studying Flandrian (Holocene) sea-level change, because the magnitude of the change can be similar to the magnitude of the errors". Allen (1999) also infers that vertical errors are likely to be much greater when working with Holocene peat sequences as the material is extremely variable and does not conform to standard laboratory tests, e.g., Atterberg (1911) limits. Kidson and Heyworth (1973), Devoy (1982) and Smith (1985) encountered similar problems in their attempts to use sediment mechanics theory to decompact Holocene peat sequences.

Mathematically quantifying the problem of autocompaction in Holocene sequences is always unlikely to deal with issues of spatial variability across the former marsh or mudflat surface. Kaye and Barghoorn's (1964) geometrical measure of peat autocompaction, using deformation of logs, was limited by 'slip' between timber in the sequence and the surrounding finer peat material. Allen (1999) states that similar geometrical approaches are limited by the availability of borehole and chronological data and assumptions that some peat bed tops may be isochronous (Bloom, 1964; Belknap and Kraft, 1977; Allen, 1996; Haslett *et al.*, 1998). The use of geotechnical tests and equations that resolve the post-depositional consolidation of Holocene sediments is currently the most widely used methodological approach to the problem (Paul *et al.*, 1995; Pizzuto and Schwendt, 1997; Paul and Barras, 1998).

1.6.4 Indicative meaning

The quantification of indicative meaning (the height at which a sea-level indicator was deposited in sediment in relation to a reference tide level) is necessary to reconstruct reliable palaeo-water levels (van de Plassche, 1986). Previous studies in south-west Britain have not accurately determined the indicative meaning of radiocarbon dated samples used to reconstruct Holocene sea-levels (e.g., Churchill, 1965a,b; Hawkins 1971, 1979; Kidson and Heyworth, 1973, 1976, 1978; Hails, 1975a,b; Morey, 1976, 1983a; Heyworth and Kidson, 1982). Methods of obtaining SLIPs include estimates from pollen zones, archaeological remains, dendrochronologies, and plant fossils. Haslett *et al.* (1997) suggest that these authors have not assigned indicative meaning to their SLIPs and thus cannot be considered reliable interpretations of former sea levels. However, many index points from these earlier works in south-west Britain are now listed in the Durham SLIP database, and have been assigned vertical error estimates on the basis of the available sedimentological, micropalaeontological, palynological, field survey and tidal information.

Many recent studies use foraminifera as sea-level indicators (e.g., Thomas and Varekamp, 1991; Gehrels, 1999; Gehrels and van de Plassche, 1999; Horton, 1999; Horton *et al.*, 1999a,b; Edwards and Horton, 2000; Gehrels, 2000; Horton *et al.*, 2000; Edwards, 2001). The vertical distribution of these animals is controlled by elevation, flooding frequency and salinity in the inter-tidal zone (Gehrels, 2000). Whilst foraminifera are often mooted as reliable sea-level indicators (Scott and Medioli, 1986), research has shown that using multiple indicators would provide a more accurate and precise reconstruction of palaeo-tidal levels (Gehrels *et al.*, 2001). Diatoms (Shennan *et al.*, 1994; Zong, 1997; Zong and Horton, 1999), testate amoebae (Charman *et al.*, 1998), pollen (Long *et al.*, 1999) and plant macrofossils (Belknap *et al.*, 1989) are typical of such proxies found at specific heights within the tidal frame. Healy (1993, 1995, 1996) is the only author in south-west Britain to have attempted a reconstruction of Holocene sea-levels using diatoms and pollen.

The indicative meaning of fossil salt-marsh sediment in this project is quantified using a transfer function based on the vertical distribution of local modern foraminiferal assemblages (see chapter 3). The vertical position of a SLIP in a time-height diagram is calculated from its height relative to MTL (H), its indicative meaning (I) and the amount of autocompaction (A) (see Gehrels, 1999):

$$SLIP = H - I + A$$

No adjustment is made for changes in palaeo-tidal range in this project. However, Austin (1991, p.281, Figure 4a,b) has modelled the tidal range in the English Channel throughout the early Holocene and indicates that the M_2 tidal amplitude south of Devon was around 0.3 – 0.9 m higher *ca.* 9000 – 8000 cal BP years ago. This reduced to 0.3 – 0.5 m higher than today *ca.* 7500 cal years BP. A further reduction in tidal energy during the mid- to late Holocene is predicted off south Cornwall and Devon by the models as tidal range increased in the eastern Channel following the opening of the Straits of Dover (Austin, 1991). Gehrels (1999) states that the indicative meaning of a SLIP derived from the upper part of the inter-tidal zone may only reflect current MTL if the tidal range has remained unchanged through time. This is an important consideration when evaluating vertical errors of SLIPs. However, although this project could benefit further from modelling Holocene tides along the south Devon coastline, Austin's (1991) results indicate that changes in palaeo-tidal range have been relatively small, especially for the middle and late Holocene. Regional corrections for palaeo-tidal range changes would therefore not significantly alter the results of this thesis.

1.6.5 Surveying errors

The reliability of any SLIP depends first on the accuracy of the initial site survey. Errors affecting the measured altitude of stratigraphic boundaries may occur at three stages of an investigation: 1) depth measurement in a borehole; 2) levelling of the site to an Ordnance Survey (OS) benchmark; and 3) the assessment of the benchmark's accuracy to OD Newlyn (Shennan, 1986b). The accuracy of the measured depth from ground level to each core section is dependent on the drilling technique and the user. Sampling density and local stratigraphic surface roughness may also affect the accuracy of the measured altitude of a stratigraphic boundary (Makarovič, 1973). The reliability of an OS benchmark, and the levelling accuracy to that benchmark, largely depends on the system order used in relation to OD Newlyn. For example, a third order benchmark may have potentially greater vertical error than a second order type which, in turn, may be less accurate than a first order benchmark.

Shennan (1986b) estimated local comparisons of benchmarks to be accurate to within ±0.01 m relative to OD Newlyn, and inter-regional comparisons at ±0.15 m for England and Wales. However, the Ordnance Survey suggest significantly lower error estimates dependent on the order of the benchmark used (Table 1.3). These figures were estimated during the 1970s and land subsidence around coastal sites may have since increased the maximum error considerably. This can result from natural or unnatural processes, e.g., on-going post-glacial isostatic adjustment or sediment extraction in the coastal zone respectively. Steps to improve the reliability of site survey data should therefore include the following: 1) careful and consistent depth measurement during the drilling process; 2) an electronic distance measurement (EDM survey) technique; and 3) a prudent choice of benchmark (see section 2.2.2). The Ordnance Survey was consulted to account for possible errors associated with benchmarks used in this study and section 2.2.2 explains the steps taken to limit any errors during the survey.

Order of benchmark	Maximum error (mm)
Fundamental	Error free
Geodetic	±2.0
Secondary	±5.0
Tertiary	±12.0

Table 1.3. Accuracy of Ordnance Survey benchmark levelling. Source: Ordnance Survey (2003).

1.6.6 Selection of a local tidal datum and changes in tidal range

Coastlines with high tidal ranges may induce greater errors in the calculation of indicative meaning than those with low tidal ranges (Heyworth and Kidson, 1982). Higher tidal ranges amplify water level changes and thus increase the difficulties in relating water table level to sea level. Much of the south Devon coastline is subject to spring tidal ranges around 4.5 metres. However, tidal range and the height of MHWST may change significantly further up an estuary thus hampering the interpretation of indicative meaning. It may therefore be difficult to quantify an altitude, in relation to MTL, of samples taken from up-estuary sites because of probable differences in tidal heights between the open coast and the 'inland' site (Haslett *et al.*, 1998).

Palaeo-tidal changes during the Holocene are likely to induce further error in the assessment of the altitudinal reference of a SLIP. Previous authors have modelled tidal changes (Austin, 1991; Gehrels *et al.*, 1995), in some cases to reduce vertical errors when reconstructing RSL histories from coastal lithofacies (Gehrels, 1999). Alternatively, the local litho-, bio-, and chronostratigraphic data has been used to reconstruct error bands of MHWST (Shennan, 1986b). Trends in MTL are also likely to have been influenced by meteorological factors during the Holocene. Gordon and Suthons (1963) and Rossiter (1972) removed the variance associated with surface air pressure, temperature and other forcings to obtain more reliable estimates of changes in mean sea level (MSL). Woodworth (1987) calculated a MSL rise of approximately 2 mm yr^{-1} at Newlyn between 1916 and 1962 using similar methods.

Coastal barrier systems are notoriously difficult sites from which to obtain reliable altitudinal data (Healy, 1993). Such coastal lagoonal areas can present large sources of error when attempting to assess the indicative meaning of samples from rapidly evolving morpho-sedimentary structures. For example, the flooding of back-barriers by unusual storm events could destroy trees, and the water table responsible for the death of the trees may be 1 m higher than that represented by their rooting horizon (Heyworth and Kidson, 1982). The selection of such samples for radiocarbon dating could therefore induce large age and elevation errors. Heyworth and Kidson (1982) suggest that 'exceptional tides' and storm surges may have had an influence on transgressions and that peat may have been deposited in the back-barrier environment at a lower level than expected. Careful interpretation of the lithostratigraphic records in this study will reduce errors in the assignment of indicative meaning from the back-barrier facies of south Devon cores. Relatively small changes in tidal range during the mid- to late Holocene (Austin, 1991) are likely to alter SLIP heights predicted in this thesis, but not significantly (see section 1.6.4).

1.7 Study sites and rationale for selection

Four back-barrier systems along the south Devon coast (Figure 1.12) were selected based on the presence of a succession of well-preserved Holocene inter-tidal peat and minerogenic sedimentary sequences. The sites also offer access for vehicle and drilling rig by nearby road or track. The subsurface lithostratigraphy of the back-barrier systems was established from a pilot study carried out by the supervisory team during the early part of 1997. A 12.50 m borehole log, extracted from North Sands (SX 730 382), was obtained from South West Water (1992) from the construction of the Salcombe foul water discharge facility. It revealed bedrock overlain by a basal peat unit, covered by minerogenic sediments and shells. This provided impetus for the study outlined here because it seemed likely that other basal peat units existed in similar settings along the south Devon coastline. Subsequent reconnaissance coring with an Eijkelkamp hand corer produced a 4.5 m section, from Bantham Sands (SX 665 435) revealing sand and shells (4.50 – 3.90 m), a wood peat (3.90 – 3.80 m), *gyttja* or lacustrine clay (3.80 – 2.80 m), and freshwater peat (2.80 – 0.00 m). These sequences clearly indicated the site as being suitable for the purpose of sea-level study because they contain transgressive and regressive overlaps. Deeper cores were required to establish this more clearly and obtain the basal sequences that would provide the long-term (10^3 yr) RSL history for south Devon.

The sites chosen on the basis of these results are Bantham Sands (SX 664 435); North Sands (SX 729 382); Slapton Sands (SX 825 432); and Blackpool Sands (SX 852 478). All are part of the South Hams District Area of Outstanding Natural Beauty and Coastal Preservation Area (Mildmay-White, 1985; Salcombe and Kingsbridge Estuary Association, 1985, Figure 1.12). Slapton Sands, Slapton Ley and the surrounding woodland is also a UK Grade 1 Site of Special Scientific Interest managed by the Field Studies Council as a National Nature Reserve (O'Sullivan, 1994; Slapton Line Defence Group, 2002).

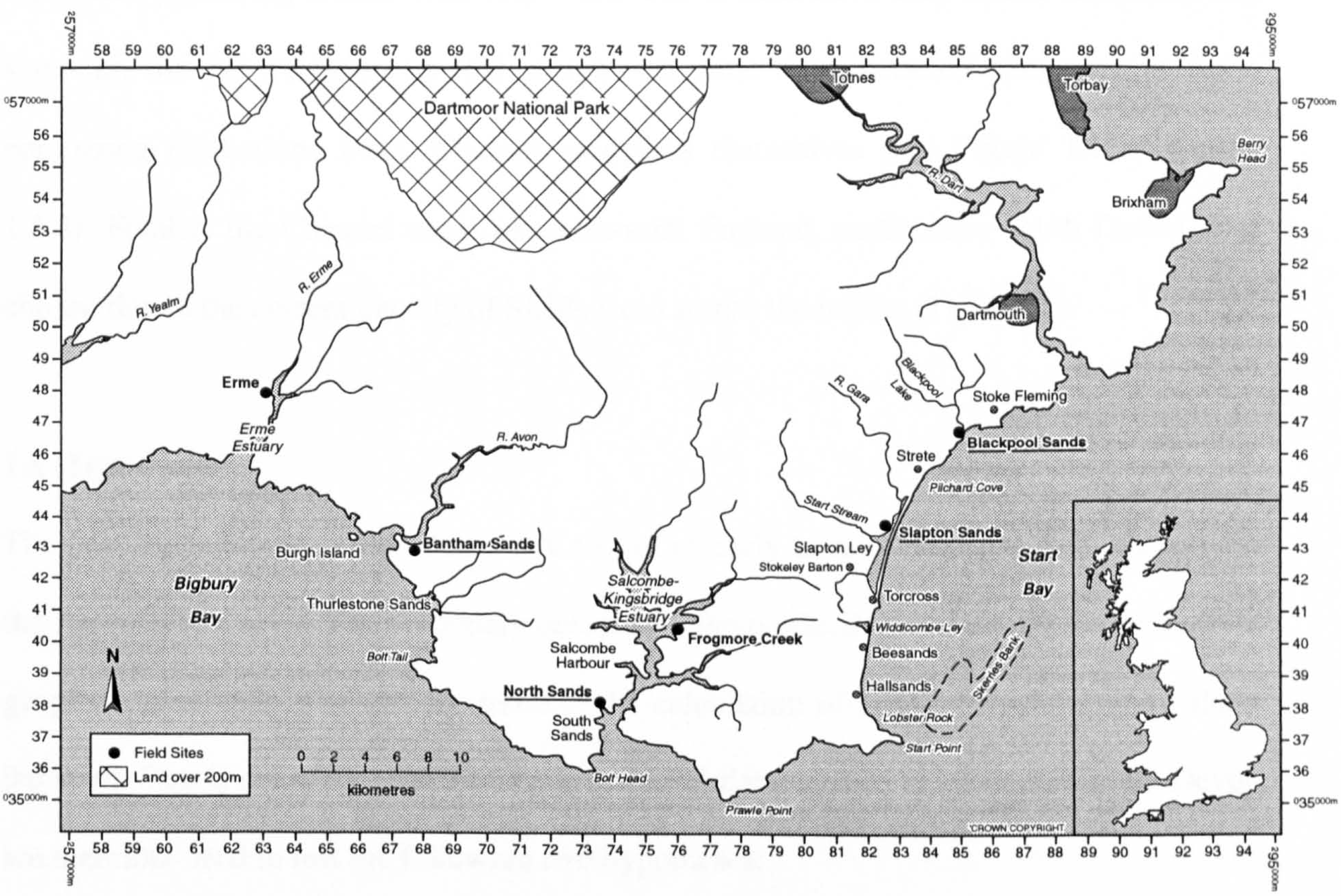


Figure 1.12. Location of study sites along the south Devon coastline. Coring sites (see chapters 4, 5, 6 and 7) are shown bold and underlined.

The back-barrier sites in south Devon were once marine environments. Bedrock and Pleistocene head deposits are overlain by basal peat facies containing fossil indicators of inter-tidal environments which in turn are buried by thick Holocene minerogenic sediments of shallow marine or lagoonal origin (Hails, 1975a,b; Lees, 1975; Clarke, 1970; Morey, 1976, 1983a,b). Fringing salt marshes are typical in the rias (post-glacially drowned river valleys) of south-west England (Long and Mason, 1983). This geomorphological setting means that sediments preserved in the present-day back-barrier environment are suitable for sea-level reconstruction because an indicative meaning can be obtained from them. Holocene basal sequences occur at a range of depths offering the potential for a RSL history to be reconstructed. These facies also provide compaction-free SLIPs. The selected sites are geographically close to one another allowing for a between-site comparison to be carried out. This is because the isostatic histories of the sites should be more or less the same thus producing similar RSL responses. The models have very coarse resolution and show no distinct variation in RSL height per time slice between sites, except when comparing predictions from different modellers themselves (see Figure 1.11a, section 1.5.3). Finally, the Channel coast of south-west England, particularly south Devon, was chosen due to the current paucity of SLIPs from across the region (Table 1.1).

1.8 Hypotheses

The primary outcome of this thesis is a record of early to mid-Holocene RSL change for the south Devon coast. The secondary outcome is the production of data able to test current geophysical models. The tertiary result is the calculation of rates of crustal motion along the south Devon coastline. The Holocene sea-level data derived from sites in south Devon are therefore used to test the following two hypotheses:

1) The Holocene RSL history for south Devon is in agreement with existing data from the Holocene SLIP database for SW Britain, and rates of crustal motion calculated from new SLIPs along the south Devon coast are similar to those calculated by Shennan and Horton (2002) from the existing SLIP database.
2) Geophysical models are reliable predictors of Holocene RSL change in south Devon, but the Peltier model, which predicts a glacio-isostatic forebulge, is in closer agreement with the geological data than the Lambeck model.

1.9 Wider scope

The collapse of a glacio-isostatically induced bulge in the lithosphere created south of the former British Ice Sheet (Carter, 1982a; Devoy, 1983; Wingfield, 1995) is likely to be reflected in the early Holocene RSL history for south Devon. A similar mechanism has been proposed by Pardi and Newman (1987), Gehrels and Belknap (1993), Barnhardt *et al.* (1995), Davis and Mitrovica (1996) and Gehrels *et al.* (1996) along the U.S. eastern seaboard. For example, if empirical rates of RSL rise are faster than those predicted by the models it raises the possibility that ice load was greater than described in the models and forebulge collapse more significant. This is of particular importance when evaluating Holocene RSL results derived from Lambeck's (1991, 1993a, 1995) models (e.g., Figure 1.11b), which do not predict a forebulge mechanism (Devoy, 1995). Such a conclusion would favour Peltier's (2000) and Peltier *et al.* (2002) modelled sea-level curves for south Devon (Figure 1.2) given Peltier's (1998) conclusion that ice-load over Scotland was sufficient to produce such a response.

Rates of crustal motion along the Channel coast of south-west England will be calculated according to Shennan and Horton (2002). They calculated the rate for each area around the British Isles using index points younger than 4000 cal years BP, similar to Shennan (1989) because the eustatic model adopted over that time period was approximately zero. Shennan (1989) and Shennan and Horton (2002) have shown that uplift is still ongoing in the north of Great Britain and that subsidence is still occurring in the south (Figure 1.13a,b). The 2002 analysis shows that the highest rates of subsidence occur in SW England, whereas subsidence in SE England is less significant than previously thought. Furthermore, it is apparent from a comparison between the two maps of crustal residuals that Holocene SLIPs remain scarce for southern Britain and SW England in particular. Rates of crustal motion cannot be fully elucidated until an increase in the number and improvement in the reliability of data has been achieved.

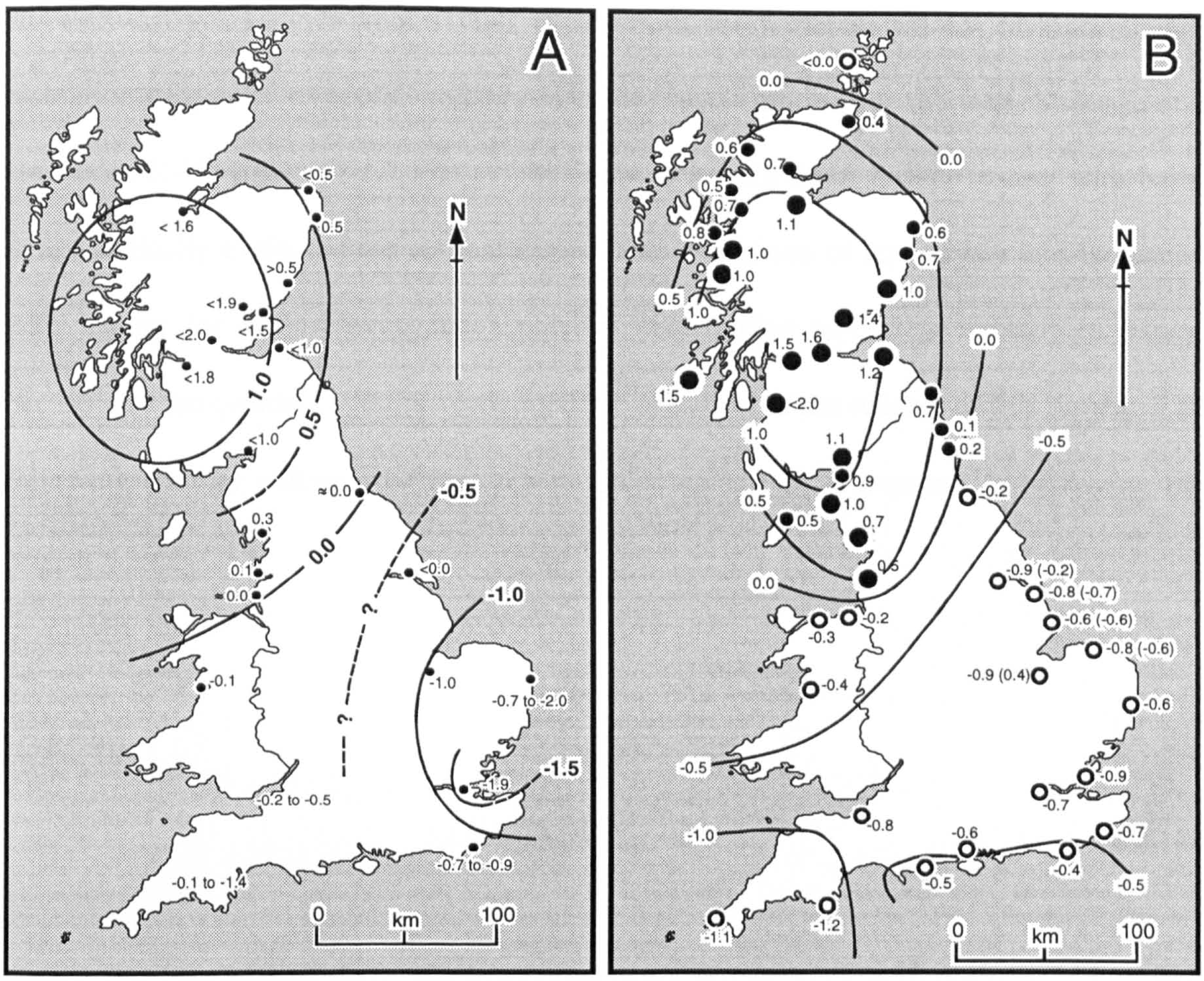


Figure 1.13. Comparing maps of late Holocene crustal movement in the British Isles. Source: (A) Shennan (1989), (B) Shennan and Horton (2002). See Figure 1.3 caption for explanation.

1.10 Summary

There is a paucity of reliable Holocene sea-level index points along the Channel coast of south-west England. The Holocene SLIP database for south-west Britain currently holds only 5 SLIPs from south Devon, 6 from south Cornwall and 4 from Dorset. There are no index points pre-7200 or post-1100 cal years BP and the vertical disagreement between them is as much as 3.5 m (*ca.* 6000 cal years BP). Current geophysical models for the region exhibit even greater RSL height differences, particularly *ca.* 7000 cal years BP where a 6 m disagreement exists between the models of Lambeck (1995) and Peltier (1998). The majority of mid- to late Holocene SLIPs are overestimated by the models and there is a current lack of data from the English Channel coast to test the model divergences *ca.* 7000 cal years BP (Figure 1.11a). SLIPs from south Devon are not of high enough resolution or quality to predict reliably rates of crustal motion for the area. Shennan and Horton's (2002) figure of 1.2 mm yr^{-1} is likely to benefit from further testing with basal SLIPs critically evaluated for vertical errors. The production of an early to mid-Holocene RSL history for south Devon and a positive contribution to current unresolved issues, particularly the quantification of vertical errors, is crucial to the success of this project. The following chapter outlines the research methodologies that are central to achieving the project's aims.

Chapter 2

Research Methods

2.1 Introduction

A number of authors have discussed various approaches to sea-level reconstruction (van de Plassche, 1977, 1986; Shennan, 1983; Devoy, 1987; Carter, 1992). However, critical issues, particularly the quantification of vertical and horizontal margins of error (Heyworth and Kidson, 1982; van de Plassche, 1982; Shennan, 1986a,b; Paul and Barras, 1998; Gehrels, 1999; Gehrels *et al.*, 2001) are still ignored by some sea-level researchers. This chapter will elucidate the methods used to produce a reliable reconstruction of Holocene sea-level change along the Channel coast of south Devon and contribute to the resolution of these issues.

2.2 Field techniques

2.2.1 Coring

The back-barrier systems (Figure 1.12) were cored between 1997 and 1999 using a percussion drilling rig (Figure 2.1a,b) manned by a crew from Site Investigations Services (Gainsborough, Lincolnshire). All core sections were acquired in 94 mm diameter core sleeves, 1 m in length, for Bantham Sands (2 cores), North Sands (4 cores) and Blackpool Sands (3 cores) and 76 mm diameter sleeves for Slapton Sands (2 cores). Geotechnical Engineering (Gloucester) recovered a third core from Slapton Sands in April 2001 during an Environmental Impact Assessment (EIA) Report for English Nature, preceding a proposed road building project at the site. This yielded 75 mm diameter core sections up to 1.5 m in length, for sections where recovery was possible, using a rotary coring rig with a hammer percussion attachment. All core sections were sealed in plastic sleeves and placed in a cold store at +4°C at the end of each working day.

A. Core drilling

B. Core extraction, showing metal outer casing (drill bit guide)

Figure 2.1. Percussion drilling rig in operation at Slapton Sands.

Partial recovery of some core sections occurred upon extraction and such volume changes may be due to a number of factors. For example, sandy sediment may have settled in the liner (tube or sleeve) after it was lifted above the water table. This is known as sediment de-watering. The vertical datum was recorded at the top of each section and core loss at the base. Compaction of clay, silt and peat is a problem with percussion rigs and other non-rotary techniques (Emery and Dietz, 1941; Richards and Keller, 1961; Rees, 1999) and was measured in the field. Sediment can also be lost from the bottom of the barrel during withdrawal from the borehole. A sediment catcher attached to the base of each sleeve prevented this and the 'captured' sediment was later retained in a bag or smaller tube. Volume increases can occur as sediment 'caves in' from a higher level than that which is cored, and through the swelling of peat lithologies. A metal outer casing was placed in the borehole as a guide to the drill bit (Figure 2.1b), also reducing the amount of sediment 'cave in' to very few occurrences. The uppermost sections in some cores were not retrieved (see chapters 4 to 7). This practice was carried out to save time and alleviate financial expenditure. These sections were described in the field. No sediment was lost that may provide a site-specific RSL history using the methods described in this chapter. Finally, core sections were stored and opened horizontally to avoid any vertical (upward or downward) movement of sediment in the liner.

2.2.2 Surveying and datum selection

A Zeiss Electronic Distance Measurer (EDM) was used to level all borehole locations to the UK Ordnance Datum (Newlyn) using the nearest Ordnance Survey (OS) benchmark. Attempts were made to keep each individual survey line (between EDM and prism) to within a maximum distance of 300 m. Zeiss recommend this method as it significantly reduces vertical and horizontal error margins caused by atmospheric disruption. Altitude and horizontal distance were recorded to the nearest millimetre. Angular measurements were recorded to the nearest 1/1000 of a degree. The horizontal and angular measurements were used to locate the borehole positions on a large-scale OS map. Locations were confirmed using a Garmin hand-held Global Positioning System (GPS).

The reduction of vertical errors in the survey is an important objective of this study. When horizontal distances from EDM to prism were greater than 300 m, several readings were taken and a mean height calculated. Multiple OS benchmarks were also used to obtain further checks on the altitude of borehole locations. This is particularly important where benchmarks were deemed unreliable on their own, e.g., structures subject to subsidence and benchmarks not recently (pre-1950) verified by the OS. First order benchmarks were used where available and any lower order benchmarks were checked with the OS before being utilised in the survey.

2.2.3 Electrical resistivity surveying

Resistivity surveying was carried out according to Overmeeren and Ritsema (1988) using the ABEM Lund Imaging System to obtain stratigraphical information in addition to the cores. Field deployment of the system can be seen in Figure 2.2. In most cases a shore-parallel and shore-normal survey of the back-barrier was completed. This was altered at Blackpool Sands where the borehole transect is shore-diagonal (Figure 7.2) and in close proximity to a road (see section 7.3.4 for explanation). A survey was carried out as close to the borehole transect as practical to enable the lithostratigraphy to be superimposed onto the resistivity profile. Resistivity readings were taken at 5 m intervals along each cable with the deepest recordings in the centre (zero-point, see Bernstone and Dahlin, 1996 and Dahlin *et al.*, 1996). Core records are used as ground-truths and individual sedimentary facies are interpreted on the basis of their resistivity ranges and median values that are given in Ωm.

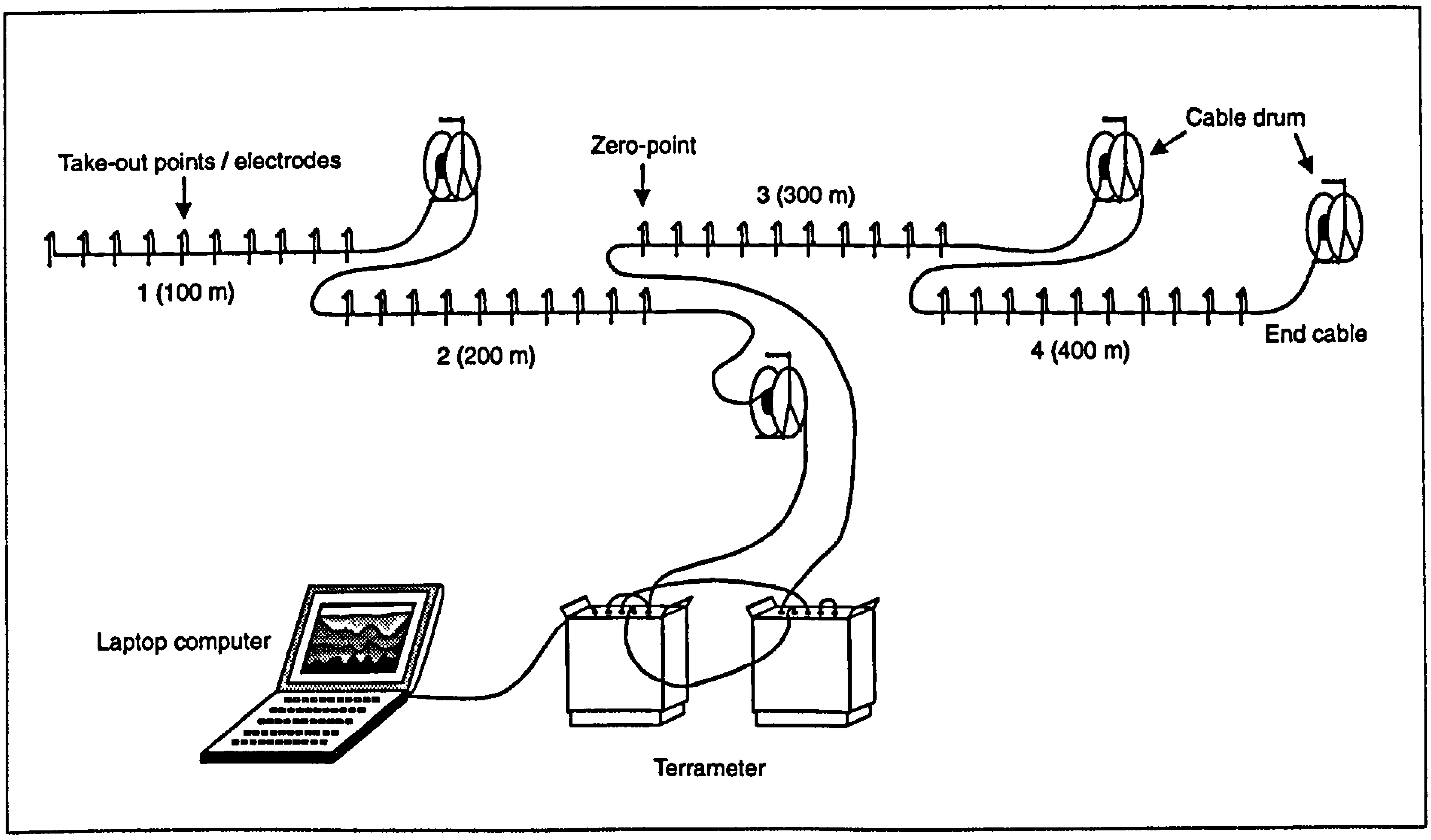


Figure 2.2. Principal system layout for roll-along resistivity surveying using four electrode cables.

Topographic surveys of the cable array were carried out using the EDM unit and surveyed to an OS benchmark at the site. Height measurements along the cable lengths were taken at each individual electrode take-out point on very undulating ground. This method was changed to every 10 m on ground with minor height changes (±0.10 m) along the survey line. On level ground the method was altered to include measurements at 'zero point' and the ends of cables. The topographic survey data is superimposed onto the resistivity profile.

2.3 Laboratory techniques

2.3.1 Lithostratigraphical analysis

Cores were opened in the laboratory using a cutting device designed by the author (Figure 2.3). A core section (plus end caps) was placed horizontally in the device and the plastic sleeve cut, then rotated through 180°, and cut again. The end caps were reattached to the two halves of the sleeves. A blade (core-cutter) was then pushed down through the cut sleeve to reveal both halves of the sedimentary profile. Once exposed to air, sections were recorded as quickly as possible before oxidisation occurred. Core sections were logged using a key based on the Troels-Smith (1955) scheme (sketches and explanatory text are included in the CD-ROM, Appendices 1a, b and c) and one half of each core section was photographed (CD-ROM, Appendix 1d) and archived. This section was preserved intact for radiocarbon sampling. The other half of the section was sampled, where required, for foraminifera, carbon and pollen (North Sands only), liquid and plastic limits, particle size analysis, sediment moisture and sediment density (North Sands and Blackpool Sands only). Core sections were then wrapped tightly in cellophane-wrap and heat-sealed in airtight plastic tubing before storing at 4°C.

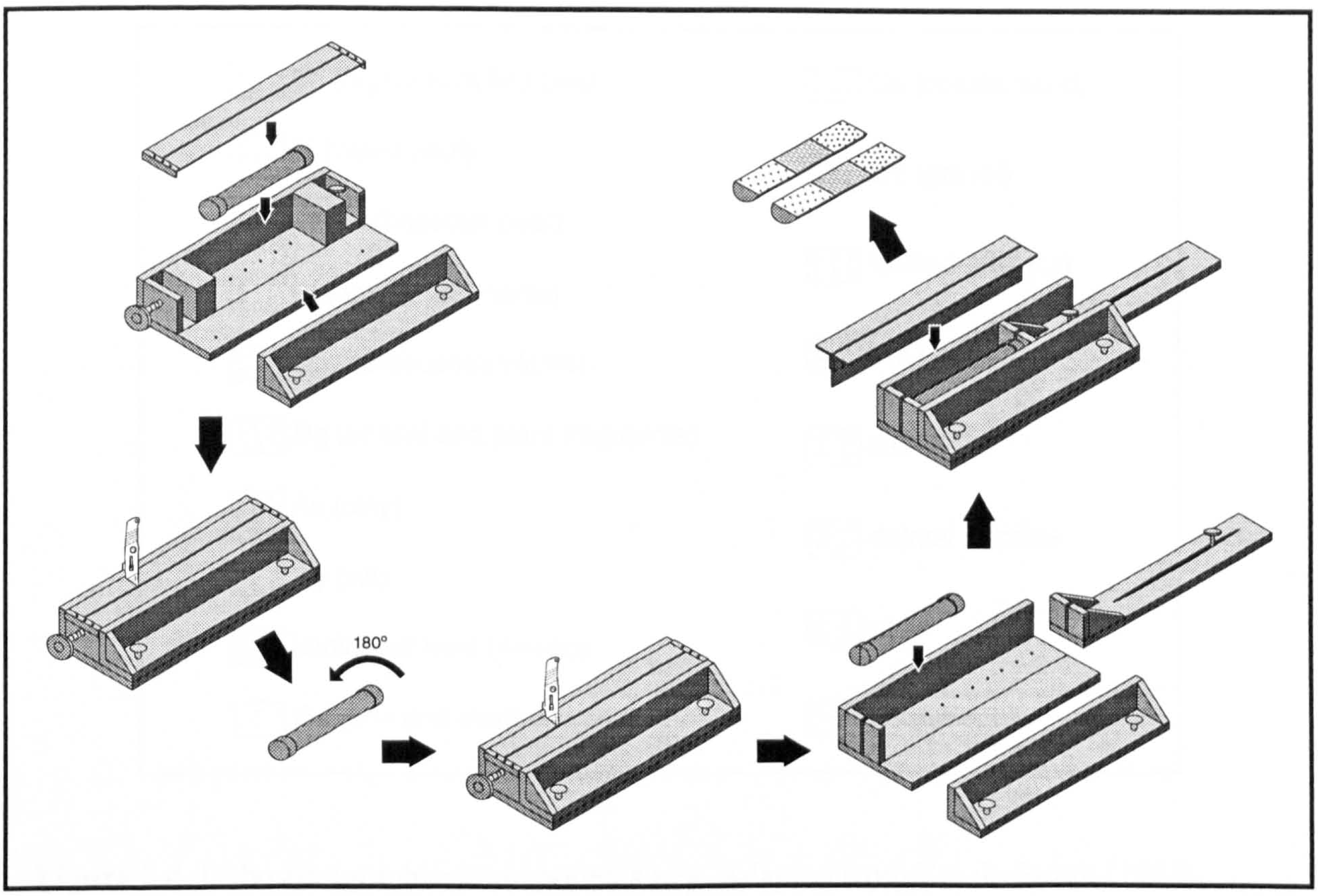


Figure 2.3. Core cutting technique. See text for explanation.

Detailed individual boreholes are depicted using the Troels-Smith (1955) symbols as a guide (Figure 2.4 and chapters 4 to 7) and the Troels-Smith Plotting Program: TSPPlus (Waller *et al.*, 1995) is used to correlate the cores. Linking lines suggest lithostratigraphic connections and only major (≥25%) components are shown for ease of illustration. Units too small (<0.25 m) to be included in the cross-sections are shown in exploded view or indicated with an arrow. Depths for individual borehole records (e.g., Figures 4.4, 4.10, 5.20, 5.21 and 5.29) are given in metres in relation to ground level zero. Depths for stratigraphic profiles and electrical resistivity profiles (e.g., Figures 4.6 and 4.7 respectively) are given in metres OD.

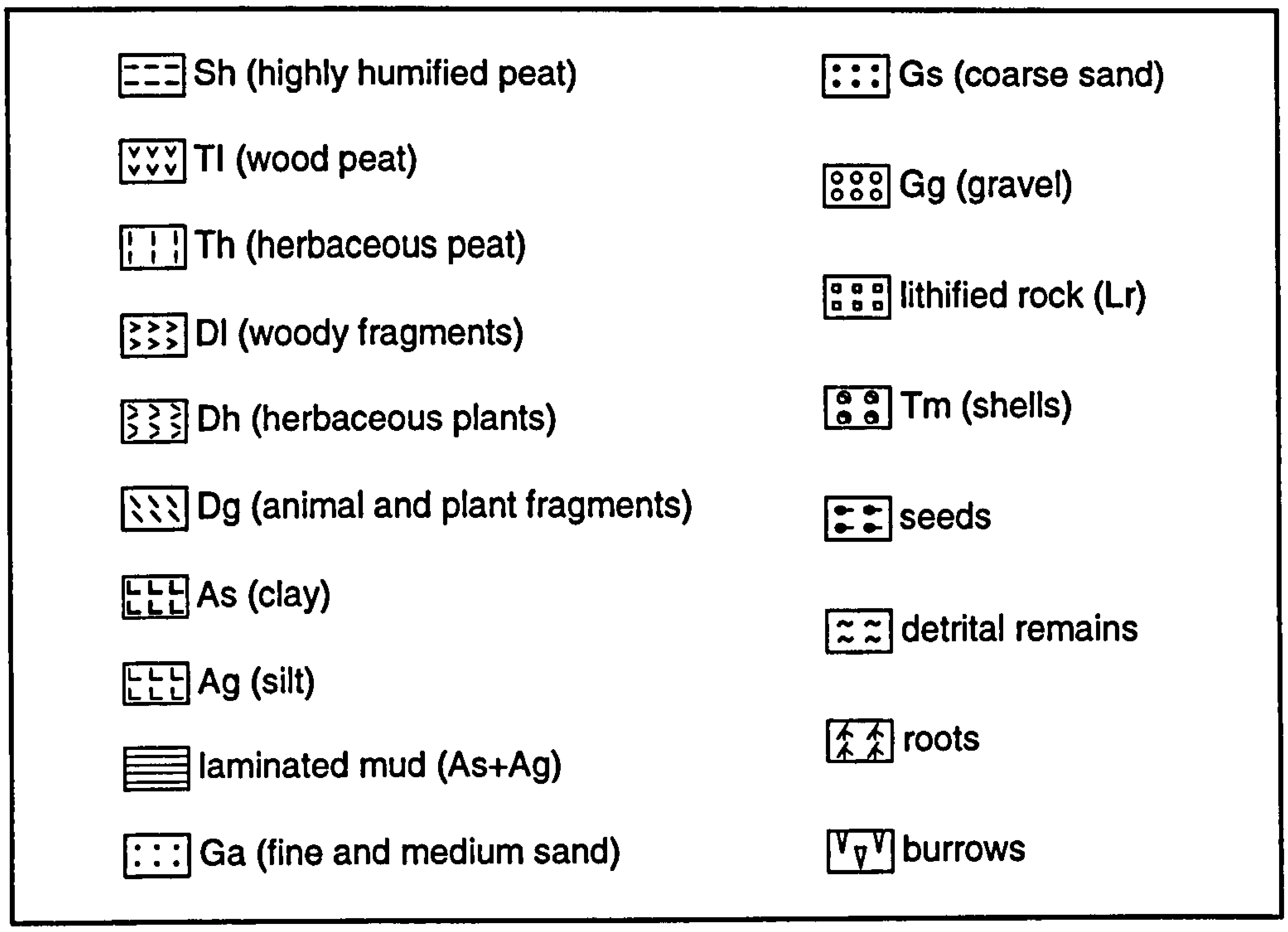


Figure 2.4. Lithostratigraphical components key. Adapted from Troels-Smith (1955).

2.3.2 Foraminiferal analysis

Fossil foraminifera are used to reconstruct Holocene sea-level change along the south Devon coastline and to identify salt-marsh (and mudflat) species in the lithostratigraphy. Modern assemblages are then used to establish quantitatively an indicative meaning for the fossil record (see chapter 3). Foraminiferal taxonomy is based on Murray (1979), Scott and Medioli (1980b), Loeblich and Tappan (1987) and de Rijk (1995). Some species were combined to generic level, e.g., *Quinqueloculina* spp. This was only undertaken where a combination of the species into a single (generic) classification would not hinder the interpretation of the biofacies. Generic names are in accordance with Loeblich and Tappan (1987). Sediments also contained various other faunas and floras that assisted in improving a qualitative interpretation of individual biofacies. For example, the presence or absence of *Acari* spp. (mites, Borror *et al.*, 1989), *Chironomidae* spp. (midges, Cranston, 1982), *Coleoptera* spp. (beetles), *Chara* oospores (aquatic plant macrophytes), ostracods (Boomer and Godwin, 1993; Boomer, 1998), diatoms and moss leaves were noted in all core samples.

Foraminifera were sampled and analysed using standard techniques (Scott and Medioli, 1980b; Gehrels, 2002). Samples measuring 5 ml were initially selected at 2 cm intervals across each lithostratigraphical contact to locate salt-marsh (agglutinated) foraminifera, and increased to 5 cm intervals until mudflat (calcareous) foraminifera dominated. The sampling resolution was then adjusted to allow a detailed reconstruction of the biofacies to be made. All samples were preserved in a 98% ethanol and 2% distilled water solution, and refrigerated to reduce retardation of the tests by biological and chemical action.

Samples to be investigated for the presence of foraminifera were sieved between 500 μm and 63 μm and 'wet-split' into eight equal parts (Scott and Hermelin, 1993; Gehrels, 1994; de Rijk, 1995; Gehrels, 2002). Foraminifera were wet-picked from a Bergeroff tray under an Olympus SZ40 low power binocular microscope with fibre optic illumination. A target of at least 200-300 foraminifera per 1/8th fraction was set (Scott and Leckie, 1990; de Rijk, 1995). Where individuals numbered greater than 1000 per 1/8th fraction, subsequent samples were re-split and a 1/64th or 1/32nd fraction was analysed. Where individuals numbered less than 50 per 1/8th fraction, analysis was increased to a half or whole of the sample. Counts of agglutinated foraminifera were extremely low in some basal and in-core sections (<5 per 5 ml). These include *Trochammina inflata*, *Miliammina fusca*, *Haplophragmoides wilberti* and *J. macrescens*. The sampling resolution was increased to a contiguous (1 cm) strategy, where counts of salt-marsh foraminifera were extremely low, to identify assemblage boundaries more precisely.

A number of protective measures were taken to reduce contamination and damage to tests. These include sampling from a layer of sediment that was not in contact with the core-cutting tool, careful separation of the animals in sieves, and increasing the settling time in the wetsplitter to maximise the capture of lighter specimens e.g., *Trochammina ochracea*.

In addition, organic debris in highly organic sediments was carefully decanted out to improve the recognition of foraminifera.

The reliable species identification of salt-marsh foraminifera seldom proved problematic. For example, juveniles from a small number of samples were unclassifiable due to their simple test form and were therefore recorded as 'unidentified'. Calcareous species were only occasionally discovered at varying stages of dissolution, a problem common in such palaeo-settings (Scott and Medioli, 1978), but were easily identified by their test lining (Edwards and Horton, 2000). The presence of damaged or deformed agglutinated and calcareous tests also hindered identification but the majority of these were mounted on a gridded slide for later classification. Gehrels and van de Plassche (1999) noted that the chambers of *Balticammina pseudomacrescens* often collapsed when dry. This sometimes occurred with juvenile and damaged or deformed salt-marsh species (e.g., *J. macrescens*) but did not significantly hinder classification. Differentiating between poorly preserved *Elphidium williamsoni* and *Elphidium earlandi* proved difficult on occasion but the latter was identified by its fewer retral processes and fewer chambers in the outer whorl (Murray, 1979). Scott (1976) further assisted in the identification of *Polysaccammina ipohalina*, the shape of which can be extremely variable. TILIA and TILIA*GRAPH (Grimm, 1991) is used to describe and depict the assemblage frequencies and palaeo-biozones are reconstructed using a CONISS (Grimm, 1987) cluster analysis. Values for indicative meaning representing optima, minima and maxima of error bars are shown also (see section 3.4.2). Spreadsheets containing raw and percentage count data for all fossil and modern assemblages are included in the CD-ROM (Appendices 2 and 7 respectively).

2.3.3 Carbon analysis

Samples for carbon analyses were selected from a North Sands core to (1) identify potential salt-marsh facies not clarified by the lithostratigraphical analysis, (2) to replace the variable foraminiferal sampling strategy designed to locate those facies, and (3) determine the carbon content of the sediment for ^{14}C dating (see section 5.5). Sedimentary facies were sampled for total and organic carbon content according to Avery and Bascomb (1982). Samples weighing approximately 3 gms were extracted at 2 cm to 5 cm intervals. They were oven dried over-night at 80°C, ground with a pestle and mortar, then sieved to <1 mm. Sub-samples of 50 mg were then prepared for carbon analysis and total carbon content was recorded by a Shimadzu TOC5000/SSOI5000 Carbon Analyser. To determine the organic carbon content sub-samples of 500 mg were digested in 1 ml of 10% hydrochloric acid (HCl) until all inorganic carbon was removed (indicated by the cessation of frothing). After 1 hour, samples were diluted with 10 ml of distilled water, oven dried and analysed for organic carbon content. The acid pre-treatment caused significant problems with the carbon analyser readings and had to be abandoned. However, time was the primary limiting factor of the method and carbon analysis was found to be less efficient at establishing salt-marsh facies than the foraminiferal method. It was not used in other sites. Carbon percentage values are included in the CD-ROM (Appendix 3).

2.3.4 AMS ^{14}C radiocarbon dating

Radiocarbon analyses are used to date portions of the cores from which an indicative meaning could be established on the basis of the foraminiferal analyses. Compared to the conventional ^{14}C method, AMS ^{14}C radiocarbon dating allows smaller samples to be selected, e.g., horizontally embedded individual plant fossil remains (Gehrels and Belknap, 1993; Gehrels *et al.*, 1996; Gehrels, 1994, 1999). The re-deposition of fragile plant fragments is unlikely, thus providing a more accurate dating tool of former surfaces.

Samples were selected primarily on the presence of agglutinated foraminifera identified in salt-marsh facies (Scott and Medioli, 1978), and sampling was carried out according to Gehrels (1999). The radiocarbon dating of Holocene basal salt-marsh facies provides compaction-free SLIPs, establishing the long-term (10^3 yr) sea-level trend. In-core samples were selected to provide additional SLIPs and to determine the timing of barrier closure in some cores (North Sands and Blackpool Sands). A 1 cm slice was cut from selected horizons and the edges removed to reduce contamination. Where present, horizontally embedded plant macrofossil remains were selected from individual horizons with tweezers. In their absence, wood fragments or bulk samples were selected. The material was washed over a 63 μm sieve with distilled water, oven-dried overnight at 55°C then weighed. Dried samples weighed approximately 10 - 20 mg. Bulk samples were heavier. Samples were wrapped in aluminium foil and stored in airtight plastic containers.

Bulk samples were digested in hydrochloric acid then rinsed with distilled water to eliminate inorganic carbon (shells and calcareous foraminifera) from the sediment. Samples were sent immediately to the Natural Environment Research Council (NERC) Radiocarbon Laboratory at East Kilbride in Scotland, where they were prepared to graphite. Two submissions were made which resulted in radiocarbon analysis allocations (see acknowledgements, p.xix, for allocation numbers). AMS ^{14}C measurements were conducted in the NSF-AMS Facility at the University of Arizona, and the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, University of California. The first batch from North Sands was oven dried. The second batch of samples, from Bantham Sands, Slapton Sands and Blackpool Sands, was not dried but only cleaned in distilled water by ultra-sound and placed in airtight vials. All radiocarbon ages provided by NERC are calibrated according to Stuiver *et al.* (1998a) using the Extended ^{14}C Data Base Revised Calib 4.2 Radiocarbon Calibration Program 2000 (previously used in accordance with Stuiver and Reimer, 1993). Ages are given in calibrated years BP with the $\pm 2\sigma$ range.

2.3.5 Pollen analysis

Pollen has long been established as a useful indicator for reconstructing Holocene coastal palaeoenvironments (e.g., Godwin, 1940; Crabtree and Round, 1967; Kidson and Heyworth, 1973; Morey, 1983a; James and Guttmann, 1992; Hewlett and Birnie, 1996; Lloyd *et al.*, 1999; Long *et al.*, 1999; Waller *et al.*, 1999). Fossil pollen is used to test the ^{14}C chronology and the continuity of one core in this project (see section 5.7.2). Pollen analysis was carried out by Dr Katherine Selby from the School of Geography and Archaeology, University of Exeter.

Pollen was sampled, prepared and analysed based on Moore *et al.* (1991). Samples measuring 2 ml were initially selected directly above and below AMS ^{14}C sample horizons. This method served as a pilot study. The sampling resolution was then adjusted to approximately 5 cm intervals to allow a detailed reconstruction of the pollen zones to be made. Three *Lycopodium* tablets per sample were added and potassium hydroxide was used to remove humic acid in the sediment and disaggregate any acid peat. Fine sieving to 10 μm removed silt and clay particles. A hot hydrofluoric acid treatment was carried out to remove silica, and cellulose was removed with an acetylation mixture (glacial acetic acid). Samples were then mounted on slides with silicone oil prior to microscopic analyses.

Identification followed Andrews (1984) and Moore *et al.* (1991). Nomenclature follows Stace (1991) and Bennett *et al.* (1994). Pollen samples were counted under an Olympus CH2 microscope with a x10 objective and a x40 lens. A target of at least 300 total land pollen grains per sample was set (Moore *et al.*, 1991). The total land pollen included arboreal and non-arboreal pollen and all spores and aquatics were recorded although they did not form part of the pollen sum. Traverses covered the whole slide to account for the non-randomness of pollen distribution on the slide (Brookes and Thomas, 1967).

The pollen density in a few samples from the basal section was extremely low (app. 0 to 80 grains). The majority of slides analysed from in-core sections contained only 100 to 200 land pollen grains, and a significant number of damaged grains were discovered in these samples. This is similar to the problems observed in the foraminiferal analyses (section 2.3.2) and is discussed in the results (section 5.7.2). Steps were taken to reduce damage and contamination to the pollen samples during the preparatory stages (see Moore *et al.*, 1991). The high resolution (1 to 5 cm) sampling technique was also used to improve interpretation of the pollen zones. TILIA and TILIA*GRAPH (Grimm, 1991) are used to calculate and graph the pollen assemblage frequencies, and biozones are reconstructed using a CONISS (Grimm, 1987) cluster analysis. Raw and percentage counts for pollen assemblages are included in the CD-ROM (Appendix 4).

2.3.6 Geotechnical correction for sediment autocompaction

Correcting sedimentary sequences for autocompaction is a crucial objective of this project because previous studies have either ignored the issue or only qualitatively estimated the errors. Post-depositional compression (autocompaction) causes significant vertical margins of error (Paul and Barras, 1998; van de Plassche, 1982; Gehrels, 1999) that may displace SLIPs from intercalated sequences on the order of decimetres to metres. A geotechnical correction is carried out on four cores, two from North Sands and two from Blackpool Sands, using the Paul and Barras (1998) model (see chapters 5 and 7 and Figure 2.5). Geotechnical modelling was conducted by Professor Mike Paul from the School of Life Sciences, Department of Science, Engineering and Technology, Heriot-Watt University, Edinburgh.

The model requires a number of parameters to be investigated including (1) the compressibility of the sediment, (2) the bulk sediment density, (3) the pore water pressure, (4) the total stress of the sedimentary sequence, and (5) temporal and spatial information

(Paul and Barras, 1998). The methods and classifications for parameters 1 to 4 are based on standard techniques and are included in the CD-ROM (Appendix 5). The determination of temporal information (^{14}C analysis) is discussed in section 2.3.4, and spatial information (the geological and geographical setting) is given in chapters 5 and 7.

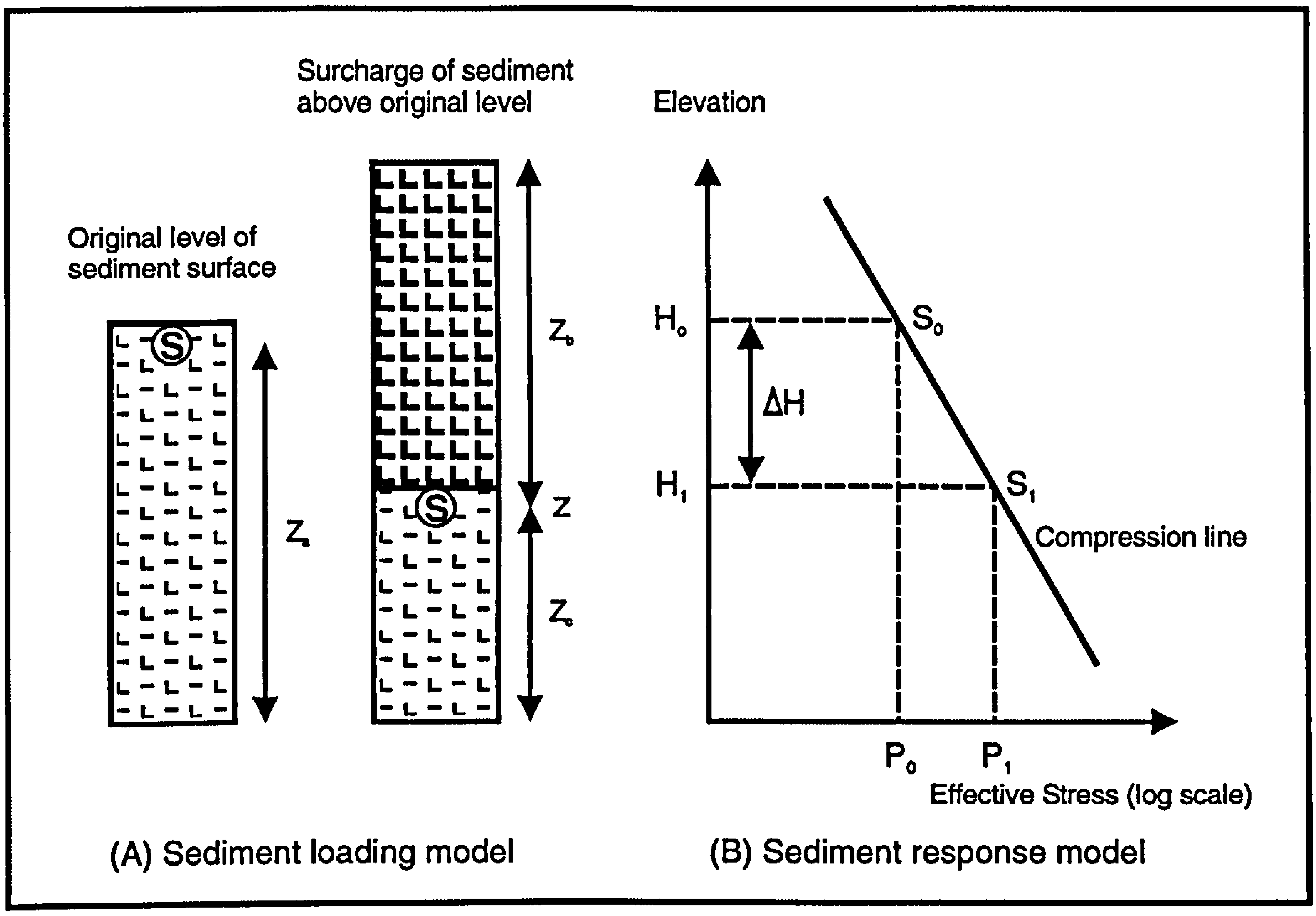


Figure 2.5. The geotechnical basis of the Paul and Barras (1998) model. (a) A sample *S* lies at a height Z_a above a datum level. The sample is subsequently buried by Z_b depth of sediment that reduces its elevation by ΔH to Z_c. (b) The geotechnical response of the sediment: At elevation H_0 the effective stress on the sample is P_0 but when it is buried to a depth Z the effective stress is raised to P_1. In response the underlying sediment compresses along the line from S_0 to S_1 thus lowering the sample elevation to H_1. ΔH is therefore the vertical displacement, in metres, calculated by the model.

2.3.7 Determination of the sea-level index point

The heights of SLIPs in metres relative to MTL in this project are calculated according to Gehrels (1999). The equation is:

$$\text{SLIP} = \text{H} - \text{D} - \text{I} + \text{T} + \text{A},$$

where H = the height of the back-barrier marsh surface relative to MTL, D = sample depth down-core relative to the ground surface, I = indicative meaning of the sample sediment surface relative to MTL, T = difference between the present and former mean high water level relative to MTL, and A = autocompaction of the sediment as determined by geotechnical correction. Vertical error margins associated with the altitudinal reference are based on transfer function results (see section 3.4.2).

SLIPs are plotted as grey boxes for $\pm 1\sigma$, white boxes for $\pm 2\sigma$ calibrated age ranges, while the height of the box reflect the vertical errors. Crosses indicate the median calibrated ^{14}C age and mid-point of the MTL range. The mid-point of the SLIP MTL range is used to compare model and empirical data and heights are shown in relation to metres RSL. A colour-coded key for all graphs showing model RSL histories and SLIPs can be seen in Figure 2.6. Raw numbers for all geophysical modelling runs are included in the CD-ROM (Appendix 6).

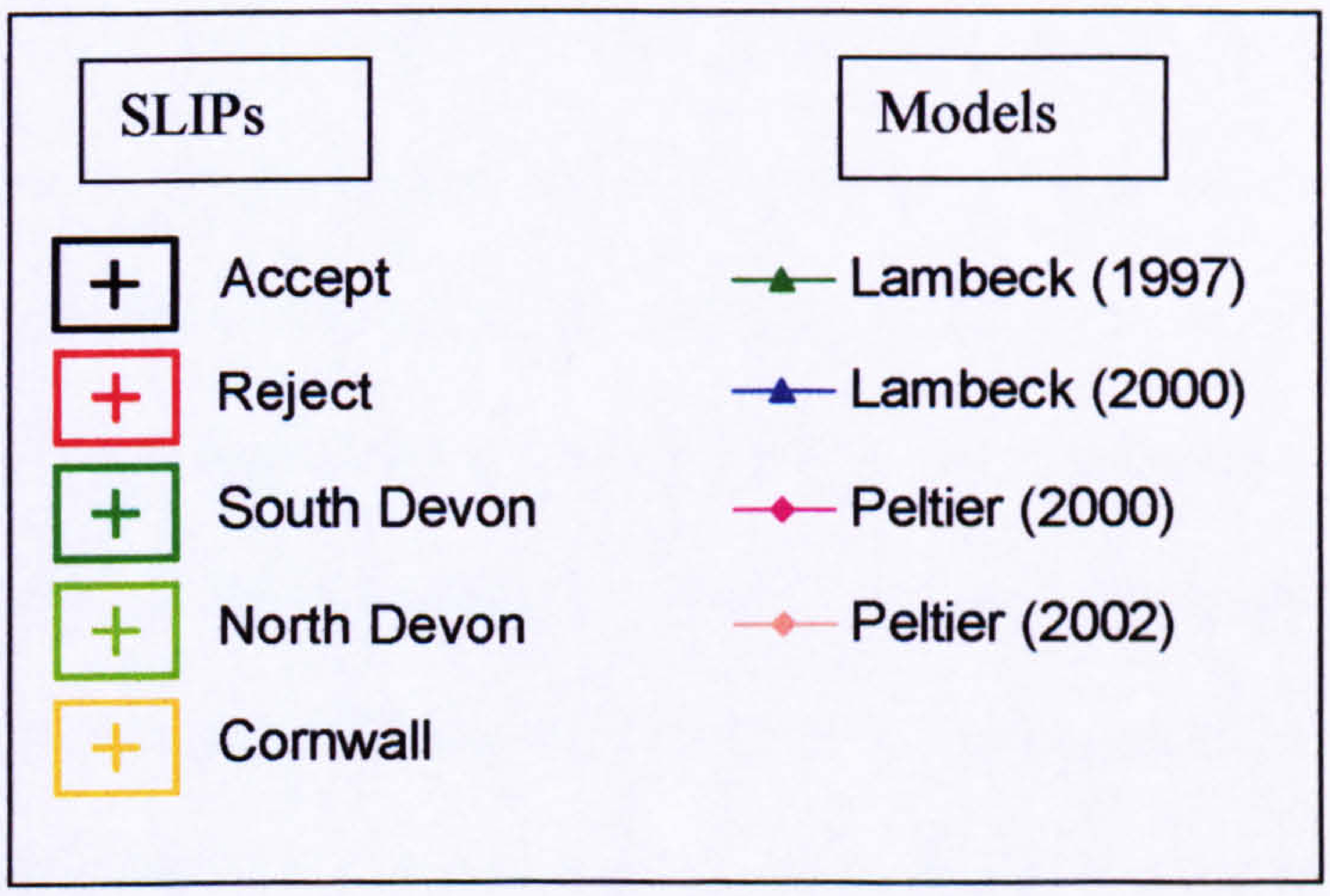


Figure 2.6. Colour-coded key for SLIPs and models shown in this thesis. Symbol for Cornwall includes data from the Isles of Scilly.

2.4 Summary

The aim of this chapter is to present methods that successfully facilitate the reconstruction of Holocene RSL change along the south Devon coastline. The use of drilling rigs capable of reaching deep (20 m) Holocene sedimentary sequences to the Pleistocene bedrock is a primary method. Precise surveying using an EDM reduces altitudinal errors. Electrical resistivity surveying provides further lithostratigraphical information of the back-barrier sediments. Biostratigraphical sampling methods quickly identify basal and in-core inter-tidal fossil foraminiferal sequences and AMS ^{14}C dating provide a more precise Holocene chronology. In one core, the total carbon content of the sediment is determined for ^{14}C dating and fossil pollen analysis assesses the possibility of sediment reworking. A geotechnical correction is carried out on sedimentary sequences containing in-core SLIPs. A modern analogue data set is developed to provide a regional transfer function for south-west England to calculate the indicative meaning of the AMS ^{14}C dated sample. This is explained further in the following chapter.

Chapter 3

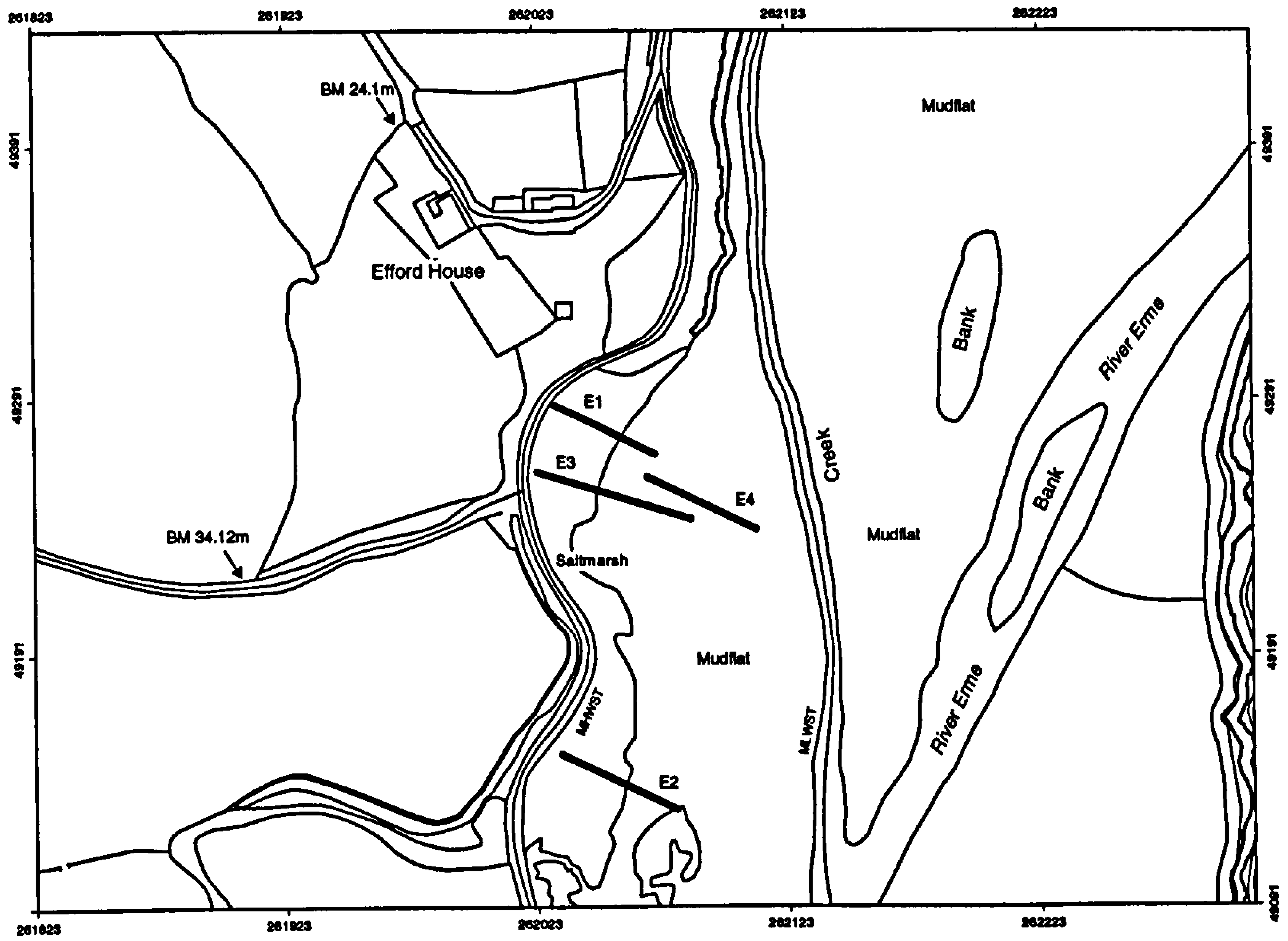
Transfer Function

3.1 Introduction

Foraminifera preserved in Holocene coastal sediments from Bantham Sands, North Sands, Slapton Sands and Blackpool Sands are assigned an indicative meaning based on the vertical distribution of their modern counterparts (chapters 4 to 7). Surface samples were collected from the Erme estuary and Frogmore Creek in the Salcombe-Kingsbridge estuary, south Devon (Figure 3.1). Both sites are part of the South Hams District Area of Outstanding Natural Beauty and are Sites of Special Scientific Interest and Coastal Preservation Areas (Mildmay-White, 1985; Salcombe and Kingsbridge Estuary Association, 1985). These contemporary salt marsh and mudflat environments provide a modern analogue data set for the sedimentary palaeoenvironments.

The Erme salt marsh is an estuarine fringing marsh and its suitability as a modern analogue for a back-barrier site may be questioned on the grounds that river discharge and the absence of a barrier could produce a tidal environment that is different from a back-barrier setting. However, the Erme estuary is selected as it is one of the most 'pristine' estuaries in the south-west and not significantly impacted by human activities such as mining and farming. It also contains the most extensive area of salt marsh in south Devon between the Tamar and Exe estuaries. Frogmore Creek is a close modern analogue for the palaeoenvironmental setting of one of the coring sites (North Sands), at least in the mudflat stage of its development, and it may be noted that the upper foraminiferal assemblages in Frogmore Creek and the lower assemblages in the Erme estuary show distinct similarities. The foraminiferal assemblages in the Erme estuary may therefore indeed be regionally representative. However, this remains an assumption due to the lack of contemporary back-barrier salt-marsh settings in south Devon.

Erme estuary



Frogmore Creek

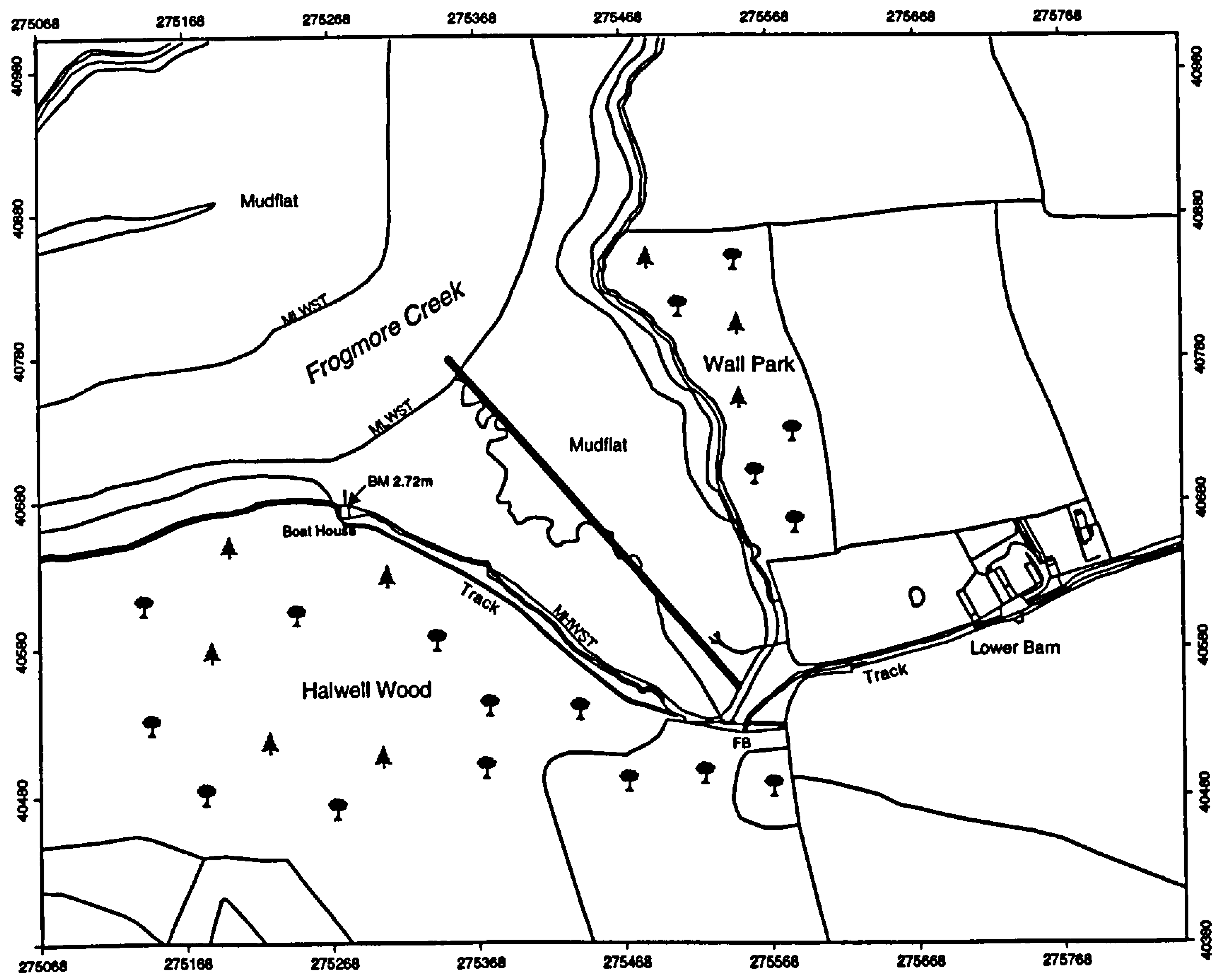


Figure 3.1. Erme estuary and Frogmore Creek mudflat sites with transect locations. Erme estuary 1 and 2 (Gehrels *et al.*, 2001); Erme estuary 3 (White, 2001); Erme estuary 4 and Frogmore Creek (this project) transects. Map extract at scale 1:2500 (2002). Source: http://digimap.edina.ac.uk/service/advanced/carto_welcome.html.

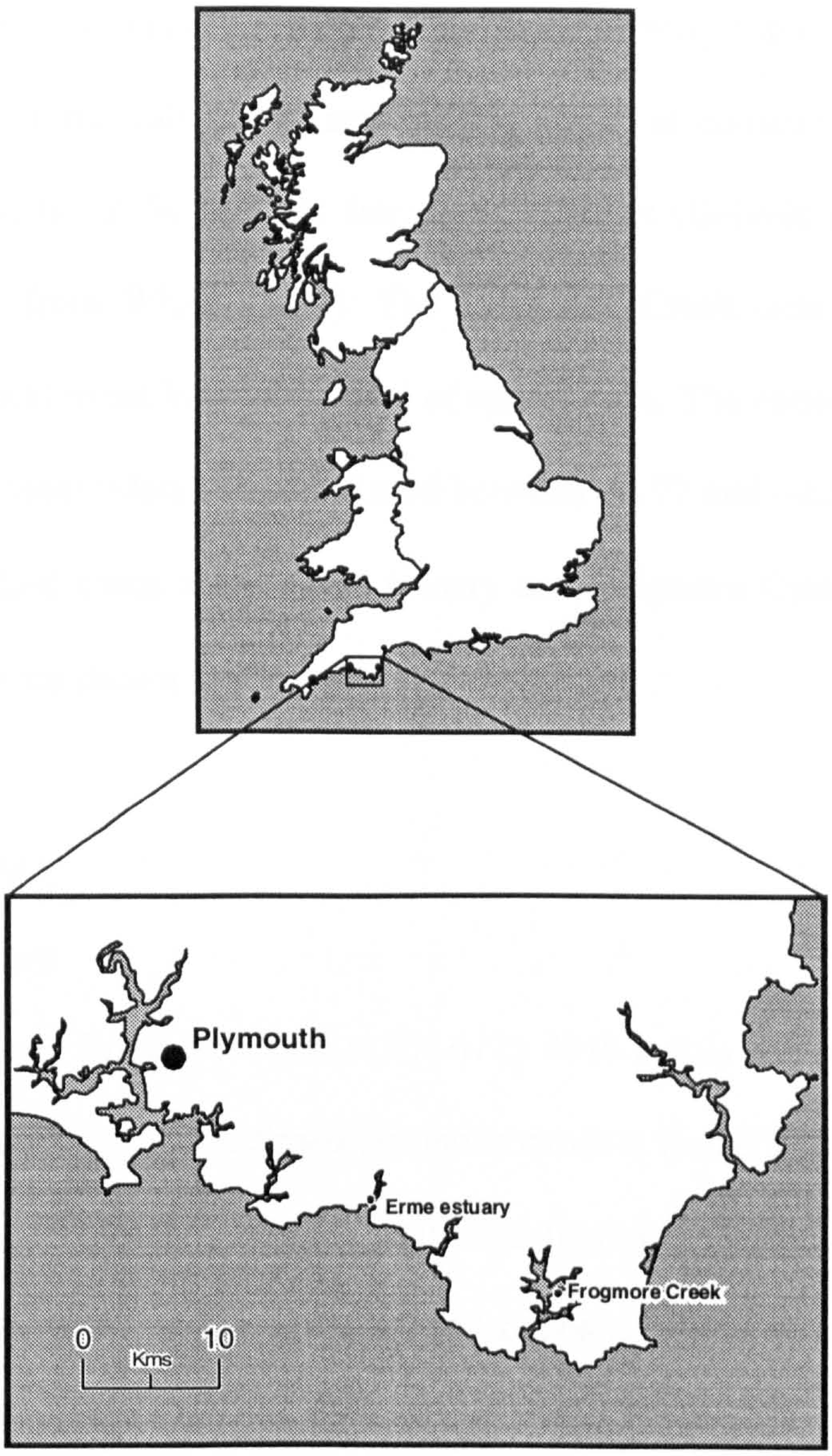


Figure 3.1 continued. Location of the Erme estuary and Frogmore Creek on the south Devon coastline, south-west England.

Transfer functions have been applied in a limited number of sea-level studies (Horton *et al.*, 1999a,b; Edwards and Horton, 2000; Edwards, 2001; Gehrels *et al.*, 2001). Modern analogue data presented here will enhance the current UK inter-tidal foraminifera database and provide a regional transfer function to reconstruct inter-tidal water levels in south-west England. The data set includes samples from Gehrels *et al.* (2001) and White (2001) as well as new low marsh and mudflat samples from the Salcombe-Kingsbridge estuary collected to increase the vertical range of the existing data set.

The total number of foraminifera samples from contemporary inter-tidal zones in south Devon is 113. The Erme salt marsh and mudflat data set consists of 8 new samples supported by an additional 54 samples from two transects (Gehrels *et al.*, 2001), and a further 30 samples from White (2001). The Frogmore Creek data set consists of 21 samples, down to local mean low water level of spring tides. The entire data set represents modern inter-tidal foraminifera samples found between +2.77 and –2.54 m MTL. Sample heights and main plant zones at the Erme estuary and Frogmore Creek in the Salcombe-Kingsbridge estuary are shown in Figure 3.2.

3.2 Site descriptions

3.2.1 The Erme estuary

The mouth of the River Erme is located at SX 6125 4710 and is one of the smallest south Devon rivers (Figure 3.1). The sedimentary environments in the Erme estuary comprise sandy beaches, tidal channels, tidal flats, salt- and freshwater marshes. Cliffs of Lower Devonian Dartmouth slates and periglacial head deposits rise above the inter-tidal zone (Creak, 1991). The estuary contains approximately 20.75 ha of estuarine fringing salt marsh (Burd, 1989) but the inter-tidal marsh sampled in this survey covers only 0.6 ha (Gehrels *et al.*, 2001). The entire catchment extends inland about 23 km (National Rivers Authority, 1993) and the open coast is about 2.5 km from the marsh site.

The Erme marsh (SX 6205 4928) is characterised by distinctive vegetation zones (Figure 3.2, Massey, 1998). A range of plant sub-environments is found on the fresh and saltwater marsh surfaces, the most dominant of which are described here. The uppermost terrestrial environment contains small deciduous trees and shrubs. The high marsh zone is dominated by *Phragmites australis* and bordered by *Cyperaceae* spp., and *Juncus maritimus*. The middle marsh consists of predominantly *Puccinellia maritima* and the low marsh contains a mixture of *Puccinellia maritima*, *Halimione portulacoides* and *Spartina anglica*.

3.2.2 The Salcombe-Kingsbridge estuary

The estuary entrance is located at SX 7450 3900 (Figure 3.1) and, in the strictest geomorphological sense, the entire feature is not an estuary as there is no river flowing into it (Spooner, 1979). Hiscock (1986) suggested that the several drowned river valleys, now creeks, of the Salcombe-Kingsbridge estuary were originally formed by the River Avon. The estuary extends 8.3 km inland covering an area of 674 ha, of which 446 ha is inter-tidal (Fahy *et al.*, 1993). The sedimentary environments are predominantly sand and mudflat bordered by low marsh and steep cliffs of altered Devonian rock types (Ussher, 1904; Durrance and Laming, 1982; Born, 1986). The estuary contains 4 ha of estuarine fringing salt marsh (Fahy *et al.*, 1993) but the inter-tidal zone sampled in this survey is predominantly fine to coarse silty-clayey-sand and gravel covering approximately 3.9 ha. The open coast is about 4 km from the mudflat site and low wave and wind energy promote high sediment deposition especially in the more shallow upper estuary creeks, e.g., Frogmore Creek (BMT Environment and Adams Hendry for English Nature, 1994).

The cove containing the sample sites is situated near the entrance to Frogmore Creek (SX 7545 4068) and is entirely inter-tidal mudflat (Figure 3.2) bordered by a narrow zone of *Spartina anglica* and *Aster tripolium.* It is almost completely surrounded by a rocky shoreline covered in the ancient mixed deciduous woodland of Halwell Wood and Wall Park (Figure 3.1). Frogmore Creek is over 3 km in length, the longest tributary of the estuary, and contains salt marsh in its upper reaches (BMT Environment and Adams Hendry for English Nature, 1994). The flora in these fringing salt marshes is predominantly *Aster tripolium, Puccinellia maritima* and *Plantago maritima.*

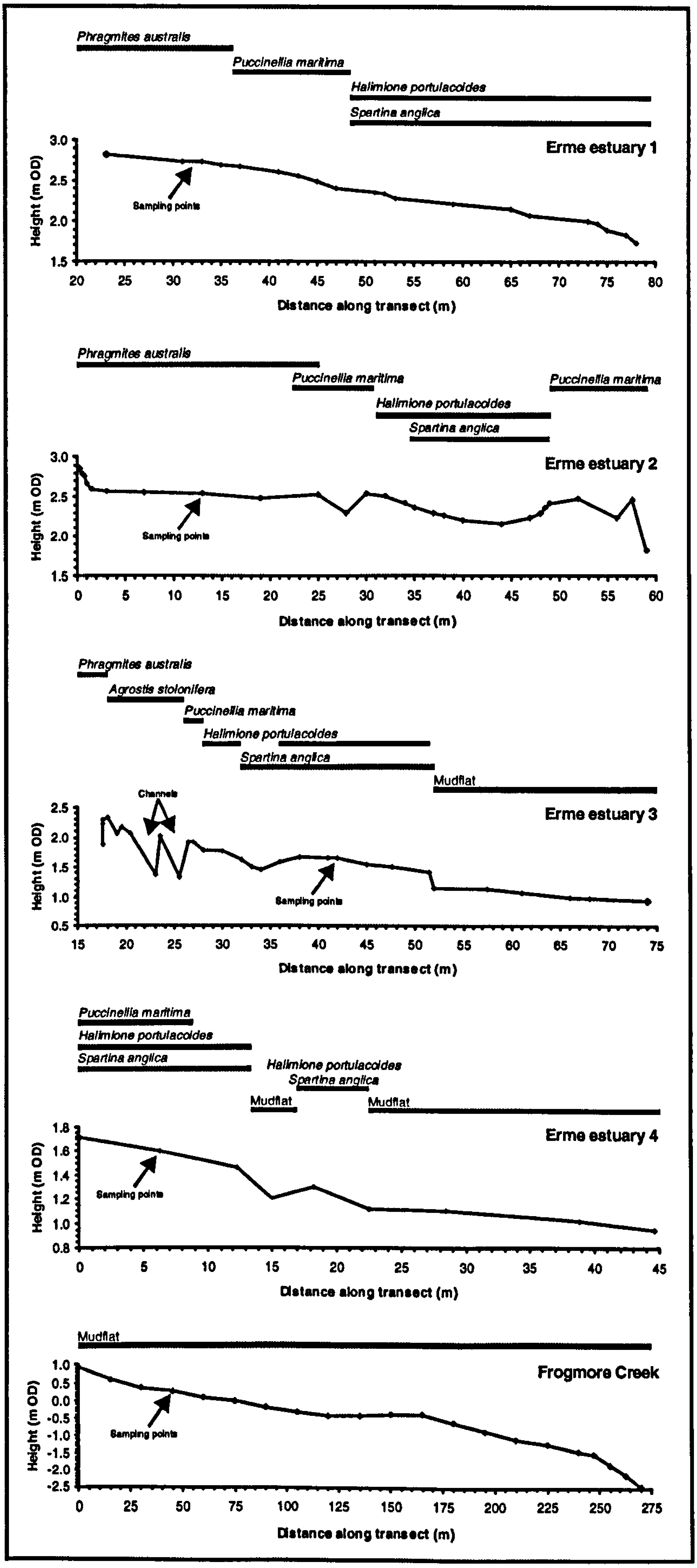


Figure 3.2. Cross-sections from the Erme estuary and Frogmore Creek. Changes in height, main plant and mudflat zones along Erme estuary 1 and 2 (Gehrels *et al.*, 2001); Erme estuary 3 (White, 2001); Erme estuary 4 and Frogmore Creek (this project) transects.

3.2.3 Tidal regime

Tidal heights for the Erme and Salcombe-Kingsbridge estuaries are shown in Table 3.1. Heights are interpolated from the River Yealm and Salcombe Harbour tidal station data (Hydrographic Office, 2002). Sample sites from the Erme estuary are on inter-tidal marsh and mudflat situated between +1.8 and +1 m MTL and Frogmore Creek samples are on mudflat sediment between +1 and –2.5 m MTL. Gehrels' *et al.* (2001) Erme salt-marsh data ranges from around +2.7 to +1.6 m MTL and White's (2001) Erme data is also predominantly salt marsh (+2.4 to +1 m MTL).

Tidal Station / Site	Latitude N	Longitude W	Distance (m)	MHWST (m)	MLWST (m)	Tidal Range (m)	HAT (m)	Chart Datum (m)	MSL (m)	MTL (m)
River Yealm (entrance)	50° 18'	4° 04'	8600 west	2.35	-2.35	4.7	2.75	-3.05	3.20	0.15
Salcombe Harbour	50° 13'	3° 47'	11425 east	2.25	-2.35	4.6	2.65	-3.05	3.14	0.09
Erme Estuary (entrance)	50° 18'.55	3° 56'.64	-	2.31	-2.35	4.66	2.71	-3.05	3.17	0.12

Table 3.1. Tidal heights for field sites interpolated from nearest tidal stations. Heights for Frogmore Creek are the same as for Salcombe Harbour. All heights are relative to the UK Ordnance Datum. Distance – approximate distance between the Erme Estuary and Tidal Station (using eastings). Source: Hydrographic Office (2002).

3.3 Methods

3.3.1 Field techniques

Fieldwork was carried out at Frogmore Creek in October 2000 and the Erme estuary during June 2001. The sites are accessible by nearby track or road and OS benchmarks are within 300 m of the field transects (Figure 3.1). Field sampling methods are based on Gehrels *et al.* (2001). Sediment samples were collected with 7.5 cm diameter by 5 cm deep (Pitman Corer) tins sealed at both ends with plastic lids. Surveying was carried out, at the time of sample collection i.e., MLWST, according to the methods described in section 2.2.2. A vertical sampling resolution (the height differential between consecutive samples) of generally <0.3 m and <0.2 m was used along the Frogmore Creek and Erme 4 transects respectively (Figure 3.2). Gehrels *et al.* (2001) achieved around 0.05 m, and not >0.3m, along Erme 1 and 2, and White (2001) generally used <0.4 m for Erme 3. Greater height differentials occur where transects crossed channels in the marsh or mudflat surface, e.g., Erme 3 (Figure 3.2).

3.3.2 Laboratory techniques

A 5 ml sub-sample was removed from the top 1 cm slice of each sediment sample on the day of collection. These were then submerged in an ethanol and rose Bengal solution (consisting of a 10:1 dilution of 98% ethanol, 5 gms per litre of sodium bicarbonate and 5 gms per litre of rose Bengal) to stain living foraminifera. In studies of live and dead populations, the in-water examination of stained tests significantly aids the identification of species (Scott and Leckie, 1990). Samples were placed in cold storage before preparation for microscopy analysis using standard techniques (Scott and Medioli, 1980a, see 2.3.2) and numbers of live, dead and total foraminifera were recorded from each sample.

3.4 Results

3.4.1 Modern analogue of dead foraminifera

Murray (2000) suggests that the use of total foraminiferal assemblages is "hindering progress in understanding the full extent of taphonomic change between live and dead assemblages and therefore hindering the interpretation of fossil assemblages". Current palaeoecological practice dictates that the dead assemblage should be used as it more closely resembles the fossilised assemblage (Murray, 1976, 1982, 2000). This project therefore uses the dead data set and statistical analysis will show that the assemblage produces the most significant results (see section 3.4.2). Dead foraminifera are found along surface transects at the Erme estuary and Frogmore Creek between approximately –2.6 and +2.6 m MTL (Figure 3.3). A total of 34 species are identified of which 9 are agglutinated and 25 are calcareous. A number of tests at unidentifiable stages of development or condition were recorded. A stratigraphically constrained CONISS cluster analysis (Grimm, 1991) divides the dead assemblage into four main zones.

The highest zone (+1.25 to +2.55 m MTL) is dominated by *M. fusca*, *T. inflata* and *J. macrescens*, with minor occurrences of *Haplophragmoides* spp. The next zone (around +0.05 to +1.25 m MTL) is dominated by a combination of agglutinated and calcareous species. *M. fusca* and *J. macrescens* share the zone with high numbers of *Ammonia beccarii*, *Elphidium* spp., and *Haynesina germanica*. The zone between –0.65 and +0.05 m MTL is preferentially occupied by *Reophax* spp., *Eggerella scabra* and *Elphidium oceanensis*. The diversity and percentage abundance of calcareous foraminifera notably increase in this zone. Data from the lowest zone (–2.55 to –0.65 m MTL) are almost exclusively derived from Frogmore Creek. It is dominated by most calcareous species, in particular *A. beccarii* and *H. germanica* with significant occurrences of species such as *Asterigerinata mamilla*, *Brizalina variabilis*, *Elphidium margaritaceum* and *Trifarina angulosa*. *T. ochracea* and *J. macrescens* also occur in significant numbers in the lowest zone.

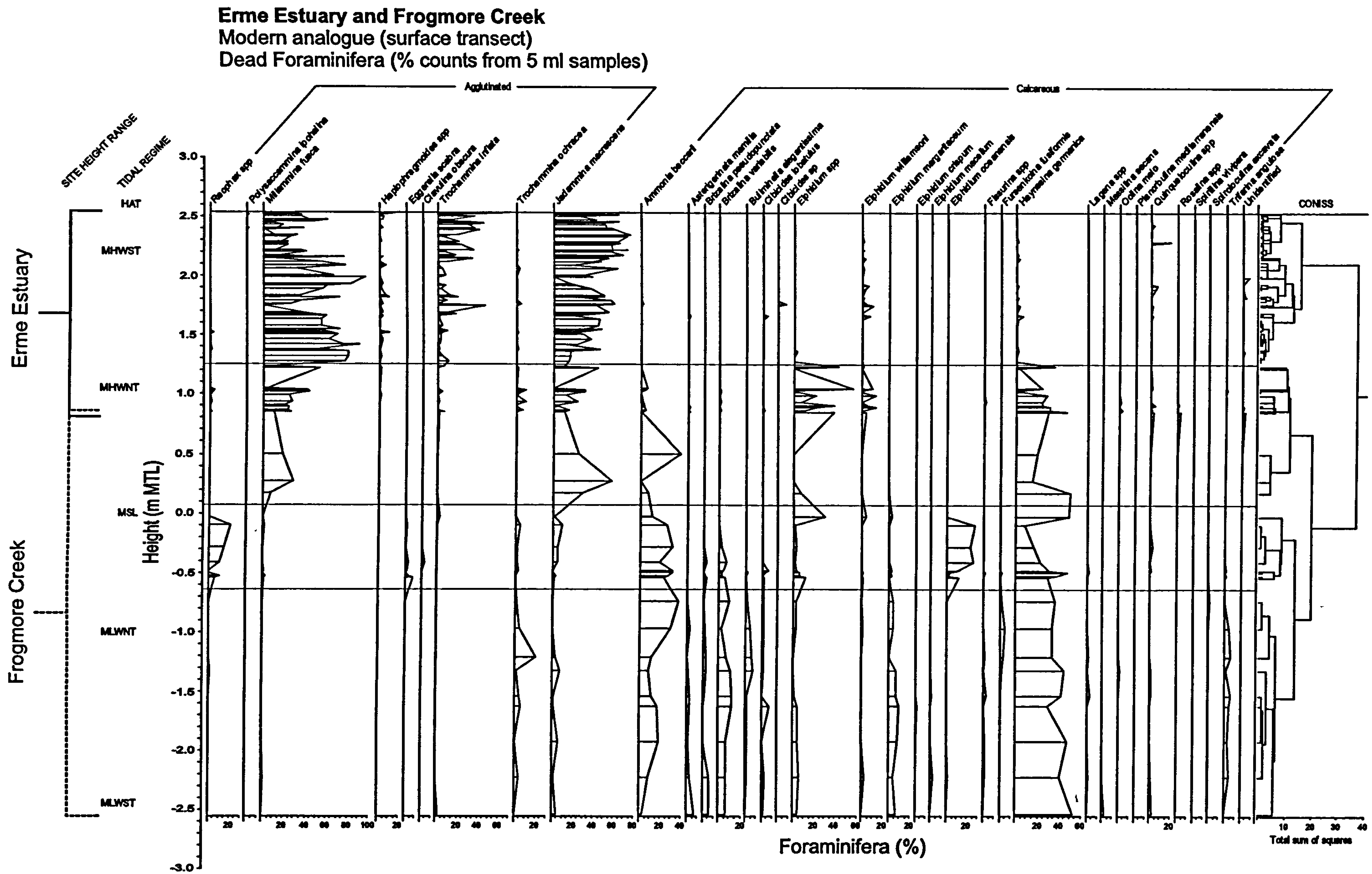


Figure 3.3. Modern surface analogue of dead foraminifera. A CONISS cluster analysis (Grimm, 1991) identifies four principal zones. Some species are combined to a single generic classification (see section 2.3.2 and Appendix 7), e.g., *Elphidium* spp., includes *E. earlandi*, *E. gerthi* and *E. incertum* (Totals shown in Appendix 7a).

3.4.2 Quantitative analyses

Samples with low counts (n < 30: an arbitrary cut-off) are excluded from quantitative analyses and the data set was divided into three training sets of dead, live and total (dead + live) foraminifera. Raw and percentage counts of live and dead foraminifera are included in the CD-ROM (Appendix 7). Total counts for Erme estuary 1 and 2 (Gehrels *et al.*, 2001), Erme 3 (White, 2001) and Erme 4 and Frogmore Creek (this project) samples are also shown in Appendix 7. The relationship between foraminiferal assemblages and MTL is explored using multivariate statistical analyses. This was assessed by performing a Detrended Canonical Correspondence Analysis (DCCA), using the CANOCO programme (Ter Braak, 1995), to provide an estimate of the gradient length in standard deviation (SD) units (Birks, 1995) and determine the most suitable model for the training sets. The majority of species optima will be located within the gradient when gradient lengths are >2 SD units. In this case the most appropriate regression and calibration technique to use is a unimodal response model. Linear response models are used when the gradient length is <2 SD units (Birks, 1995). DCCA calculates gradient lengths of >3 SD units indicating the unimodal nature of all training sets (Table 3.2). Indicative meaning can be usefully reconstructed as 25 to 30% of the variance is explained by axis 1 for species data, and 80 to 90% for species-environment (elevation: in m MTL) data. Species variance explained seems low but compares favourably with Gehrels *et al.* (2001) and Horton *et al.* (1999a,b). Eigenvalues of 0.7 show that the data are well correlated (Table 3.2).

				Axis 1 % variance explained		
Training set	*N*	n	Axis 1 length	(of species data)	(of species-environment relationship)	Axis 1 eigenvalue
Dead	88	33	3.575	28.3	87.3	0.689
Live	90	30	3.432	24.8	82.4	0.647
Total	99	35	3.756	29.7	88.0	0.681

Table 3.2. DCCA on training sets of dead, live and total foraminifera. Data are from Appendix 7. *N* = number of samples; n = number of taxa.

The CALIBRATE programme (Juggins and Ter Braak, 1998) is used to calculate regression statistics for models. Unimodal regression models used here include WA (weighted averaging), WA-Tol (weighted averaging with tolerance downweighting) and WA-PLS (weighted averaging partial least squares). The main linear model is PLS (partial least squares, results are not included here because gradient length >2 SD units, Table 3.2). WA-PLS combines the preferential features of inverse linear regression and WA regression (high correlation) and of stable predictors of high variance, i.e., PCR (principal components regression) and CAR (correspondence analysis regression). WA-PLS is recommended for species-rich data where gradients are >2 SD units (Birks, 1995).

The regression statistics produced by the CALIBRATE programme include the root mean square error prediction (RMSEP: vertical error of the SLIP in metres relative to MTL), the coefficient of determination (r^2: indicative meaning, height in metres relative to MTL), and values for the maximum bias (Table 3.3). The difference between observed and predicted values of the training sets (RMSEP) is around ±0.3 to ±0.4 m MTL. WA-PLS on dead foraminifera produces the smallest vertical error margin (±0.29 m MTL). The proportion of variance explained by the regression models (r^2) is very high (0.85 to 0.93). The best measure of the potential accuracy for the reconstruction of indicative meaning of fossil assemblages is again produced by WA-PLS on the dead assemblage (0.93). This combination also produces the least maximum bias, or difference between mean observed and predicted values along the gradient (Birks, 1998), of 0.29. Regression results show that the WA-PLS response model is the most appropriate one to use with the dead modern analogue data sets.

Training set	N	n	Regression model	RMSEP (m MTL)	r^2 (m MTL)	Max-Bias
Dead	88	33	WA	0.339	0.901	0.451
Live	90	30	WA	0.410	0.838	0.505
Total	99	35	WA	0.374	0.882	0.559
Dead	88	33	WA-Tol	0.328	0.907	0.378
Live	90	30	WA-Tol	0.395	0.850	0.555
Total	99	35	WA-Tol	0.371	0.884	0.502
Dead	88	33	WA-PLS	0.290	0.927	0.291
Live	90	30	WA-PLS	0.355	0.878	0.539
Total	99	35	WA-PLS	0.326	0.910	0.416

Table 3.3. Performance of the regression models. Results for training sets of dead, live and total foraminifera are shown. Data are from Appendix 7. N = number of samples; n = number of taxa; Models and terms are explained in the text.

Had these results showed that the dead assemblage is statistically weaker than the others it would remain the choice training set for palaeoecological reasons (Murray, 1976, 1982, 2000, see section 3.4.1). The relationship between this training set and elevation is very strong (r^2 = 0.93) but only marginally more so than both live (r^2 = 0.88) and total (r^2 = 0.91) assemblages. It is apparent from Table 3.3 that both WA (r^2 = 0.90) and WA-Tol (r^2 = 0.91) models also show that the relationship between the dead assemblage and elevation is strong. However, as the RMSEP (0.29) and Maximum Bias (0.29) are lowest from WA-PLS analysis of dead foraminifera and the gradient length is >2 SD units WA-PLS is the chosen combination for the reconstruction of indicative meaning of fossil foraminiferal assemblages.

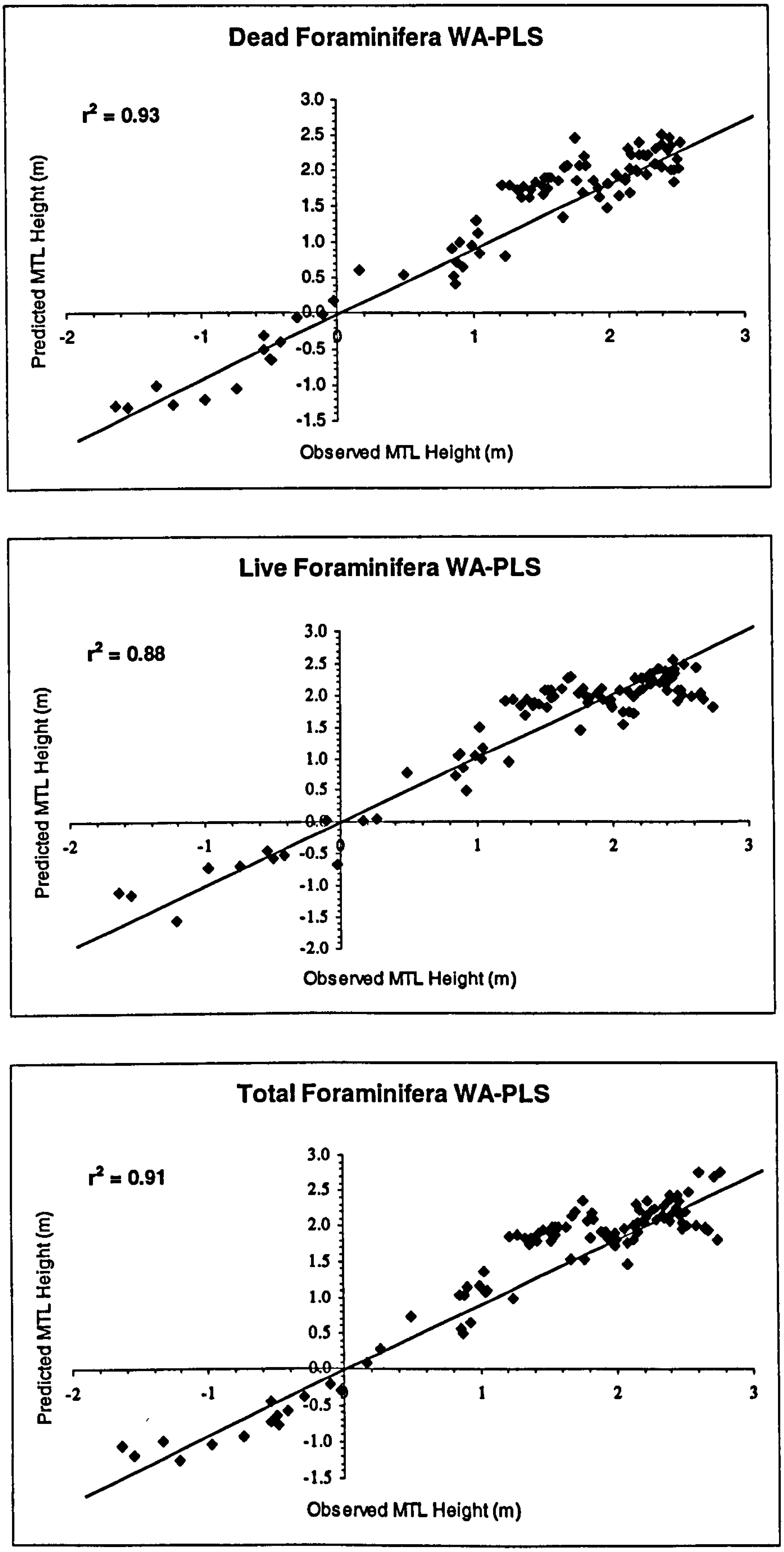


Figure 3.4. Observed and predicted values of indicative meaning (m MTL). Results include data for dead, live and total foraminifera using the WA-PLS regression model (the best predictor of the three models: Table 3.3) and show the r^2 value indicating the ability of the training set to accurately predict indicative meaning in fossil samples presented in chapters 4 to 7.

3.5 Discussion and interpretation

Whilst live populations are generally the focus of ecological investigations it has been argued that this training set does not provide a representative analogue for the fossil record (Murray, 2000). This is because the dead assemblage is less likely to alter, compared to the live and total assemblages, and is therefore a more representative analogue for fossil foraminiferal assemblages. The inclusion of the total (live + dead) assemblage in training sets has fuelled similar debate. Scott and Medioli (1980b) argued that total modern populations provide a better representation of the fossil record. Murray (2000) opposes this suggesting that as total assemblages include the live population they hinder the interpretation of fossil populations. Also, post-mortem assemblages may be significantly altered when all the calcareous tests are lost through dissolution leaving the dead assemblage comprised of 100% agglutinated foraminifera (Murray, 2000). Horton (1999) supports the view that the dead assemblage is a better modern analogue because it minimises the influence of seasonal fluctuations. The dead assemblage represents an integration of the foraminiferal population over a longer time period (several months to a year) and hence is more analogous to what is preserved in the sedimentary record. Figure 3.3 shows that the distribution of dead foraminifera is clearly related to elevation above or below mean tide level and the preference of dead over live and total assemblages is reinforced by statistical analyses (see section 3.4.2).

Fossil assemblages not contained in the modern analogue can occur when the former depositional environment is not represented in the surface training set (Edwards and Horton, 2000). For example, the indicative meaning of mudflat assemblages from Bantham Sands (Figures 4.11 and 4.13) shows that such a problem may exist with this data set. The number of deep-water marine foraminifera, e.g., *Elphidium macellum*, *Elphidium crispum* and *A. mamilla*, found in fossil sediments has influenced the calculation by the transfer function. This is because if deep-water foraminifera are over-represented in the fossil samples compared to the training set the transfer function will calculate an indicative meaning that is well below the range of modern samples. It is interesting to note that *Reophax* spp., an unusual species identified by Long *et al.* (1999), Edwards and Horton (2000) and Edwards (2001) in Poole Harbour, and *E. oceanensis* are good indicators of MTL, especially their upper limits (Figure 3.3). Furthermore, a continuum seems to exist in the zone between +0.05 and +1.25 m MTL, despite the fact that around half the samples are from the Erme and the other half from Frogmore Creek.

The modern assemblages below MTL increase the range of samples from which an indicative meaning can be established to almost –2.6 m MTL (Figure 3.3). Horton (1999) has investigated the distribution of modern inter-tidal foraminifera at Cowpen Marsh, Tees Estuary, NE England but his mudflat samples extend to around –0.7 m MTL only. Horton *et al.* (1999a) constructed a contemporary inter-tidal foraminiferal data set from 10 UK study areas (N = 165) standardising their tidal range to allow a between-site comparison to be carried out. The standard water level index (SWLI) used by Horton *et al.* (1999a) is a function of local MTL and MHWST and the transfer function for the SWLIs produced a r^2 value of 0.75 and RMSEP (WA model) of 12.47 cm. The RMSEP would have to be multiplied by the vertical difference between MTL and MHWST to obtain the RMSEP at any given locality. At the Erme, for example, MHWST–MTL is 2.19 m, while at Frogmore Creek this value is 2.16 m (Table 3.1), producing RMSEPs of 0.273 and 0.269 m respectively, if Horton's transfer function is applied to these sites. Results from this study are therefore comparable to those of Horton *et al.* (1999a).

3.6 Summary

The surface distribution of foraminifera from the Erme estuary and Frogmore Creek is significantly controlled by elevation in the inter-tidal frame. Four very distinctive zones are identified between approximately –2.6 and +2.6 m MTL containing predominantly agglutinated foraminifera in the uppermost zone and calcareous foraminifera in the lower two zones. The second highest zone is a mixture of agglutinated and calcareous foraminifera and there seems to be a continuum in the assemblage despite the fact that samples come from both the Erme and Frogmore Creek. A number of species exhibit a distinct preference for specific elevations, e.g., *Haplophragmoides* spp., prefer the uppermost 1.5 m of the inter-tidal zone, *Reophax* spp., *E. scabra* and *E. oceanensis* prefer conditions around MTL, and species such as *Buliminella elegantissima*, *E. margaritaceum* and *T. angulosa* prefer habitats below MTL. The upper limits of *Reophax* spp., and *E. oceanensis* appear to be particularly good indicators of MTL. The data set provided by Gehrels *et al.* (2001) has been enhanced by the addition of data from this project and White (2001). Whilst the data of Gehrels *et al.* (2001) span the vertical range between HAT and (approximately) MHW, the new data set now provides a modern analogue for the entire inter-tidal zone. The most powerful predictors of indicative meaning are provided by the dead assemblage ($r^2 = 0.93$) of the WA-PLS regression model. The RMSEP result (±0.29 m) for dead foraminifera provides the vertical error to be applied to the indicative meaning of fossil samples. However, it is recognised that this is a statistical error and errors may be greater due to the absence of a matching analogue and by low numbers of foraminifera present in fossil samples.

Chapter 4

Bantham Sands

4.1 Site description

Bantham Sands is located at SX 662 437 (Figure 4.1) and derives its name from the 'bents' or marram grass that have colonised the sand dunes found on Bantham Ham at the mouth of the River Avon (Jenkins, 1902; Mildmay-White, 1985). The site lies in the lower reaches of the Avon estuary, and the back-barrier marsh and valley slopes are used as rough pasture. The area also serves as an overflow car park during the summer months for recreational visitors. Bantham has been the focus of much archaeological research (Fox, 1864; Elliot, 1901; Jenkins, 1902; Fox, 1955; Silvester, 1981; Griffith, 1986; Griffith and Reed, 1998; Devon Archaeological Society, 2001) suggesting that the site has been occupied, probably as a coastal trading 'mart', since at least pre-Roman / Dark Age times (Table 4.1). Bantham is one of a number of sites around Devon that is being mapped to create a landscape archaeology model (Community Landscapes Project, 2003). However, prior to this study no other research has been carried out to reconstruct the post-glacial sea level history of Bantham Sands or its back-barrier system.

Radiocarbon laboratory number	Location (OSGR)	Sample height (m OD)	Material	^{14}C age (years BP)	Max (Median) Min Cal age (BP) (± 2σ range)	Reference
HAR – 5775	Bantham Ham SX 66260 43581	0	Bone	1690 ± 80	1818 (1584) 1409	Griffith (1986)
HAR – 5776	Bantham Ham SX 66260 43581	0	Charcoal	1440 ± 90	1526 (1322) 1178	Griffith (1986)
AA – 33125	Bantham Ham SX 6647 3790*	8	Charcoal *Ulex/Cytisus*	2950 ± 60	3325 (3128) 2895	Griffith and Reed (1998)

Table 4.1. Radiocarbon dates and associated sample information of archaeological finds at Bantham. *Approximate location. Sample heights are recorded as 'very approximate'.

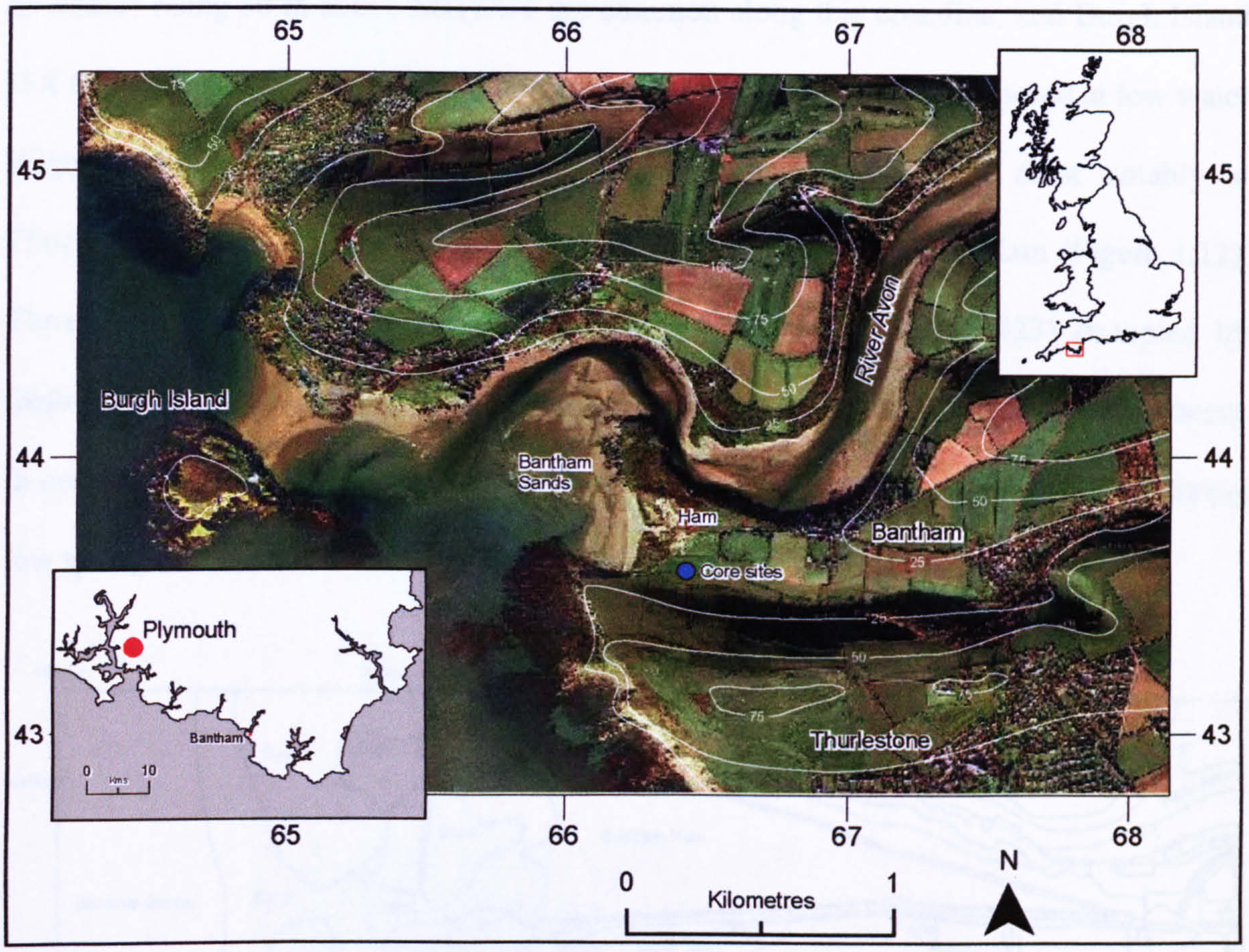


Figure 4.1. Aerial view of Bantham Sands. Source: http://www.multimap.com. Extract from Great Britain map and aerial photograph of South Hams District at scale 1:25 000 (2002). Heights in metres above MSL.

4.1.1 Geomorphology

The Bantham Sands back-barrier system is a freshwater peat marsh that is drained by Buckland Stream into Bantham Sand itself (Figure 4.2). It extends approximately 1 km inland from MHWST at 4 – 6 m above OD and covers about 10 ha. Buckland Stream and an artificial drainage system pass under sand dunes rising +15 m OD roughly 100 m west-north-west of the core site (Figures 4.1 and 4.2). Bantham Sands is a tidal flat and beach of roughly 30 ha on the eastern margin of Bigbury Bay below the sand dunes. It encroaches upon a rock shore platform below Hams End to the north and a rock spur to the south.

Sedimentary environments around Bigbury Bay include barrier and sandy beaches, tidal channels, mudflats and creeks, salt- and freshwater marshes. Stacks, caves, arches and headlands rising 50 m above MHWST are common along this coastline, and Burgh Island (SX 647 439) is currently joined to the mainland by a sand bar that is exposed at low water (Figure 4.1). Other local examples of back-barrier systems exist, most notably at Thurlestone Sands, situated about 2 km south-south-east of Bantham Ham (Figure 1.12). Three discordant valleys (SX 680 426, SX 684 421 and SX 681 423) occupied by freshwater peat marshes drain into Bigbury Bay, and were former inter-tidal environments. In conjunction with this system a submerged forest and peat bed remnant is visible during low spring tide at SX 677 417 on Thurlestone Sand (Pengelly, 1866b).

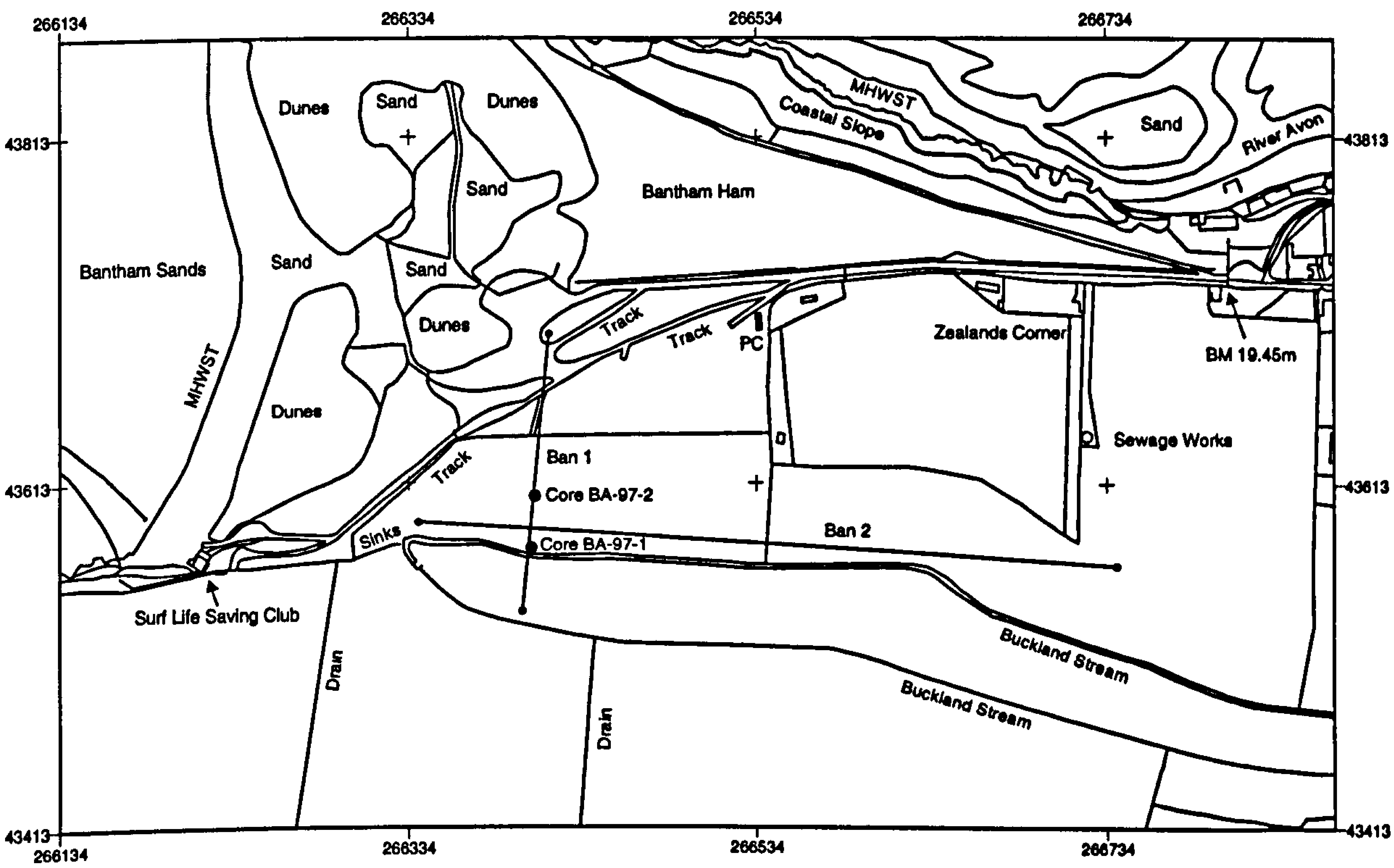


Figure 4.2. Bantham Sands back-barrier system. Borehole locations: cores BA-97-1 and BA-97-2. Electrical resistivity survey lines: Ban 1 and Ban 2. Source: http://digimap.edina.ac.uk/service/advanced/carto_welcome.html. Map extract at scale 1:2500 (2002).

4.1.2 Geology

The area containing Bantham Sands is formed of Devonian rocks consisting mainly of slates with fine-grained grit beds, silty slates and shales, and argillaceous and fine silty siliceous interlaminations (Ussher, 1890, 1904; Orme, 1960; Dineley, 1961; Figure 4.3).

The solid geology of the Bantham Sands study site is the Ringmore Type Slate Series of the Meadfoot Group from the Lower Devonian. Beeson Grits are also found to the north and south of Buckland Stream and the River Avon crosses a band of Dartmouth slates. The Quaternary deposits of the Bantham coastline include aeolian sand, river gravel and head, and alluvial sediments (Ussher, 1904). The area has been subject to widespread denudation and the shoreline of Bigbury Bay contains examples of Devonshire Trias outliers reminiscent of a previous Devonian rock surface (Pengelly, 1864). The Lower Devonian Dartmouth Group in Bigbury Bay is a predominantly fluvial mudstone probably deposited in a terminal fan on a coastal mud-plain (Smith and Humphreys, 1991).

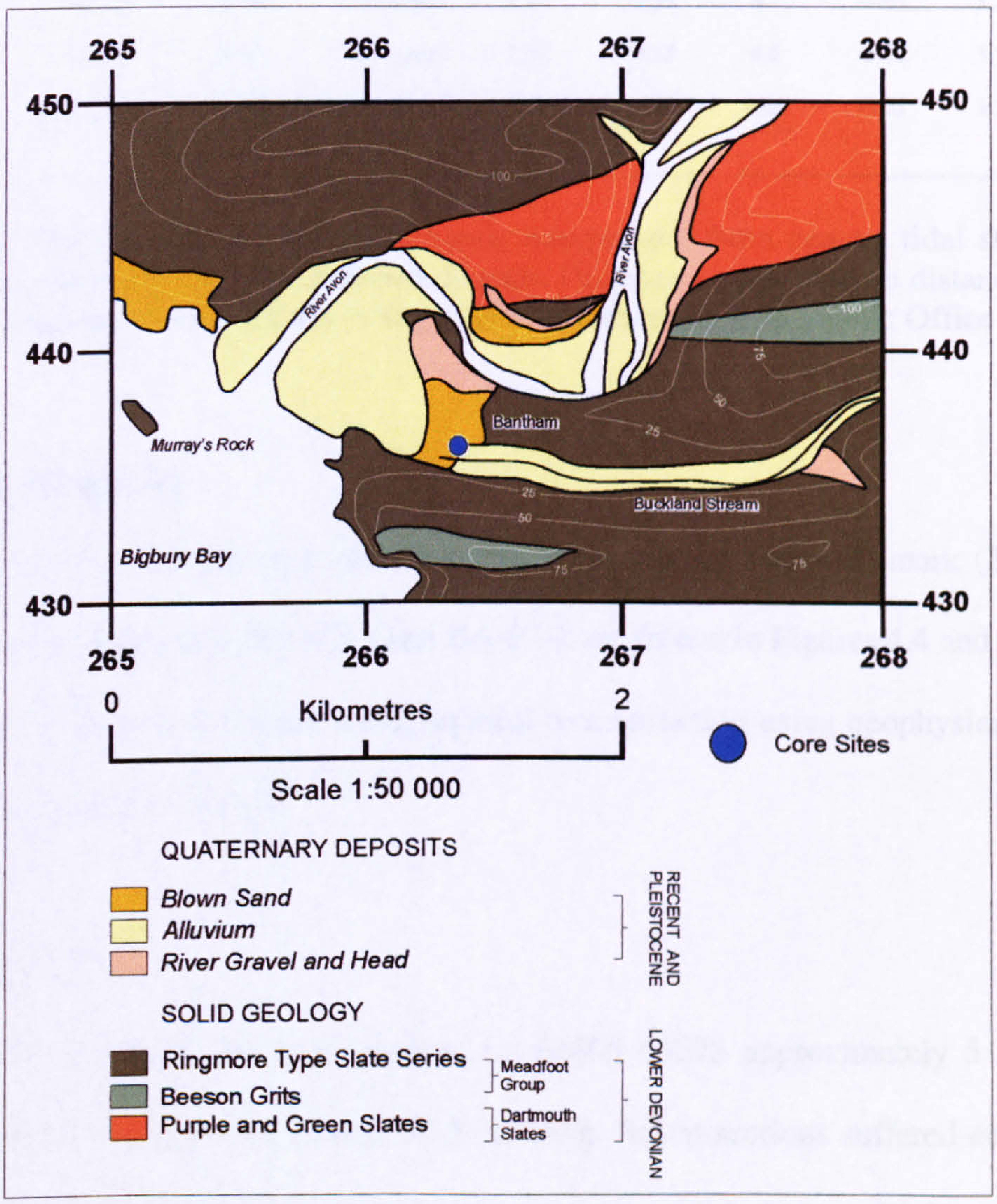


Figure 4.3. The solid geology and Quaternary deposits of Bantham Sands and the surrounding country. Source: Geological Survey of Great Britain (England and Wales) Sheets 355 and 356, Kingsbridge and Start Point, 1:50 000 (1975). Heights in metres above MSL.

4.1.3 Tidal regime

Tidal heights for Bantham Sands are shown in Table 4.2. Heights are interpolated from the River Yealm and Salcombe Harbour tidal station data (Hydrographic Office, 2002). The borehole sites are on marshland situated between +3.8 and +4.5 m above MTL and are protected from prevailing storm surges by the dune system (Figures 4.1 and 4.2). Flooding of the Bantham back-barrier freshwater marsh system is caused by a rise in groundwater levels and hillslope runoff following heavy rainfall. Buckland stream and adjacent drainage systems along the valley floor allow quick passage of the storm water to Bigbury Bay.

Tidal Station / Site	Latitude N	Longitude W	Distance (m)	MHWST (m)	MLWST (m)	Tidal Range (m)	Chart Datum (m)	MSL (m)	MTL (m)
River Yealm (entrance)	50° 18'	4° 04'	12925 west	2.35	-2.35	4.7	-3.05	3.20	0.15
Salcombe Harbour	50° 13'	3° 47'	7175 east	2.25	-2.35	4.6	-3.05	3.14	0.09
Bantham Sands (River Avon entrance)	50° 16'.80	3° 53'.13	-	2.29	-2.35	4.64	-3.05	3.16	0.11

Table 4.2. Tidal heights for Bantham Sands interpolated from nearest tidal stations. All heights are relative to the UK Ordnance Datum. Distance – approximate distance between Bantham Sands and Tidal Station (using eastings). Source: Hydrographic Office (2002).

4.2 Lithostratigraphy

Two cores were extracted approximately 210 m east of the MHWST mark (Figure 4.2). Detailed results from cores BA-97-1 and BA-97-2 are shown in Figures 4.4 and 4.5 and are correlated in Figure 4.6. Further stratigraphical reconstruction using geophysical survey is depicted in Figures 4.7 to 4.9.

4.2.1 Core BA-97-1

Core BA-97-1 (Figure 4.4) is located at SX 66405 43575 approximately 5 m north of Buckland Stream (Figure 4.2) and is 16.50 m long. Seven sections suffered core loss and eleven sections included bagged sediment. The basal lithofacies (16.50 to 16.40 m) consists mainly of fractured dark greyish-brown slates in a dark reddish-brown clayey-silty-sand. This is overlain by dark greyish-brown gravelly wet mud with organic remains

(16.40 to 15.90 m). The in-core lithofacies is dark grey fine sandy-silty-clay with shells (15.90 and 12.00 m) overlain by a minerogenic-wood peat facies (12.00 to 11.65 m). A predominantly silty-sand with large shell fragments lies between 11.65 and 6.10 m becoming more gravelly between 6.10 and 3.71 metres. The uppermost lithofacies of coarse sand and gravel contains no shells (3.71 to 3.62 m). This gravel facies is overlain by a wood peat with *Phragmites australis* remains, sand and gravel (3.62 to 0.30 m). The top of the core (0.30 m to ground level) consists of a sandy-silty peat-clay.

4.2.2 Core BA-97-2

Core BA-97-2 (Figure 4.5) is located at SX 66407 43605 approximately 30.4 m north of core BA-97-1 (Figure 4.2) and is 17.30 m long. Eight sections suffered core loss and eleven sections included bagged sediment. The basal lithofacies (17.30 to 17.13 m) of fractured light grey slates in a weak red silty-clayey-sand is overlain by a thin layer of wood and sandy-silty-clay with shells (17.13 to 17.05 m). A thin layer of sand and gravel is situated between 17.05 and 16.98 metres. The in-core lithofacies consists of dark grey sandy-gravelly-silty clay with shells (16.98 to 12.00 m) interrupted by dark olive grey gravelly-sandy-silty peat-clay (14.25 to 14.00 m and 13.46 to 12.69 m). The minerogenic sequence then becomes sandier (12.00 to 8.17 m) and more gravelly (8.17 to 5.51 m). The uppermost gravel facies (5.51 to 5.29 m) is devoid of shells and is overlain by a silty-sandy wood peat (5.29 to 4.00 m). A humified peat (4.00 to 2.32 m) is covered by very fine sand and shell fragments (2.32 to 0.20 m). The top of the core (0.20 m to ground level) consists of clayey-sandy-silt and peat.

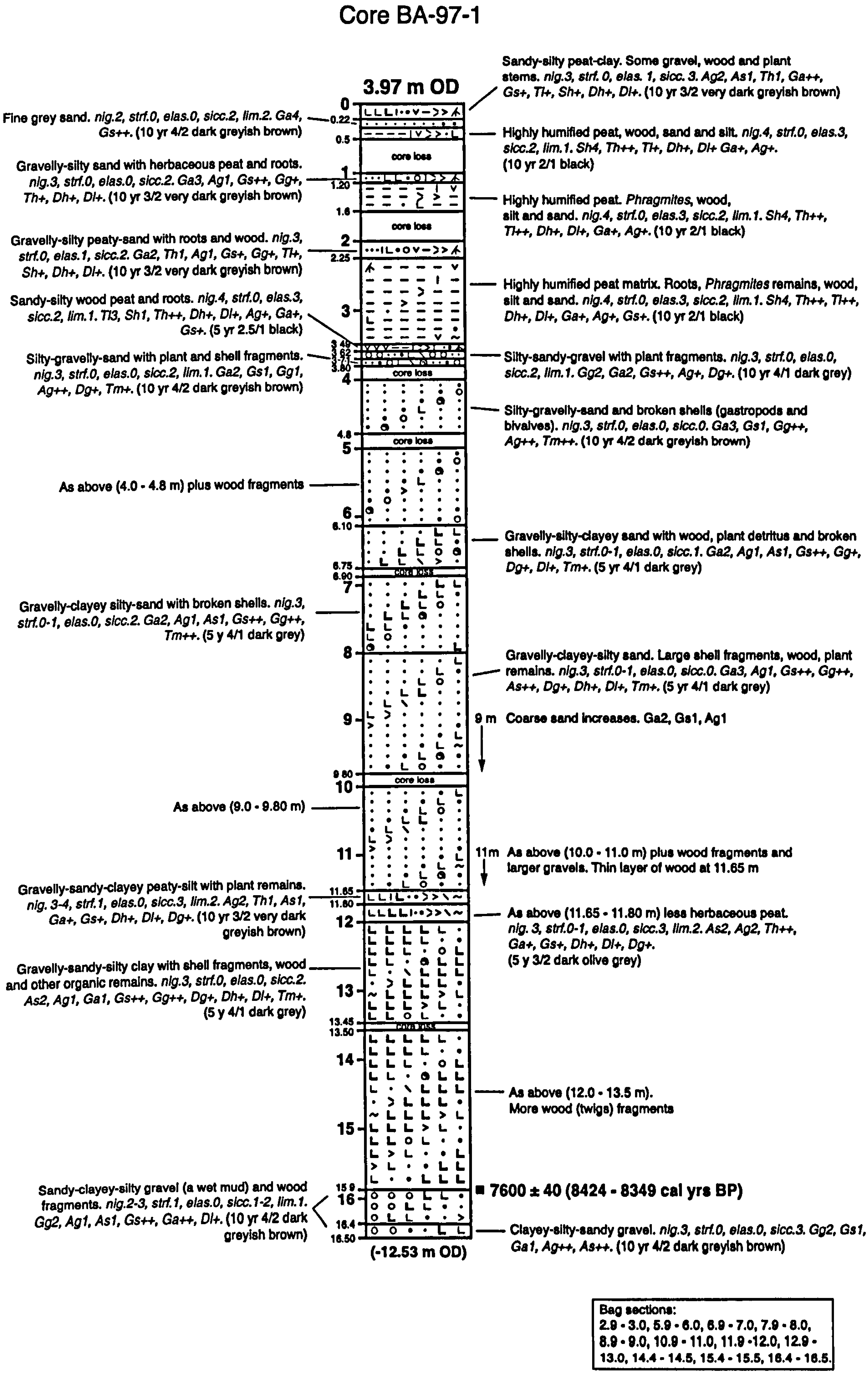


Figure 4.4. Lithostratigraphy of core BA-97-1.

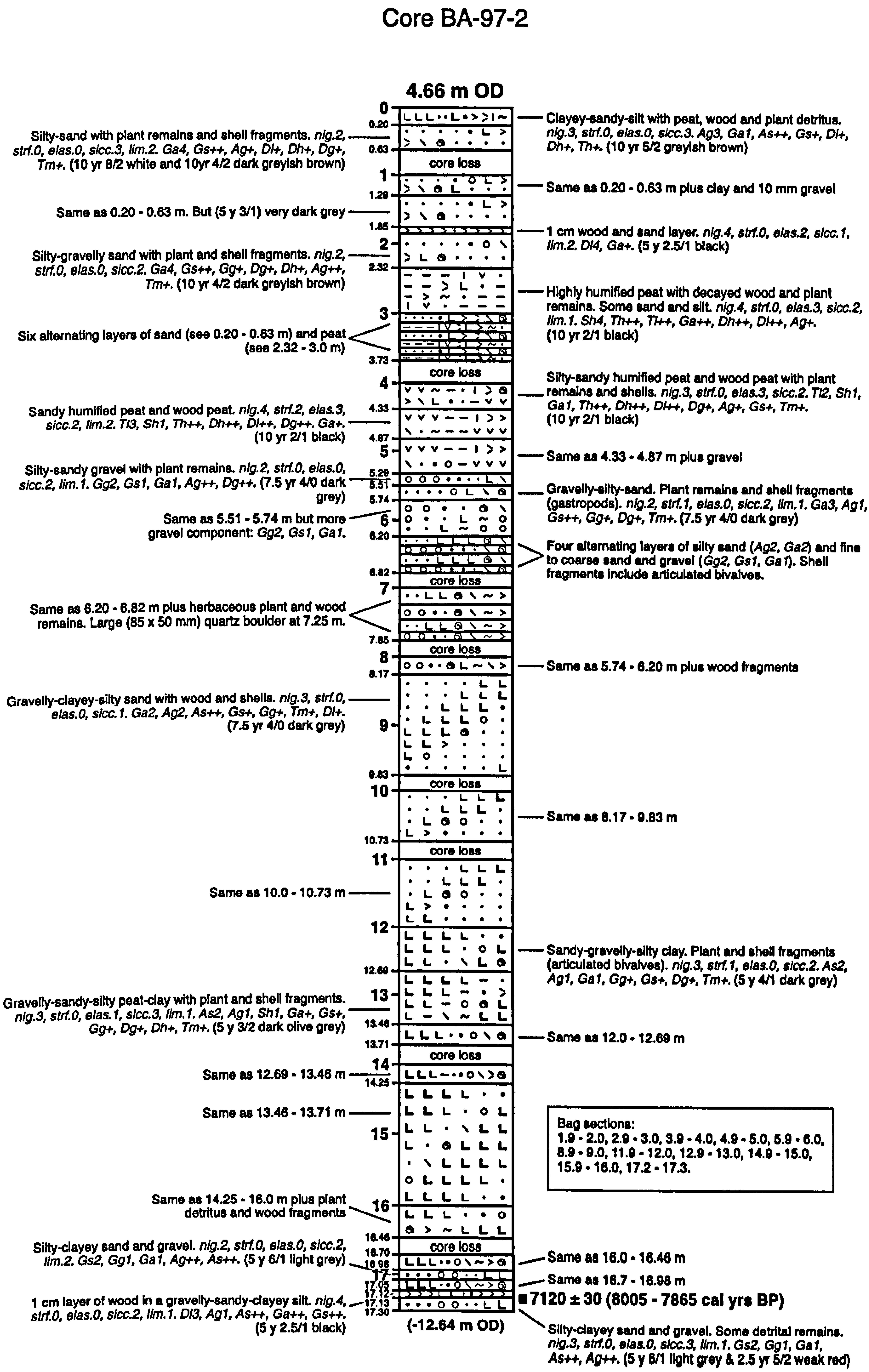


Figure 4.5. Lithostratigraphy of core BA-97-2.

4.2.3 Core correlation

The lithostratigraphic facies described in sections 4.2.1 and 4.2.2 are correlated in Figure 4.6. The basal lithostratigraphy of Bantham Sands consists of fractured slate below –12.50 m OD. Above this, to –7.0 m OD, a predominantly clay sequence is interrupted by minerogenic-peat units of greater thickness in core BA-97-2. This is overlain by a fine silty-sand sequence below –2.0 m OD that becomes more gravelly to +0.35 m OD. The minerogenic phase at Bantham Sands is then succeeded by an organic one as wood and highly humified peat dominate the uppermost lithostratigraphy, except in core BA-97-2 where a significant layer of fine sand is found above +2.34 m OD.

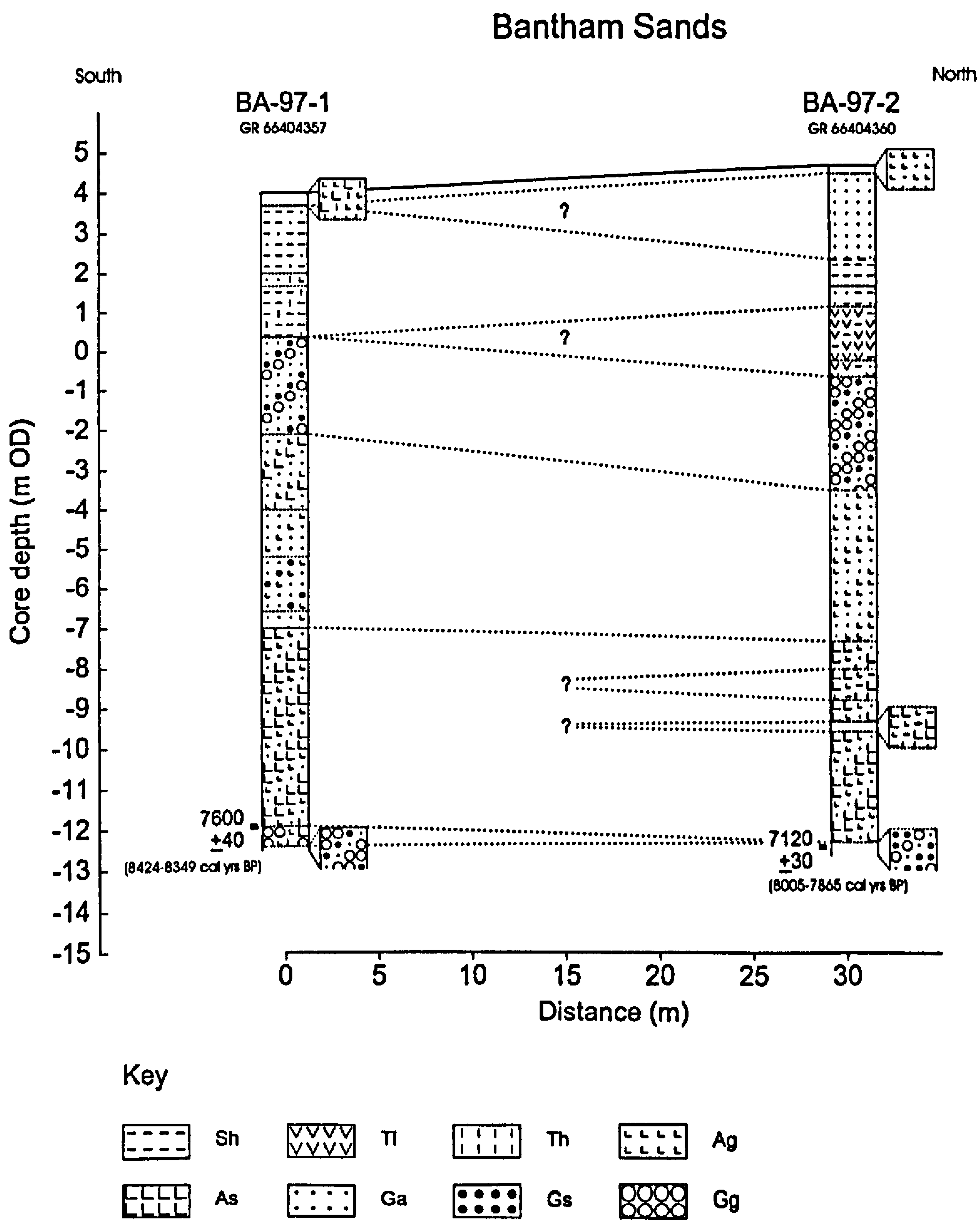


Figure 4.6. Bantham Sands core correlation.

4.3 Electrical resistivity survey

A shore-parallel (Figure 4.7) and shore-normal (Figures 4.8 and 4.9) electrical resistivity survey provide additional subsurface mapping information. Cores BA-97-1 and BA-97-2 are used as ground-truths for the survey. The geographical locations of the survey lines are shown in Figure 4.2. A summary of the electrical resistivity of sediments at Bantham Sands is given in Table 4.3.

4.3.1 Shore-parallel resistivity survey (Ban 1)

Ban 1 extends 180 m from the track at SX 66415 42700 (+6.94 m OD) to above Buckland Stream (drain) at SX 66400 43538 (+10.96 m OD). Height recordings were taken from 36 electrodes at 5 m intervals except where the survey line crossed gravel tracks (upper car park) or drains. A basin-type subsurface profile is delimited by the 150 Ωm isoline (Figure 4.7). However, basal gravel in both cores has resistivity values between 83 Ωm and 120 Ωm. The boreholes are located either side of the deepest point in the basin, over 3 m deeper than core BA-97-2. The deepest and shallowest points at which 83 Ωm occurs is –16 m OD at –10 m north of zero point and on the south slope 4.5 m below ground level respectively. Lower resistivity (<83 Ωm) is punctuated by peaks in higher resistivity at +12 and +23 m south of zero point. These peaks (–3 m OD) coincide with the locations of Buckland Stream and the drain (Figure 4.2). Higher resistivity (>83 Ωm) is recorded at approximately +2 m OD from –38 to –53 m north of zero point. This coincides with the surface gravel aggregate of the car park and tracks.

Determination of the electrical resistivity ranges of peat and fine minerogenic sediment from core BA-97-1 is complicated by interference from the stream at +12 m distance (see Dahlin and Owen, 1998). However, in cores BA-97-1 and 2 silty-clay records values of 32–83 Ωm. Sandy gravel has a resistivity of ≤68 Ωm and ≤32 Ωm in cores BA-97-1 and 2, respectively. Peat in both cores is ≤38 Ωm, and fine sand at the top of core BA-97-2 is 38–46 Ωm.

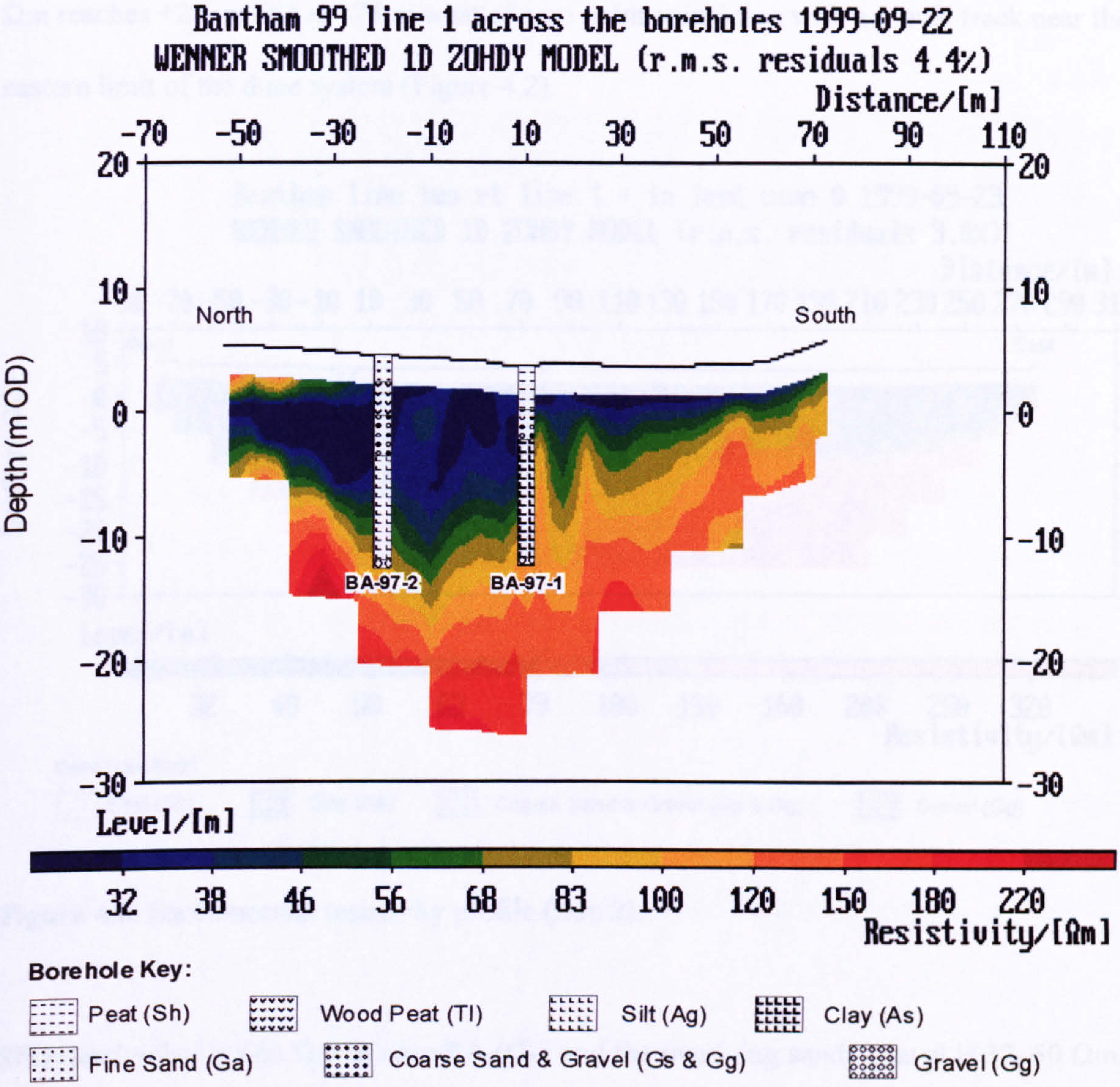


Figure 4.7. Shore-parallel resistivity profile (Ban 1).

4.3.2 Shore-normal resistivity survey (Ban 2)

Ban 2 extends 400 m from the rear of the dune system at SX 66340 43590 (+5.87 m OD) along Buckland Stream, about 15 to 22 m north of it, to SX 66740 43565 (+5.30 m OD). Height recordings were taken from 35 electrodes at 5 m intervals. A gently sloping subsurface profile is delimited by 320 Ωm rising west to east (Figure 4.8). The basal gravel in core BA-97-1 is 50–63 Ωm at zero point (Figure 4.9). Basal silt and clay (≤83 Ωm, section 4.3.1) is also present in the core (Figure 4.4). Core BA-97-1 is located in the deepest part of Ban 2 (50 Ωm) at –12 m OD around –10 m west of zero point. Isoline 50 Ωm rises steeply westward to –2.3 m OD reaching OD almost +280 m east. Isoline 100 Ωm reaches +2.5 m OD at –75 m west of zero point coinciding with a gravel track near the eastern limit of the dune system (Figure 4.2).

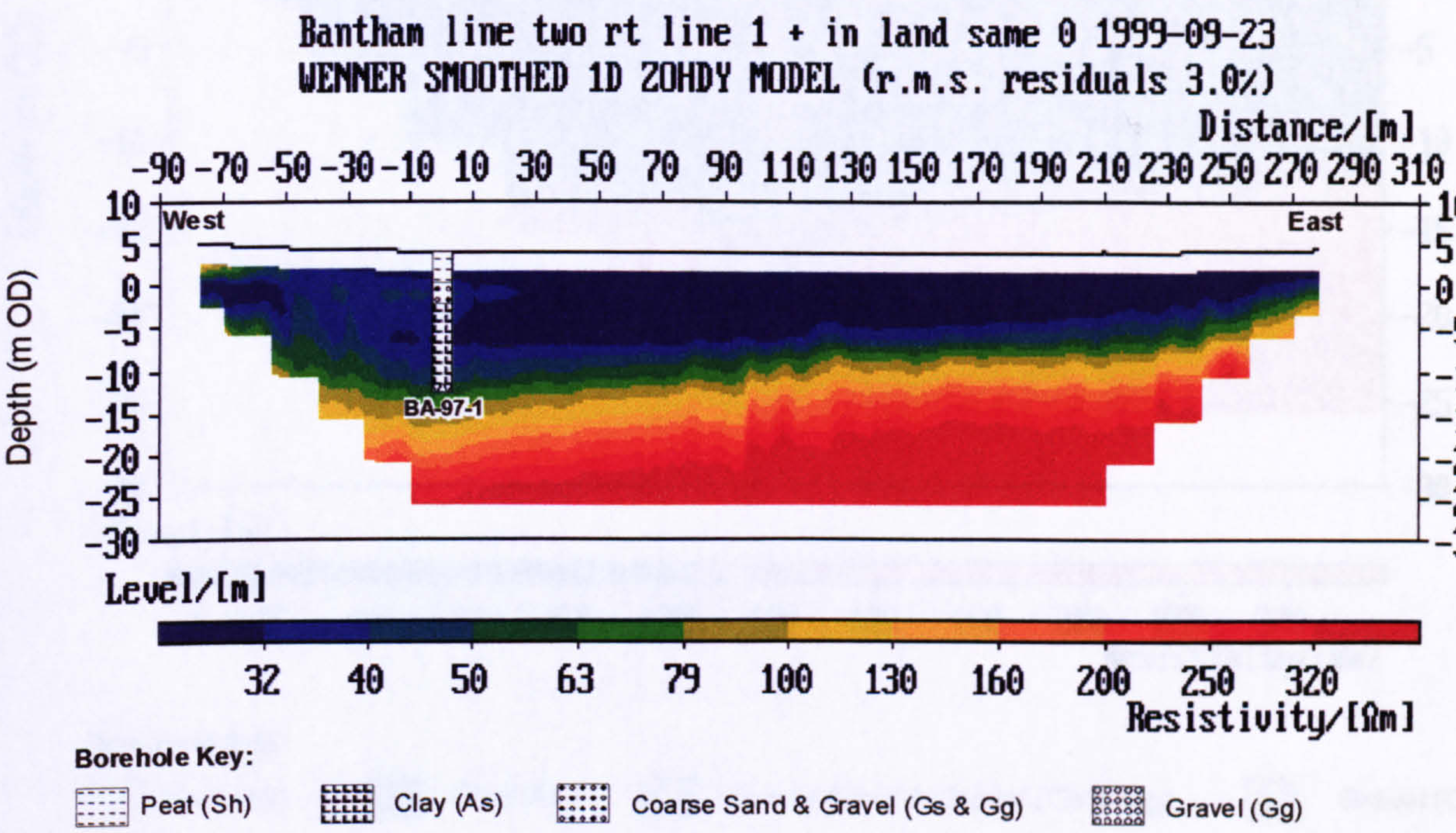


Figure 4.8. Shore-normal resistivity profile (Ban 2).

Silty-sandy-clay is ≤63 Ωm in core BA-97-1 and the overlying sandy-gravel is 32–50 Ωm. Figure 4.9 shows a marked 'tongue' of 32–50 Ωm into the 32 Ωm isoline at the borehole site and coincides with a sandy-gravel unit in the core. Peat is generally ≤40 Ωm but a resistivity of <32 Ωm to a depth of –7 m OD at zero point coincides with a minerogenic-peat layer in core BA-97-1 at –7.68 m OD (Figure 4.4).

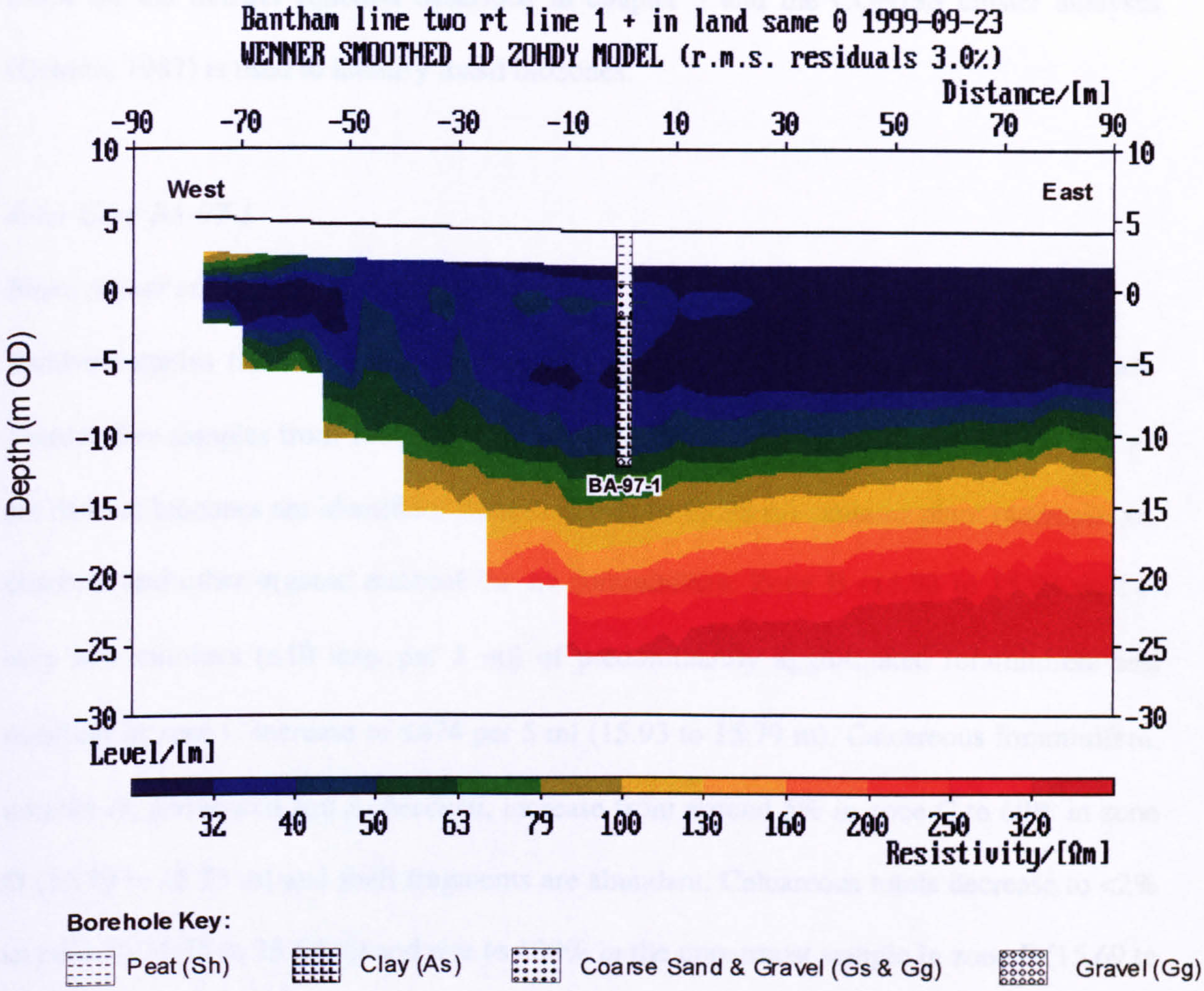


Figure 4.9. Enlarged view of shore-normal resistivity profile (Ban 2).

Survey	Basal gravel	Minerogenic fines	Peat
Shore-parallel (Ban 1)	83 – 120	≤83	≤38
Shore-normal (Ban 2)	50 – 63	≤63	≤40
Summary	50 – 120	≤83	≤40

Table 4.3. Summary of the electrical resistivity of sediments at Bantham Sands (units in Ωm). Minerogenic fines include sand, silt and clay. Peat also includes minerogenic-peat units.

4.4 Biostratigraphy

The biostratigraphy of Bantham Sands is partially reconstructed using foraminifera identified from the basal (lower contact) and in-core (upper contact) lithofacies of cores BA-97-1 and 2 (Figures 4.10 to 4.13). The indicative meaning of each sample is calculated based on the transfer function described in chapter 3 and the CONISS cluster analyses (Grimm, 1987) is used to identify fossil biozones.

4.4.1 Core BA-97-1

Basal (lower contact) section (Figure 4.10)

Twelve samples from 16.40 to 15.99 m and 15.97 to 15.95 m contain no foraminifera. Twenty-five samples from 15.99 to 15.97 m and 15.95 to 15.35 m contain foraminifera and six distinct biozones are identified. Zone A (16.40 to 15.98 m) contains plant macrophytes, charcoal and other organic material but no foraminifera. Zone B (15.98 to 15.93 m) has very low numbers (≤10 tests per 5 ml) of predominantly agglutinated foraminifera and numbers in zone C increase to ≤474 per 5 ml (15.93 to 15.79 m). Calcareous foraminifera, notably *H. germanica* and *A. beccarii*, increase from around 5% in zone C to 60% in zone D (15.79 to 15.75 m) and shell fragments are abundant. Calcareous totals decrease to <2% in zone E (15.75 to 15.69 m) and rise to 100% in the uppermost sample in zone F (15.69 to 15.35 m). The diversity of calcareous foraminifera increases four to fivefold over zone E and samples contain bivalves and gastropods.

In-core (upper contact) section (Figure 4.11)

Eight samples from 3.81 to 3.66 m contain foraminifera. Eight samples from 3.66 to 3.50 m contain no foraminifera and three distinct biozones are identified. Zone A (3.80 to 3.69 m) contains high numbers (500 to 700 per 5 ml) of predominantly calcareous foraminifera, particularly *A. mamilla* (40–60%). *Quinqueloculina* spp., *E. macellum* and *A. beccarii* comprise around 10–30% of the assemblage. The gravelly-sand facies (zone B: 3.69 to 3.65 m) contains very low numbers (≤20 per 5 ml) of primarily *A. beccarii*. Small numbers of mites (*Acari* spp.,) and midges (*Chironomidae* spp.,) are present in zone C (3.65 to 3.50 m) but no foraminifera. Samples contain *Phragmites australis* stems, freshwater bryophytes and other plant macrophytes.

Figure 4.10. Foraminifera stratigraphy (%) of core BA-97-1: basal (lower contact) section.

Figure 4.11. Foraminifera stratigraphy (%) of core BA-97-1: in-core (upper contact) section.
The transfer function has calculated an indicative meaning below the range of modern samples (see section 3.5).

4.4.2 Core BA-97-2

Basal (lower contact) section (Figure 4.12)

Twenty samples from 17.15 to 16.70 m contain foraminifera, one sample at 17.01 m contains no foraminifera and seven distinct biozones are identified. Zone A (17.15 to 17.12 m) is dominated by *A. beccarii* (50–60%) and *J. macrescens* comprises 32–16% of the assemblage. Calcareous species increase to >80% in zone B (17.12 to 17.09 m) and >90% in zone C (17.09 to 17.01 m). Sample counts in zone D (17.01 to 16.98 m) fall to ≤38 per 5 ml and agglutinates increase to 47% of the assemblage. Calcareous foraminifera dominate (75–97%) zones E (16.98 to 16.87 m) and F (16.87 to 16.82 m) and samples in zone F also contain moss leaves, *Acari* spp., and ostracods. Species diversity increases threefold at 16.70 m in zone G (16.82 to 16.70 m).

In-core (upper contact) section (Figure 4.13)

Seven samples from 5.60 to 5.47 m contain foraminifera. Thirteen samples from 5.46 to 5.00 m contain no foraminifera except one individual, *A. beccarii*, identified at 5.36 metres. Zone A (5.60 to 5.46 m) contains high numbers (700 to 1600 per 5 ml) of primarily calcareous foraminifera (95–100%) and is similar to the upper contact of core BA-97-1. The zone is dominated by *A. mamilla* (≤64%), *A. beccarii* (≤26%) and *E. macellum* (≤13%), and also contains articulated gastropods and ostracods. Samples in zone B (5.46 to 5.00 m) contain small numbers of *Acari* spp., *Chironomidae* spp., and moss leaves but no foraminifera except one *A. beccarii*.

Figure 4.12. Foraminifera stratigraphy (%) of core BA-97-2: basal (lower contact) section.

Figure 4.13. Foraminifera stratigraphy (%) of core BA-97-2: in-core (upper contact) section.
The transfer function has calculated an indicative meaning below the range of modern samples (see section 3.5).

4.5 Chronology

Index number	1	2
Radiocarbon laboratory number	CAMS-75518	CAMS-75519
Core	BA-97-1	BA-97-2
Material	Bulk sediment (minerogenic)	Wood (twig)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	38.84 ± 0.18	41.24 ± 0.14
^{14}C age (years BP $\pm 1\sigma$)	7600 ± 40	7120 ± 30
Calibrated BP $\pm 1\sigma$ age ranges	(8409-8373)	(7962-7934) (7892-7874)
Calibrated BP $\pm 2\sigma$ age ranges	(8424-8349)	(8005-7993) (7974-7922) (7899-7865)
Median calibrated age (cal yrs BP)	8390	7885
Carbon content (% by weight)	1.1	60
δ^{13}C ± 0.1 (‰)	-25.6	-29.0
MTL sample height (m)	-12.05	-12.56
Miliammina fusca	8	0
Polysaccammina ipohalina	2	0
Trochammina inflata	90	96
Jadammina macrescens	174	216
Elphidium gerthi	0	136
Elphidium williamsoni	0	16
Haynesina germanica	0	512
Ammonia beccarii	10	1824
Indicative meaning (m MTL)	2.51 ± 0.29	0.41 ± 0.29
SLIP (m MTL)	-14.56 ± 0.29	-12.97 ± 0.29

Table 4.4. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 4.14 and 4.15. See Table 4.2 for MTL datum.

Bantham Sands chronology is determined from two age measurements by AMS ^{14}C dating the basal sediment of cores BA-97-1 and 2 (Table 4.4) and provide the long-term (10^3 yr) Holocene RSL history for the site (Figure 4.14). SLIPs 1 and 2 date the early Holocene marine transgression of Bantham Sands between *ca.* 8400 and 7850 cal years BP (Figures 4.10, 4.12 and 4.14). SLIP 1 shows that high marsh (around +2.5 m MTL) conditions existed at Bantham Sands *ca.* 8425 – 8350 cal years BP and SLIP 2 shows that low marsh or mudflat (around +0.4 m MTL) conditions developed *ca.* 8000 – 7850 cal years BP. The data show that mean sea level rose locally by ~1 – 2.2 m in approximately 560 to 350 years respectively during the early Holocene, a rate of ~2 – 6 m/ka (Figure 4.14). Lambeck's (pers. comm. 1997) geophysical model of early Holocene RSL rise in south Devon is the best fit of the Bantham Sands results (Figure 4.15).

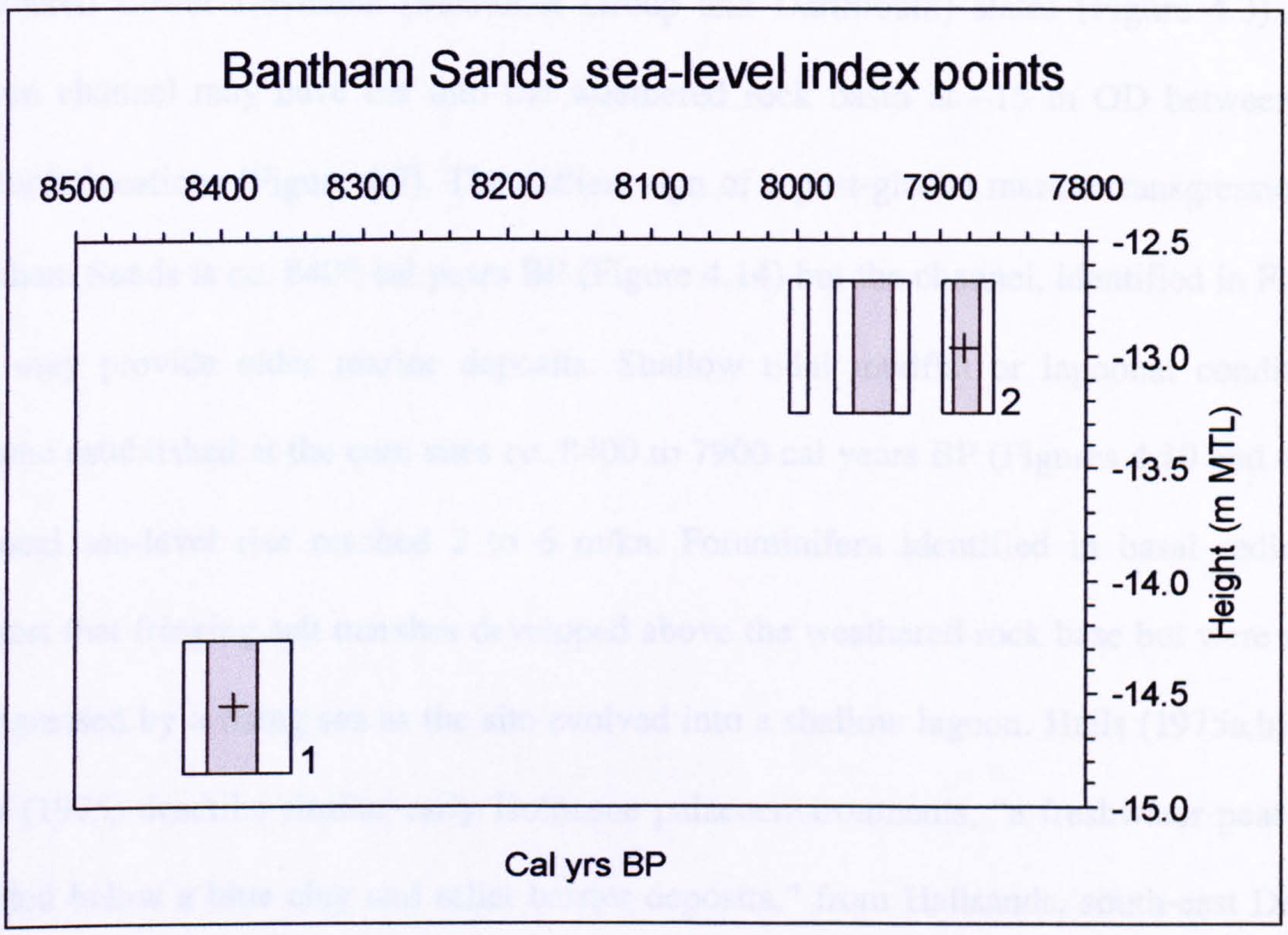


Figure 4.14. Bantham Sands SLIPs. Index numbers correspond with Table 4.4.

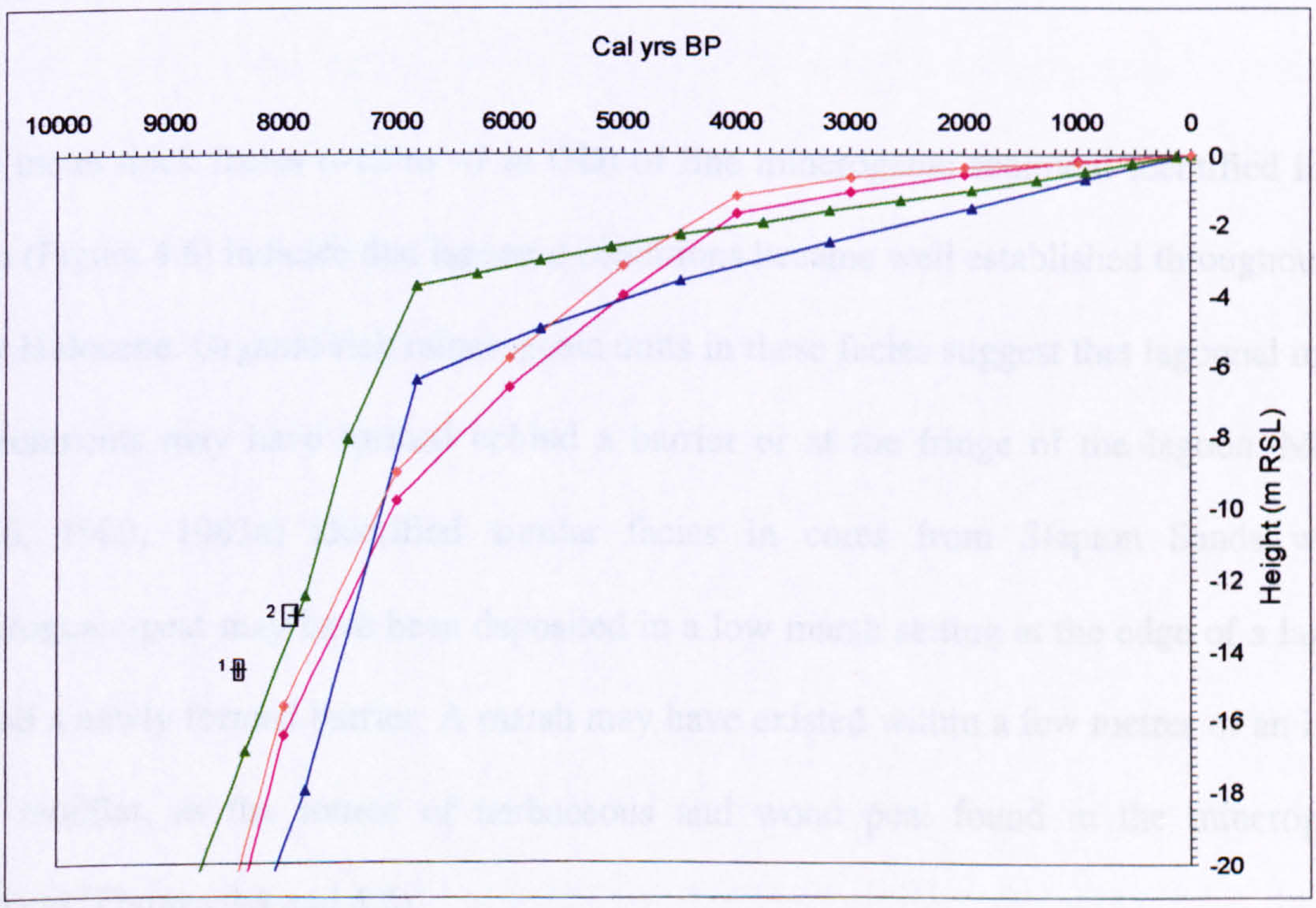


Figure 4.15. Bantham Sands SLIPs and geophysical modelling results. Index numbers correspond with Table 4.4.

4.6 Palaeoenvironmental interpretation

The early Holocene Bantham coastline probably resembled a wooded valley drained by a stream through marshland onto a rock foreshore and shifting sand flat. This developed on weathered Lower Devonian (Meadfoot Group and Dartmouth) slates (Figure 4.3). The stream channel may have cut into the weathered rock basin at –15 m OD between the borehole locations (Figure 4.7). The earliest sign of a post-glacial marine transgression of Bantham Sands is *ca.* 8400 cal years BP (Figure 4.14) but the channel, identified in Figure 4.7, may provide older marine deposits. Shallow tidal mudflat or lagoonal conditions became established at the core sites *ca.* 8400 to 7900 cal years BP (Figures 4.10 and 4.12) as local sea-level rise reached 2 to 6 m/ka. Foraminifera identified in basal sediment suggest that fringing salt marshes developed above the weathered rock base but were soon transgressed by a rising sea as the site evolved into a shallow lagoon. Hails (1975a,b) and Lees (1975) describe similar early Holocene palaeoenvironments, "a freshwater peat unit situated below a blue clay and relict barrier deposits," from Hallsands, south-east Devon where the onset of the Holocene marine transgression was dated 8108 ± 60 BP at –16.5 m RSL (±1 m).

Five metre thick facies (–12 to –7 m OD) of fine minerogenic sediment identified in the cores (Figure 4.6) indicate that lagoonal conditions became well established throughout the early Holocene. Organic-rich minerogenic units in these facies suggest that lagoonal marsh environments may have formed behind a barrier or at the fringe of the lagoon. Morey (1976, 1980, 1983a) identified similar facies in cores from Slapton Sands where minerogenic-peat may have been deposited in a low marsh setting at the edge of a lagoon behind a newly formed barrier. A marsh may have existed within a few metres of an inter-tidal mudflat, as the source of herbaceous and wood peat found in the minerogenic sediment (Figures 4.4 and 4.5).

A gravel barrier and inter-tidal sandbank system probably established itself close to the present-day Bantham coastline before the mid-Holocene. Silty-sand containing marine shells between –7 and –2 m OD (Figure 4.6) represents a marine setting in close proximity to such a system. Some of the earlier extensive sand flats would have provided the source material for the dunes of Bantham Ham. As these coastal systems extended from the north and south of the inlet the lagoon itself became choked with sand and gravel (–2 to 0 m OD, Figure 4.6) leaving only a narrow opening to Bigbury Bay.

Closure of the tidal inlet appears to have been very sudden, as there is no evidence for the development of a salt marsh. Gravel facies devoid of marine shells or foraminifera (Figures 4.11 and 4.13) are identified at –0.85 m OD (Figure 4.5) and +0.26 m OD (Figure 4.4). The coarse sandy-gravel and microscopic shell fragments is possibly the remnant of a washover event following barrier closure of the lagoon. Hails (1975a,b), Lees (1975) and (Morey (1980, 1983a,b) identified similar events at nearby Slapton Sands. Following barrier closure the site was then colonised by *Phragmites australis* and alder similar to the back-barrier environment identified at Slapton Sands by Morey (1976, 1980). Small numbers of mites and midges are present in the peat also inferring a moist vegetated terrestrial environment. The freshwater marsh system then established itself as high groundwater levels were maintained by its close proximity to the sea. Throughout the late Holocene the sand dunes gained height and extent along the coast preventing further breaching by the sea. This provided source material for the aeolian sand deposited in the back-barrier environment (+2.34 m OD, Figure 4.6).

4.7 Summary

Prior to this study no other SLIPs have been generated to date the early Holocene sea-level rise at Bantham Sands. The lithostratigraphy reveals that a barrier-lagoon coastline was initiated on top of the Lower Devonian slate shoreline during the early Holocene marine incursion. Foraminiferal and ^{14}C analyses reveal that basal salt-marsh peat existed *ca.* 8400 cal years BP but was subsequently eroded as ephemeral barriers retreated landward with the rising sea. The rate of this local sea-level rise, *ca.* 8000 BP, is around 2 – 6 m/ka. Geophysical models, particularly Lambeck's (pers. comm. 1997), agree closely with basal SLIPs. During the mid- to late Holocene the Bantham Sands basin gradually filled in with lagoonal sediments until barriers eventually closed the narrow tidal inlet initiating the start of the back-barrier fresh-water marsh system. The present-day back-barrier environment is protected from further sea-level rise and storm surges by the sand dunes on Bantham Ham.

Chapter 5

North Sands

5.1 Site description

North Sands is located at SX 731 381 and derives its name from being the northerly of two large sandy bays situated on the western edge of the entrance to Salcombe Harbour (Figure 5.1). The site lies in the lower reaches of the Kingsbridge estuary and the back-barrier wetland is used for pastoral farming. The site has witnessed Neolithic and Medieval occupation but research on settlement in the area is limited (Salcombe and Kingsbridge Estuary Association, 1985). Prior to this study West (1997) has been the only author to complete a Holocene palaeoenvironmental investigation of North Sands. His radiocarbon-dated sample (AA-14699, 6705 ± 165 ^{14}C years BP) is from coarse basal (–9.89 m OD) minerogenic gravel below organic silty-clay with marine shells. The sample suggests that the site (SX 730 382) may have been inter-tidal as early as *ca.* 7900 to 7300 cal years BP (West 1997).

5.1.1 Geomorphology

The North Sands back-barrier system is a freshwater marsh that is drained by Mill Leat into North Sands Bay (Figure 5.2). It extends roughly 0.8 km inland from MHWST at +2 to +10 m OD and covers about 5 ha. North Sands Bay is an inter-tidal zone covering an area approximately 3 ha to the north-north-west of a submerged longshore bar located at the entrance to Salcombe Harbour (Figure 5.1). The Bay is bordered by a rocky coastal slope rising 3 to 5 m above MHWST and lies in the lee of the prevailing south-westerly winds and waves, protected by the cliffs of Bolt Head (Salcombe and Kingsbridge Estuary Association, 1985).

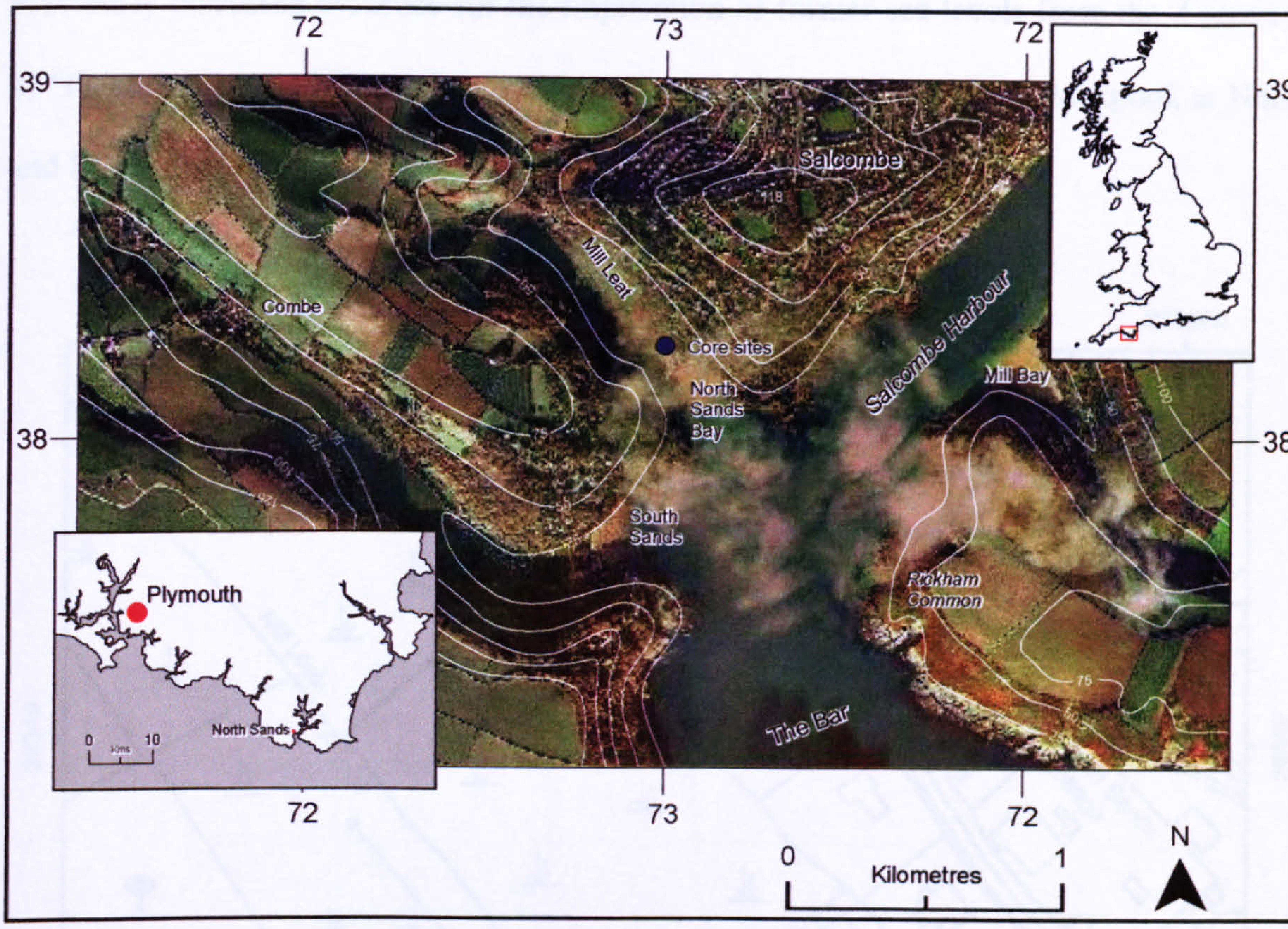


Figure 5.1. Aerial view of North Sands. Source: http://www.multimap.com. Extract from Great Britain map and aerial photograph of South Hams District at scale 1:25 000 (2002). Heights in metres above MSL.

The sedimentary environments along the neighbouring coastline to North Sands are similar to those found around Bigbury Bay (see section 4.1.1). The rocky coastline consists of many bays, coves and islets situated below cliff tops rising to 130 m within 200 to 800 m of MHWST either side of Salcombe Harbour. Other local examples of back-barrier systems exist, most notably at South Sands, approximately 0.5 km south-south-west of North Sands Bay and Mill Bay at SX 7410 3815 approximately 1 km east of North Sands Bay (Figure 5.1). A radiocarbon-dated (7425 ± 50 ^{14}C years BP) sample has been obtained from the basal (–13.28 m OD) sediment of a South Sands core dating the start of the Holocene sedimentary sequence at *ca.* 8400 to 8000 cal years BP (Jack, pers. comm. 2001). West (1997) cites evidence of a local submerged forest identified in a text entitled 'Kingsbridge and its Surroundings' (Anonymous, 1874: 169) describing a wood that is "believed to have been overwhelmed by the waves in times remote". Pengelly (1866b) had previously discussed evidence for the implication of former sea levels from the discovery of "these remnants of the ancient woods of South Devon in Torbay, at Blackpool, at North and South Sands in the Salcombe Estuary, and in Bigbury Bay".

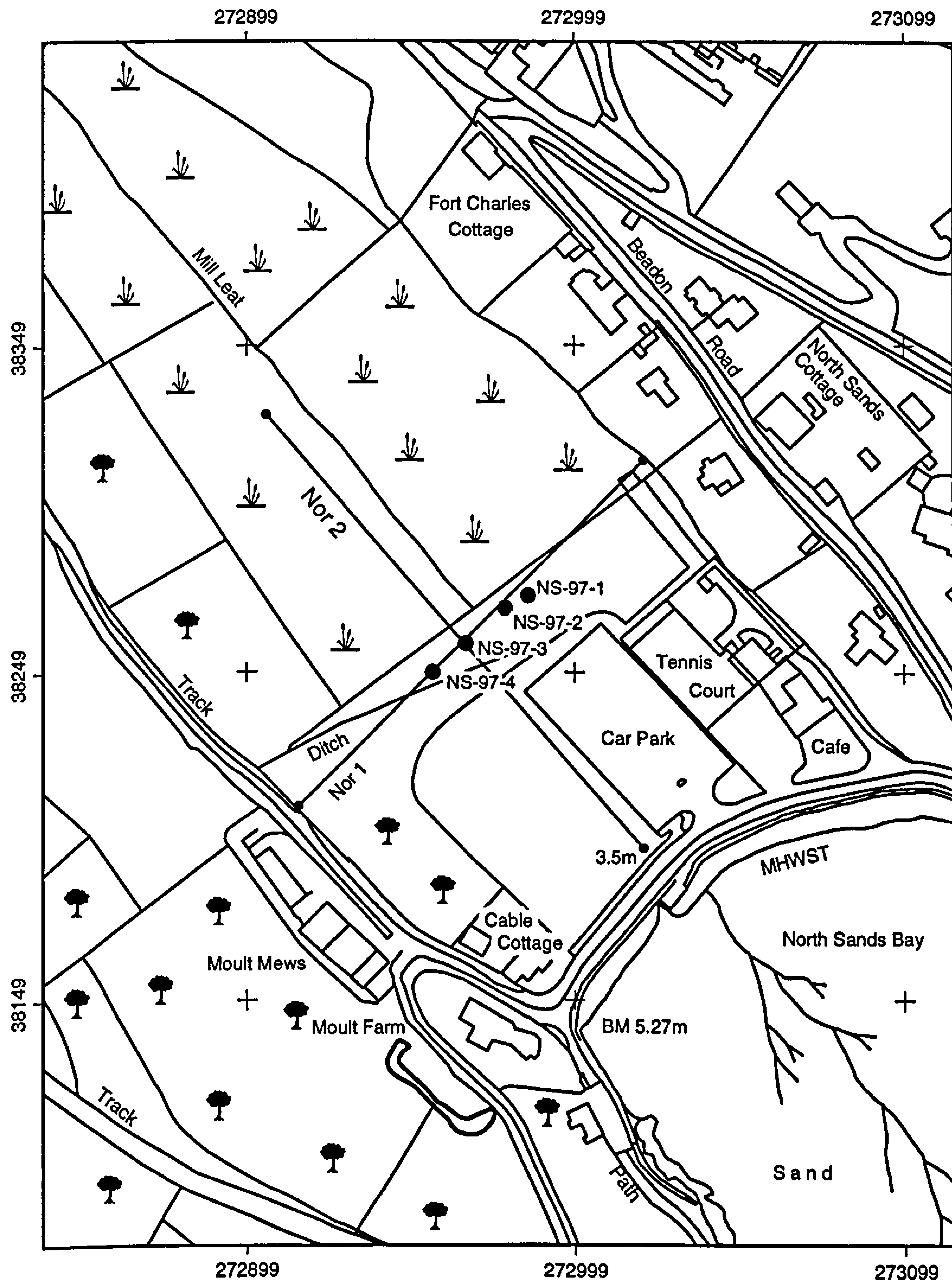


Figure 5.2. North Sands back-barrier system. Borehole locations: cores NS-97-1, 2, 3 and 4. Electrical resistivity survey lines: Nor 1 and Nor 2. Source: http://digimap.edina.ac.uk/service/advanced/carto_welcome.html. Map extract at scale 1:2500 (2002).

5.1.2 Geology

North Sands is part of the Start Point Complex comprising two groups of solid geology derived from altered Devonian rocks (Ussher, 1904; Figure 5.3). These are Green (hornblende and chlorite) Schists and Grey (pelitic) Schists. The Green Schists consist of altered basic lavas or sills (Durrance and Laming, 1982). The Grey Schists are derived from slate, siltstone and sandstone and consist of quartz and muscovite minerals (Born, 1986). The Grey Schists are more susceptible to weathering. The Quaternary deposits of the North Sands coastline include river gravel, local head deposits, and alluvial sediments. The North Sands study site lies within an area of alluvial sediments, constrained to the north and south by hornblendic and chloritic schists.

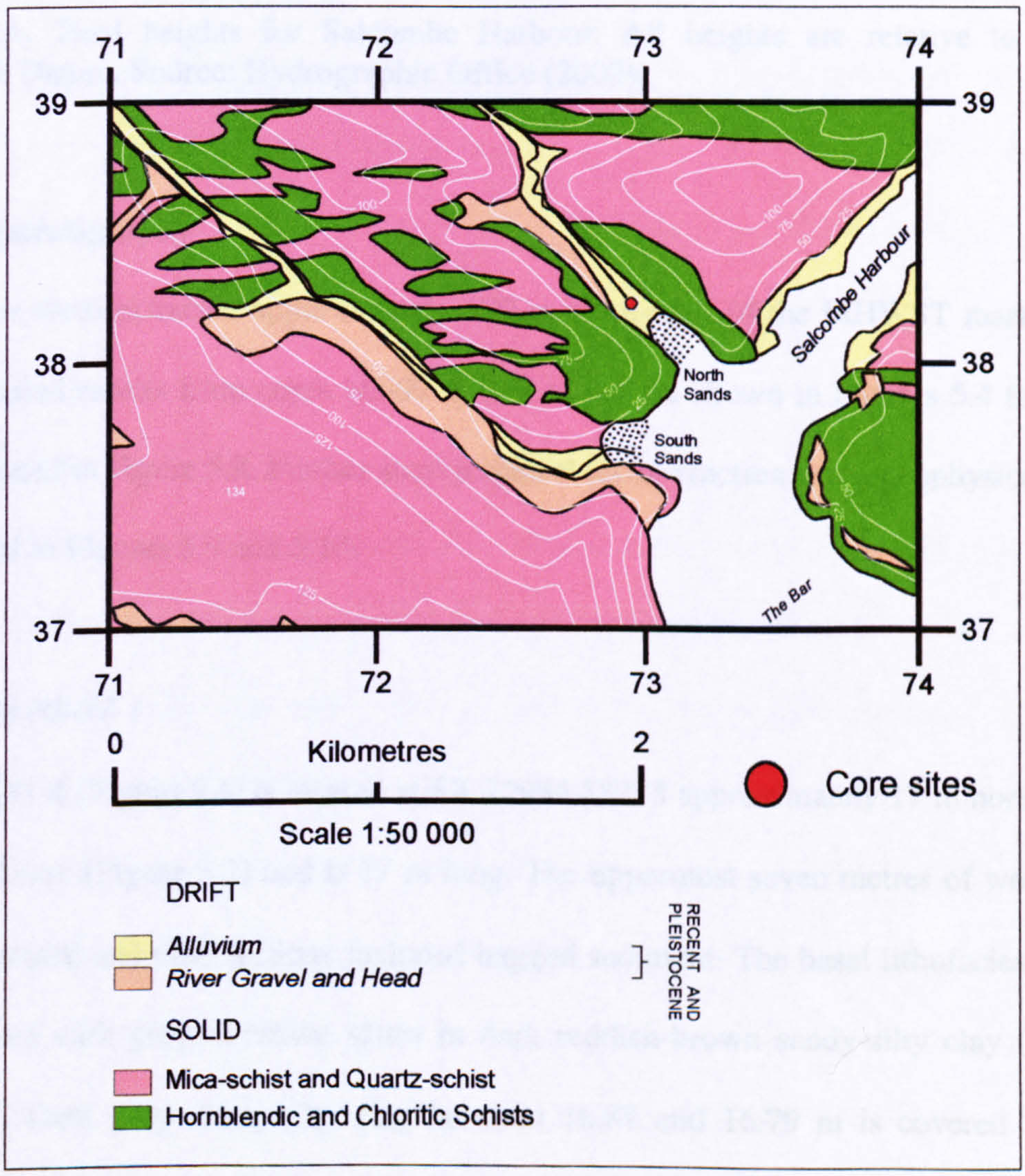


Figure 5.3. The solid geology and Quaternary deposits of North Sands and the surrounding country. Source: Geological Survey of Great Britain (England and Wales) Sheets 355 and 356, Kingsbridge and Start Point, 1:50 000 (1975). Heights in metres above MSL.

5.1.3 Tidal regime

Tidal heights for the entrance to Salcombe Harbour are shown in Table 5.1. North Sands Bay is within 1 km of the Salcombe Harbour tide gauge station and the heights are likely to be the same. The borehole sites are on a marsh +2.5 to +3 m MTL, and are exposed to southerly and south-easterly winds and waves, despite being 150 m from the MHWST mark. Any flooding of the North Sands back-barrier freshwater marsh system is normally caused by a rise in groundwater levels following heavy rainfall or by storms.

Tidal Station / Site	Latitude N	Longitude W	MHWST (m)	MLWST (m)	Tidal Range (m)	Chart Datum (m)	MSL (m)	MTL (m)
Salcombe Harbour	50° 13'	3° 47'	2.25	-2.35	4.6	-3.05	3.14	0.09

Table 5.1. Tidal heights for Salcombe Harbour. All heights are relative to the UK Ordnance Datum. Source: Hydrographic Office (2002).

5.2 Lithostratigraphy

Four cores were extracted approximately 100 m north-west of the MHWST mark (Figure 5.2). Detailed results from cores NS-97-1, 2, 3 and 4 are shown in Figures 5.4 to 5.7 and are correlated in Figure 5.8. Further stratigraphical reconstruction using geophysical survey is depicted in Figures 5.9 and 5.10.

5.2.1 Core NS-97-1

Core NS-97-1 (Figure 5.4) is located at SX 72985 38273 approximately 17 m north-east of Mill Leat sink (Figure 5.2) and is 17 m long. The uppermost seven metres of woody peat were discarded and nine sections included bagged sediment. The basal lithofacies consists of fractured dark greyish-brown slates in dark reddish-brown sandy-silty clay (17.00 to 16.87 m). Dark grey sandy-silty clay between 16.87 and 16.79 m is covered by black highly humified peat (16.79 to 16.66 m) containing a nut of *Corylus avellana* (hazel). Above the peat is a grey sand and gravel (16.66 to 16.59 m) and black wood in sandy-clayey silt (16.59 to 16.54 m). The overlying lithofacies consists of dark grey sandy-silty clay with shell and wood fragments (16.54 to 10.00 m). The uppermost lithofacies is predominantly black humified woody peat with *Phragmites australis* remains (10.00 to 7.00 m).

5.2.2 Core NS-97-2

Core NS-97-2 (Figure 5.5) is located at SX 72978 38269 approximately 9.4 m south-west of core NS-97-1 (Figure 5.2) and is 16 m long. The uppermost seven metres of peat were discarded, one section suffered core loss and eight sections included bagged sediment. The basal lithofacies consists of fractured dark grey slates in sandy-silty clay with shell and plant remains (16.00 to 15.73 m). This is overlain by black highly humified woody peat (15.73 to 15.47 m). Higher in the core the lithofacies is predominantly dark grey sandy-silty clay with large shells and wood fragments (15.47 to 10.10 m). The uppermost sequence is largely black highly humified woody peat (10.10 to 7.00 m) containing a nut of *Corylus avellana* (hazel) at 10.08 m and an alder branch at 7.45 m.

5.2.3 Core NS-97-3

Core NS-97-3 (Figure 5.6) is located at SX 72966 38258 approximately 16.8 m south-west of core NS-97-2 (Figure 5.2) and is 11.5 m long. The top 0.5 m of peat was discarded, four sections suffered core loss and seven sections included bagged sediment. The basal lithofacies consists of fractured dark grey slates in reddish-brown sandy-clayey silt (11.50 to 11.15 m). This is overlain by dark reddish-grey silty-sandy-gravelly clay (11.15 to 10.87 m), silty-clayey gravel (10.87 to 10.75 m) and dark grey clayey gravel with shell fragments (10.75 to 10.50 m). Above this, a unit of dark reddish-grey silty-sandy clayey-gravel (10.50 to 9.89 m) is covered by sandy-silty clay with shells (9.89 to 9.68 m) and light grey clayey peat (9.68 to 9.36 m). The uppermost lithofacies consists of a very dark grey silty-clayey peat (9.36 to 8.79 m) overlain by humified black woody peat (8.79 to 0.50 m).

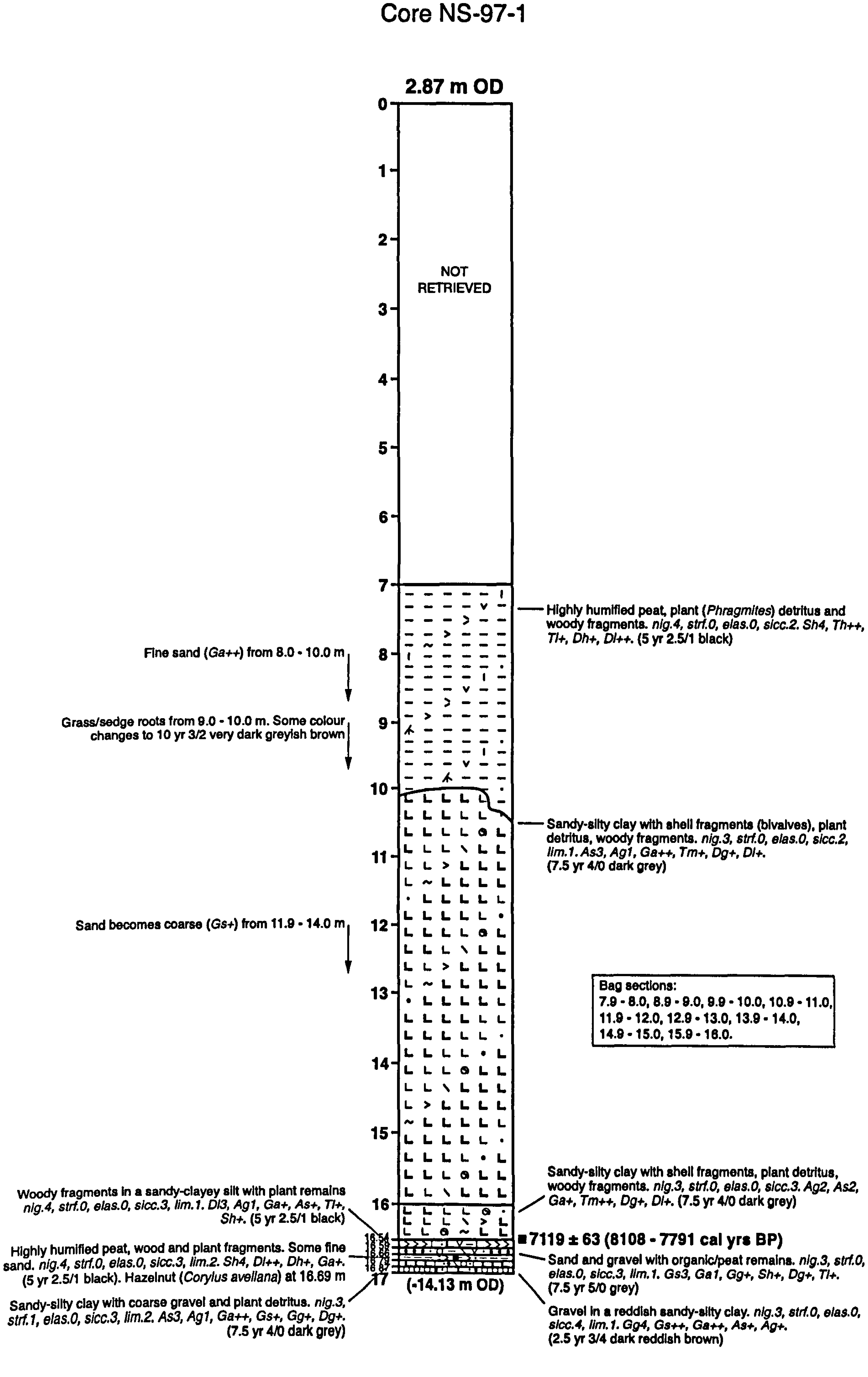


Figure 5.4. Lithostratigraphy of core NS-97-1.

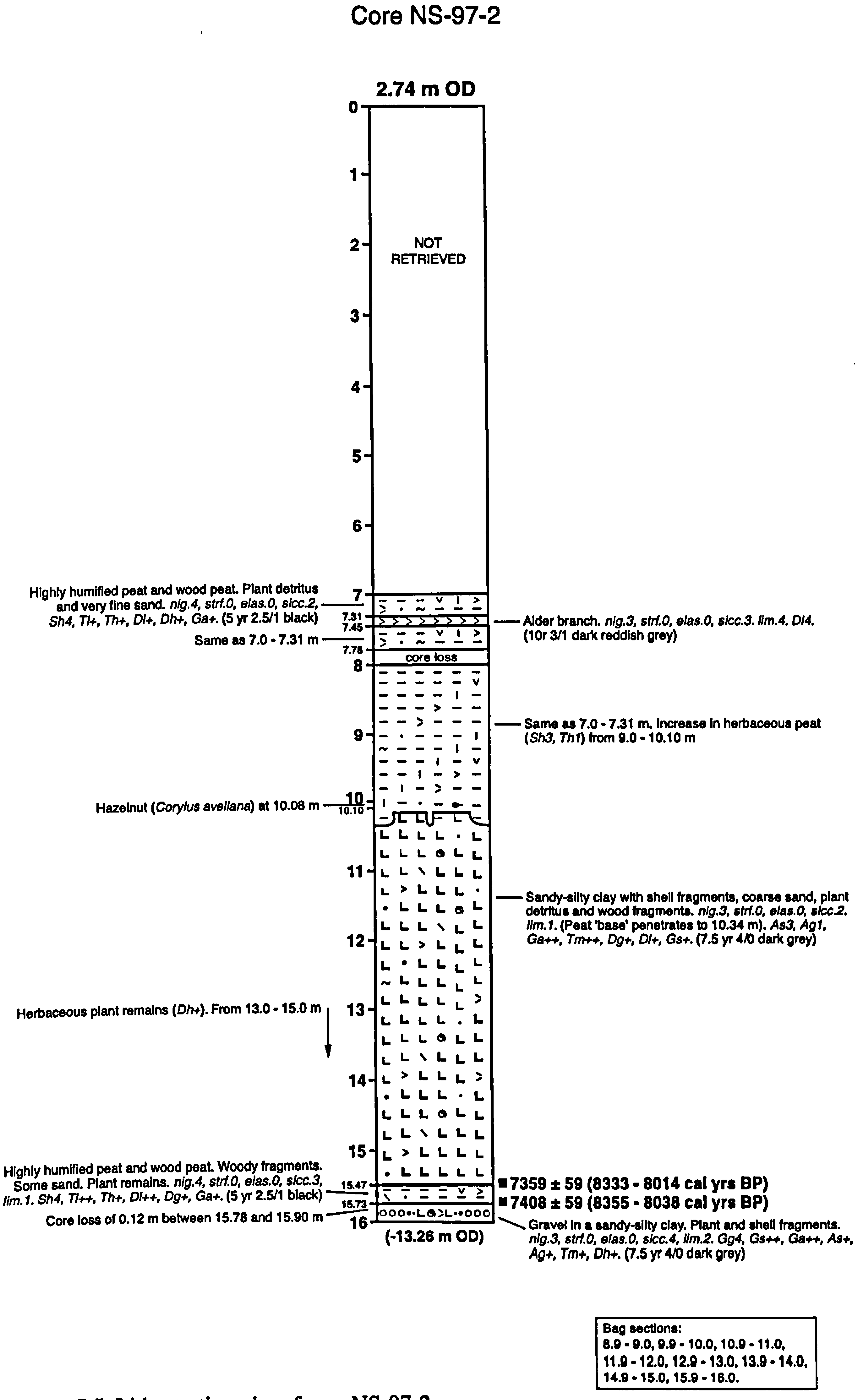


Figure 5.5. Lithostratigraphy of core NS-97-2.

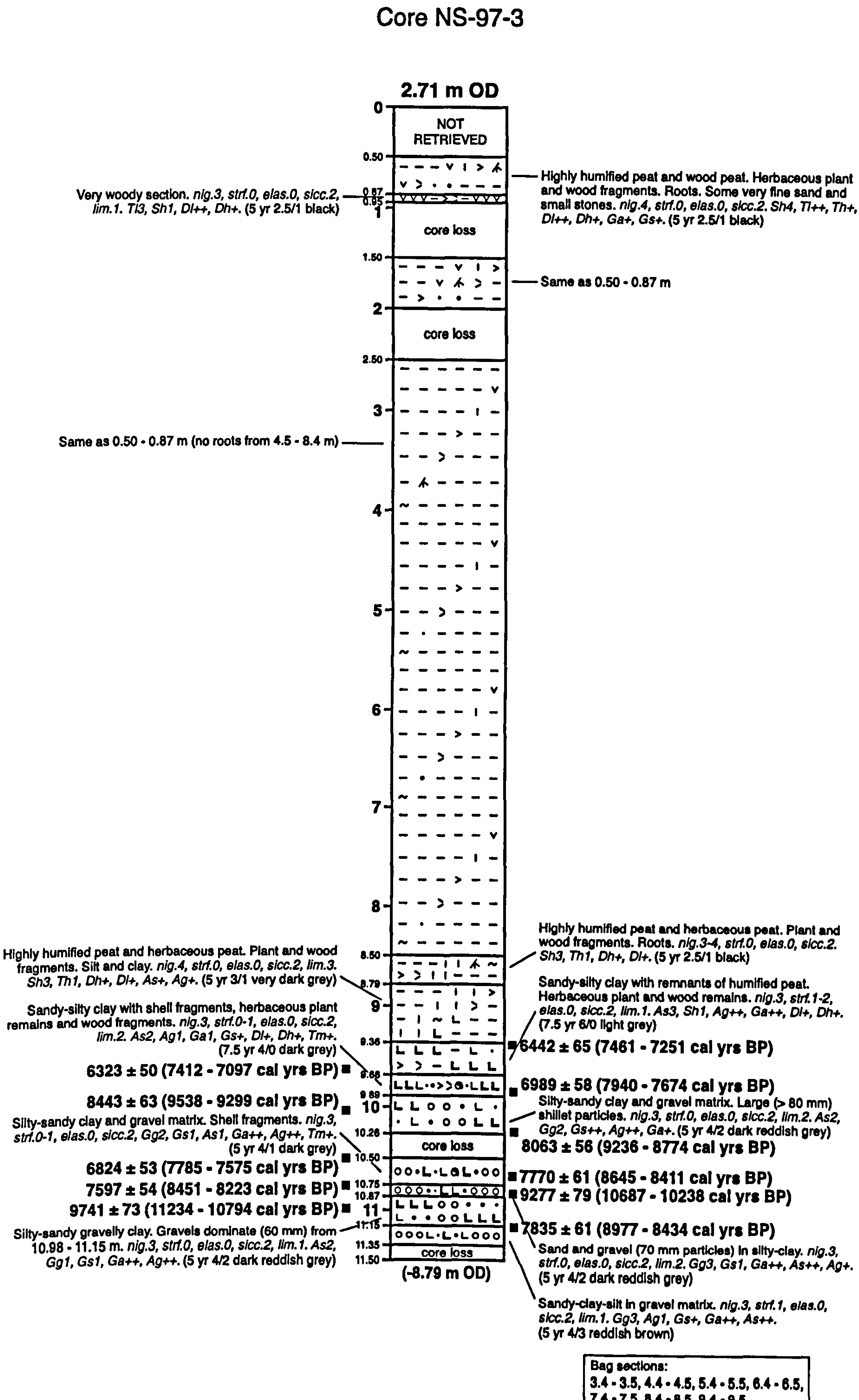


Figure 5.6. Lithostratigraphy of core NS-97-3.

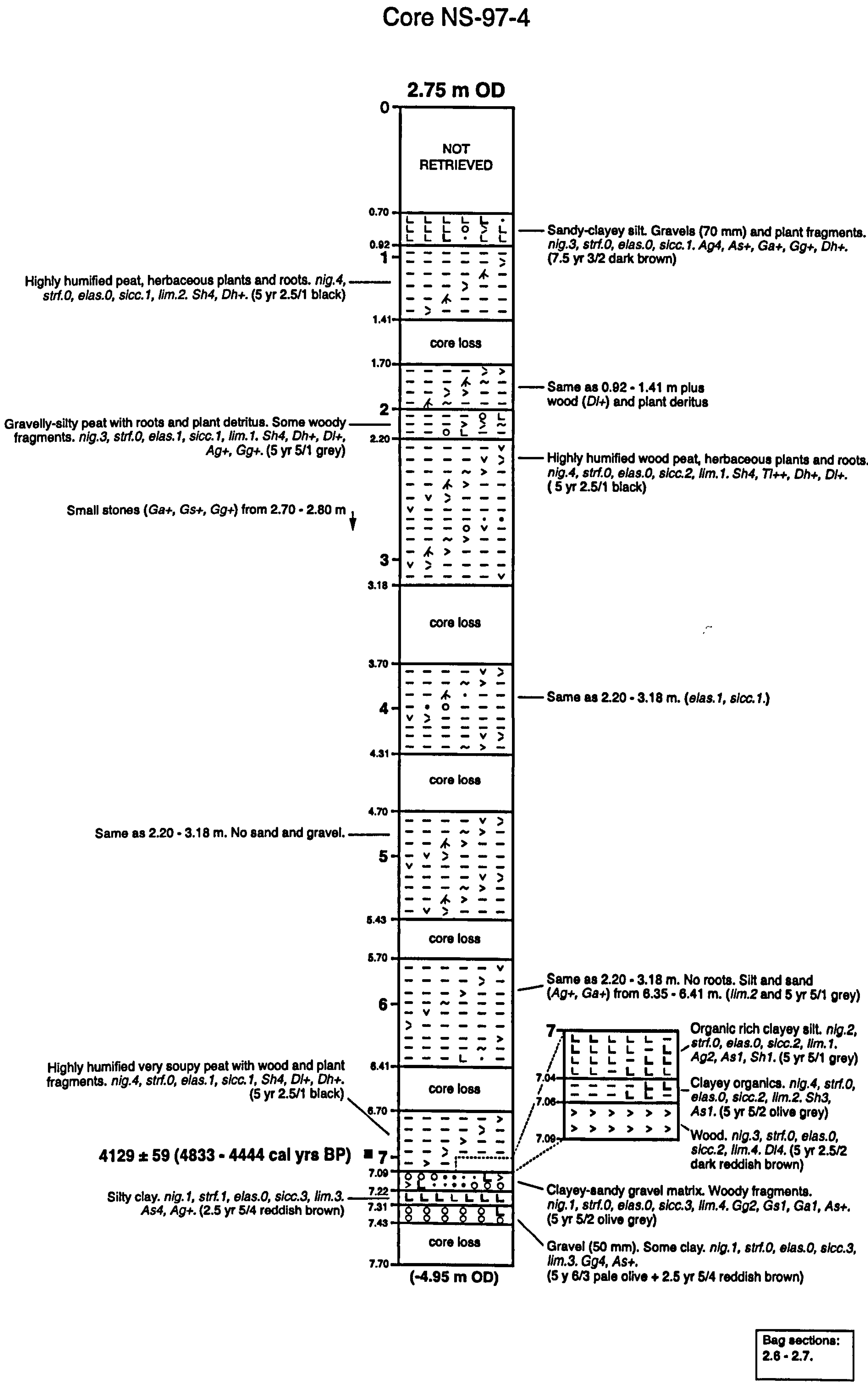


Figure 5.7. Lithostratigraphy of core NS-97-4.

5.2.4 Core NS-97-4

Core NS-97-4 (Figure 5.7) is located at SX 72956 38249 approximately 13.5 m south-west of core NS-97-3 (Figure 5.2) and is 7.7 m long. The uppermost 0.7 m of silt was discarded, six sections suffered core loss and one section included bagged sediment. The basal lithofacies consists of fractured pale olive and reddish-brown slates (7.70 to 7.31 m) overlain by reddish-brown silty clay (7.31 to 7.22 m) and olive grey clayey-sandy gravel (7.22 to 7.09 m). A very thin layer of wood, clayey peat and organic-rich clayey-silt lies between 7.09 and 7.00 m (see expanded view). The uppermost lithofacies is largely black humified wood peat with some sand and gravel (7.00 to 0.92 m). The top of the sequence (0.92 m to ground level) is dark brown sandy-clayey silt.

5.2.5 Core correlation

The lithostratigraphic facies described in sections 5.2.1 to 5.2.4 are correlated in Figure 5.8. The basal lithostratigraphy of North Sands consists of fractured slate, sand and gravel to a low of approximately –14 m OD. This is overlain in some cores by a thin basal peat (NS-97-1 and 2) or the remnants of a peat sequence (NS-97-3 and 4). The fine minerogenic phase is dominated by silty-clay to roughly –7 m OD. Very thin layers of minerogenic-peat overlie the fine minerogenic and large gravel sequences of cores NS-97-3 and 4. The organic phase at North Sands is predominantly highly humified wood peat except in core NS-97-4 where the topsoil consists of silt at +2 m OD.

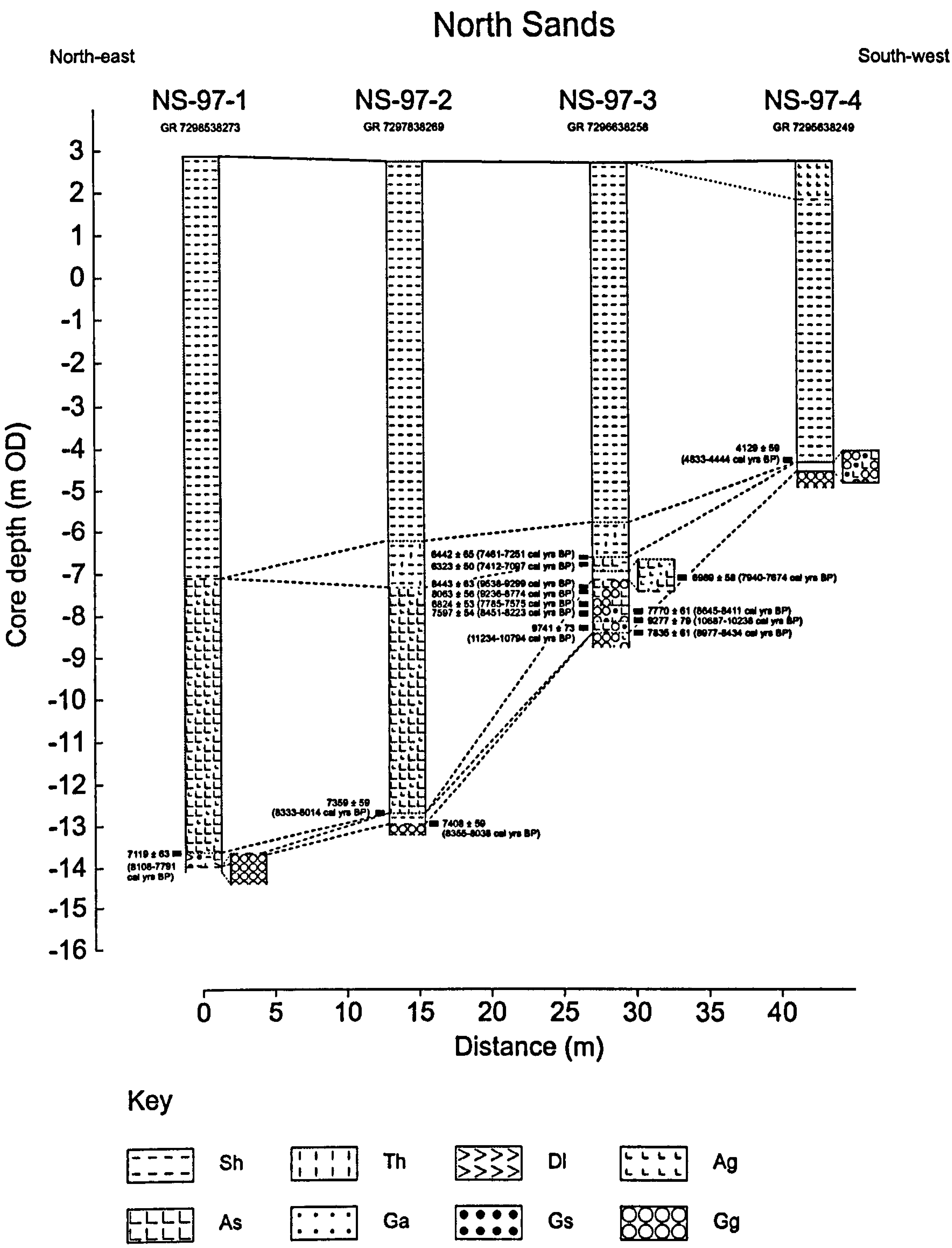


Figure 5.8. North Sands core correlation.

5.3 Electrical resistivity survey

A shore-parallel (Figure 5.9) and shore-normal (Figure 5.10) electrical resistivity survey provide additional subsurface mapping information and cores NS-97-1, 2, 3 and 4 are used as ground-truths. The geographical locations of the survey lines are shown in Figure 5.2. A summary of the electrical resistivity of sediments at North Sands is given in Table 5.2.

5.3.1 Shore-parallel resistivity survey (Nor 1)

Nor 1 extends 150 m from Moult Mews at SX 72915 38208 (+8.13 m OD) to Blue Waters Cottage at SX 73020 38314 (+4.32 m OD). Height recordings were taken from 29 electrodes at 5 m intervals except where the survey line crossed a ditch. A basin-shaped subsurface profile is delimited by the 68 Ωm isoline but basal gravel in the boreholes records resistivity of ≤83 Ωm (Figure 5.9). Core NS-97-1 is located to the south-west of the deepest (68 Ωm) point in the basin, although this is unclear from the results due to the short cable run. The deepest point at which 68 Ωm can be seen on Nor 1 is around –14 m OD at –15 m north-east of zero point. Peaks of high (68 Ωm) resistance occur around –4 m OD at –2 and –22 m north-east of zero point and coincide with the locations of Mill Leat and the ditch respectively (Figure 5.2).

The description of the electrical resistivity ranges of fine minerogenic sediment and peat from core NS-97-1 is complicated by interference from the ditch at –22 m distance. However, in cores NS-97-1 and 2 silty-clay has resistivity values between 56 and 83 Ωm, in core NS-97-3 it is 38–46 Ωm, and ≤32 Ωm in core NS-97-4 and the sandy-gravel unit in core NS-97-3 is 38–56 Ωm (Figure 5.9). Peat in cores NS-97-1 and 2 has values of ≤68 Ωm, in core NS-97-3 it is ≤38 Ωm and ≤32 Ωm in core NS-97-4. Basal peat facies are too thin to register changes in electrical resistance. The silt at the top of core NS-97-4 has values ≤32 Ωm. On the whole, results for Nor 1 suggest that no correlation exists between resistivity and lithostratigraphy across the basin.

5.3.2 Shore-normal resistivity survey (Nor 2)

Nor 2 extends from the road bridge at SX 73020 38195 (+3.74 m OD) to the back-barrier marsh at SX 72905 38328 (+2.47 m OD), approximately 6 to 10 m from Mill Leat reducing interference from the watercourse. Heights were recorded from 35 electrodes over 180 m at regular 5 m intervals. An undulating subsurface profile is delimited by 63 Ωm rising south-east to north-west (Figure 5.10). Basal gravel in core NS-97-3 records resistivity values of 50–63 Ωm at zero point. The core is located 5 m south-east of the deepest part (–15 m OD) of the shore-normal profile (79 Ωm). This value is also recorded at –22 m south-east of the borehole rising to within 10 m of ground level at –45 m from zero point. The base of core NS-97-3 is composed of significant amounts of silt, clay, sand and gravel covering the fractured slates (Figure 5.6). This unit has a resistivity of 50–63 Ωm, the overlying silty-clay facies is 40–63 Ωm, and peat is ≤63 Ωm. In general, results for Nor 2 show that it is difficult to reliably distinguish between the resistivity of peat, minerogenic and gravel facies (Table 5.3).

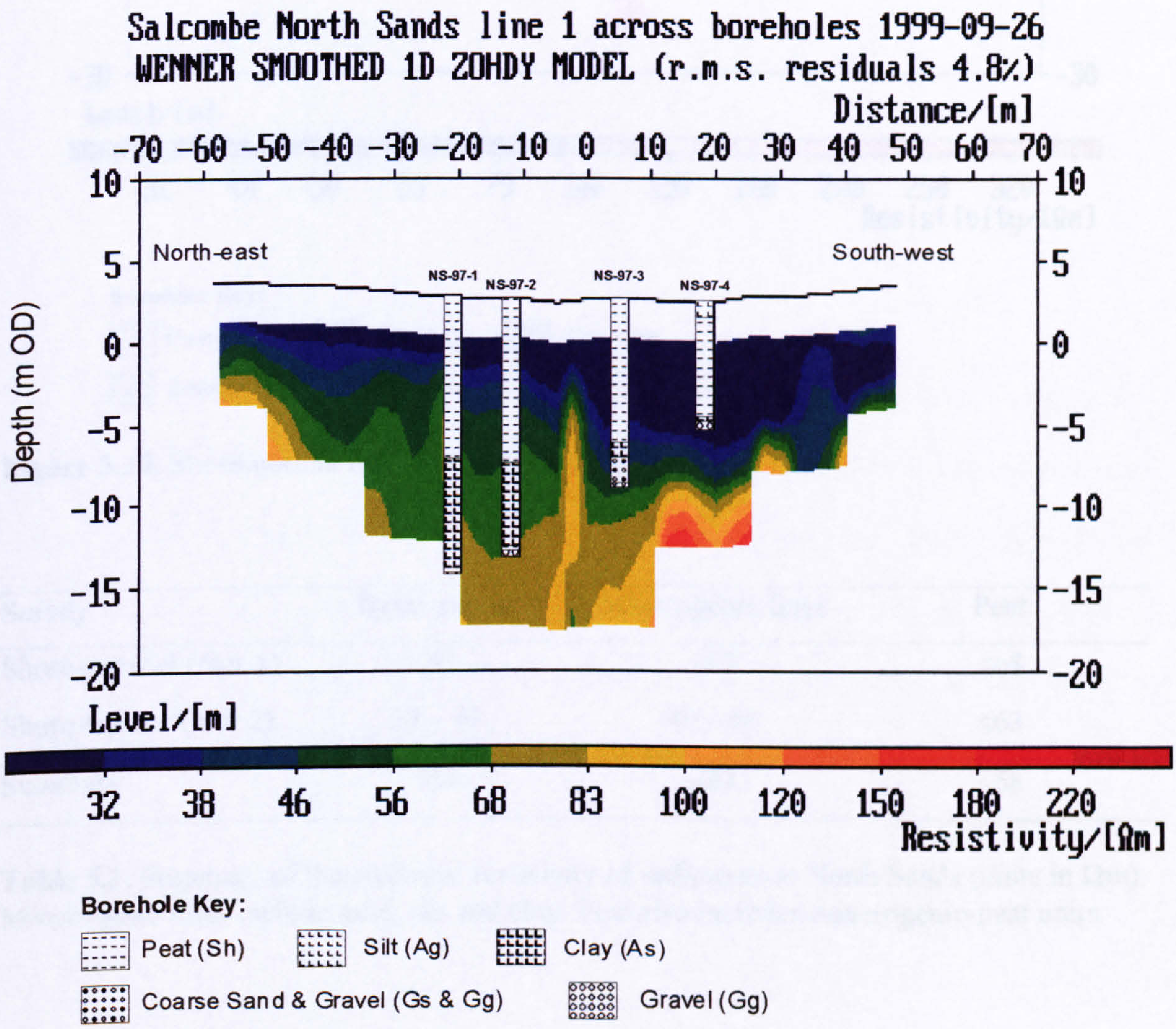


Figure 5.9. Shore-parallel resistivity profile (Nor 1).

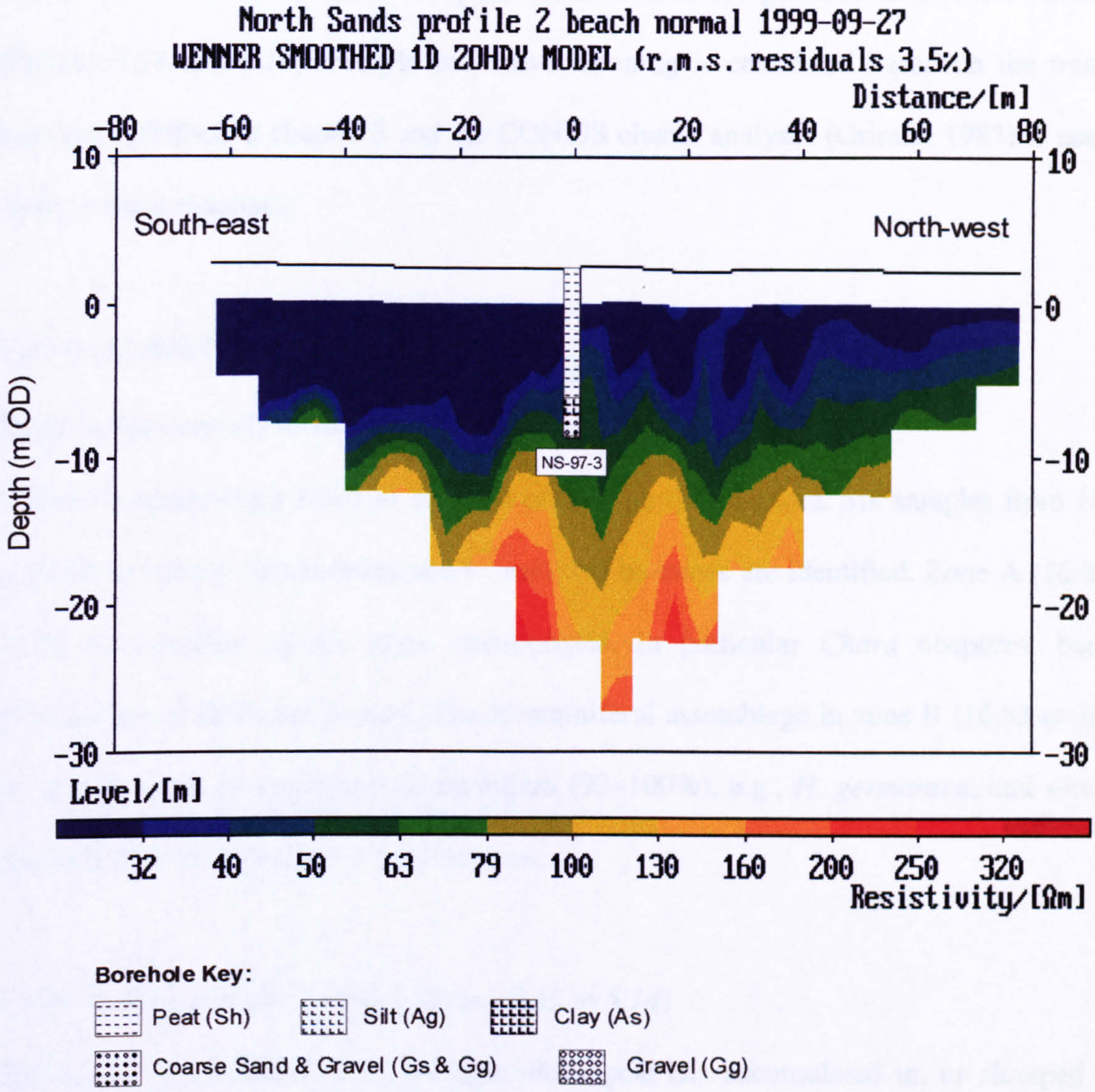


Figure 5.10. Shore-normal resistivity profile (Nor 2).

Survey	Basal gravel	Minerogenic fines	Peat
Shore-parallel (Nor 1)	≤83	≤83	≤68
Shore-normal (Nor 2)	50 – 63	40 – 63	≤63
Summary	≤83	≤83	≤68

Table 5.2. Summary of the electrical resistivity of sediments at North Sands (units in Ωm). Minerogenic fines include sand, silt and clay. Peat also includes minerogenic-peat units.

5.4 Biostratigraphy

The biostratigraphy of North Sands is partially reconstructed using foraminifera identified from the basal (lower contact) and in-core (upper contact) lithofacies of cores NS-97-1 and

2 (Figures 5.11 to 5.17). More complete fossil records are given in cores NS-97-3 and 4 (Figures 5.18 and 5.19). Sample indicative meaning is calculated based on the transfer function described in chapter 3 and the CONISS cluster analyses (Grimm, 1987) is used to identify fossil biozones.

5.4.1 Core NS-97-1

Basal (lower contact) section (Figure 5.11)

Thirteen samples from 16.86 to 16.54 m contain no foraminifera. Six samples from 16.53 to 16.48 m contain foraminifera and at least two biozones are identified. Zone A (16.86 to 16.54 m) contains aquatic plant macrophytes, in particular *Chara* oospores, but no foraminifera or shells are present. The foraminiferal assemblage in zone B (16.53 to 16.48 m) is dominated by calcareous foraminifera (92–100%), e.g., *H. germanica*, and samples contain calcareous gastropods and bivalves.

In-core (upper contact) section (Figures 5.12 to 5.14)

The contact is an uneven one, perhaps where peat has accumulated in, or slumped into hollows or creeks on a former mudflat surface. Thus peat and minerogenic sediment samples were extracted from similar horizons for biostratigraphical comparison (Figure 5.12). Four samples from 10.05 to 10.01 m (minerogenic sediment, Figure 5.13) and four samples from 10.04 to 10.00 m (peat, Figure 5.14) contain foraminifera. Four samples from 9.90 to 9.84 m contain no foraminifera. Bagged sediment between 9.90 and 10.00 m was not analysed.

Figure 5.11. Foraminifera stratigraphy (%) of core NS-97-1: basal (lower contact) section.

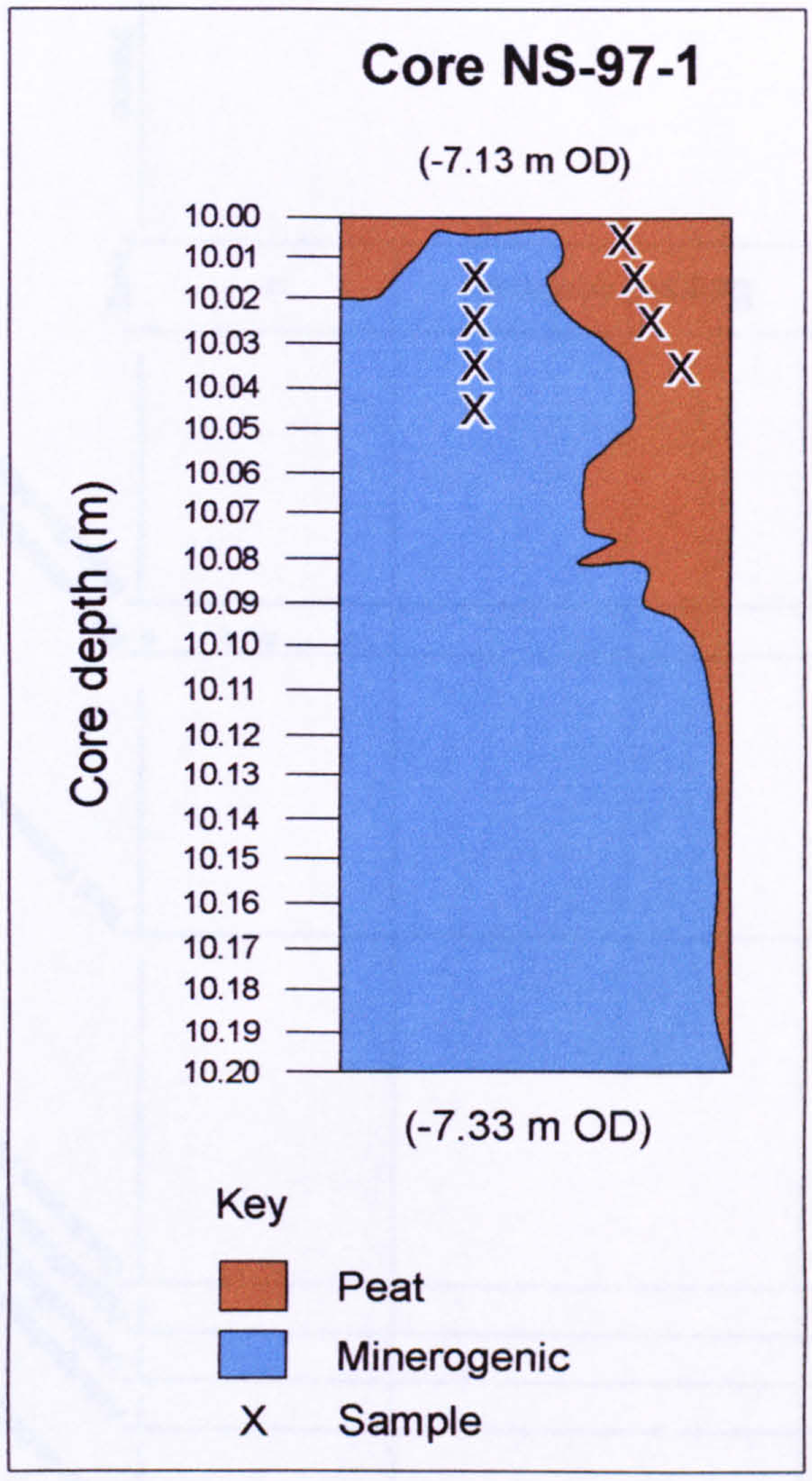


Figure 5.12. Profile of the upper contact in core NS-97-1. Crosses indicate sample locations.

The minerogenic samples in zone A (10.05 to 10.01 m, Figure 5.13) contain significant numbers of predominantly calcareous foraminifera (>7500 tests per 5 ml), shells, ostracods and diatoms. The assemblage is very diverse and *H. germanica* (~70%) and *A. beccarii* (~20%) dominate. The peat samples in zone A (10.04 to 10.00 m, Figure 5.14) contain significantly lower numbers (<600 tests per 5 ml) of calcareous foraminifera (78–98%) than the minerogenic sediment. The assemblage is less diverse, dominated by *H. germanica* and contains greater numbers of *J. macrescens*. Diatoms, *Chara* oospores, *Chironomidae* and beetle (*Coleopteran)* remains are also present. Zone B (9.90 to 9.84 m, Figures 5.13 and 5.14) contains diatoms, *Chironomidae* (mandibles) and aquatic plant macrophytes such as *Chara* oospores but no foraminifera.

Figure 5.13. Foraminifera stratigraphy (%) of core NS-97-1: in-core (upper contact) minerogenic section.

Figure 5.14. Foraminifera stratigraphy (%) of core NS-97-1: in-core (upper contact) peat section.

5.4.2 Core NS-97-2

Basal (lower contact) section (Figure 5.15)

Nineteen samples contain foraminifera, four samples between 15.69 and 15.53 m contain no foraminifera and six distinct biozones are identified. Zone A (15.74 to 15.69 m) is dominated by agglutinated foraminifera (67–100%) in particular *J. macrescens*, but total numbers are very low (3 to 40 specimens per 5 ml). Zone B (15.69 to 15.51 m) contains abundant plant macrophytes and *Chironomidae* remains but no foraminifera. There is a maximum of 8 *J. macrescens* per 5 ml sample in zone C (15.51 to 15.47 m). Species diversity (maximum 4) and numbers (maximum 88 per 5 ml) increase in zone D (15.47 to 15.43 m) and calcareous foraminifera dominate the assemblage with rising numbers of *H. germanica.* In zone E (15.43 to 15.39 m) agglutinated species comprise 60–90% of the total assemblage. In zone F (15.39 to 15.18 m) calcareous foraminifera dominate (85–95%) and species diversity (11) and numbers (1600 per 5 ml) increase in the uppermost sample. Zone F also contains ostracods, diatoms and shells.

In-core (upper contact) section (Figures 5.16 and 5.17)

The contact is an uneven one and peat 'wedges' at the sides and centre of the core sleeve did not contain enough material for a 5 ml sample. Sediment samples were therefore extracted from different horizons in the peat and minerogenic facies (Figure 5.16). Five samples from 10.25 to 10.00 m contain foraminifera, six samples from 10.05 to 9.83 m contain no foraminifera (Figure 5.17) and two biozones are identified. Zone A (10.25 to 9.99 m) contains significant numbers (≤2360 per 5 ml) of predominantly calcareous species (≤100%) similar to core NS-97-1 (Figure 5.13). Zone B (9.99 to 9.83 m) contains no foraminifera but small numbers of *Acari* spp., *Chironomidae* spp., *Coleoptera* and moss leaves are present in the peat.

Figure 5.15. Foraminifera stratigraphy (%) of core NS-97-2: basal (lower contact) section.

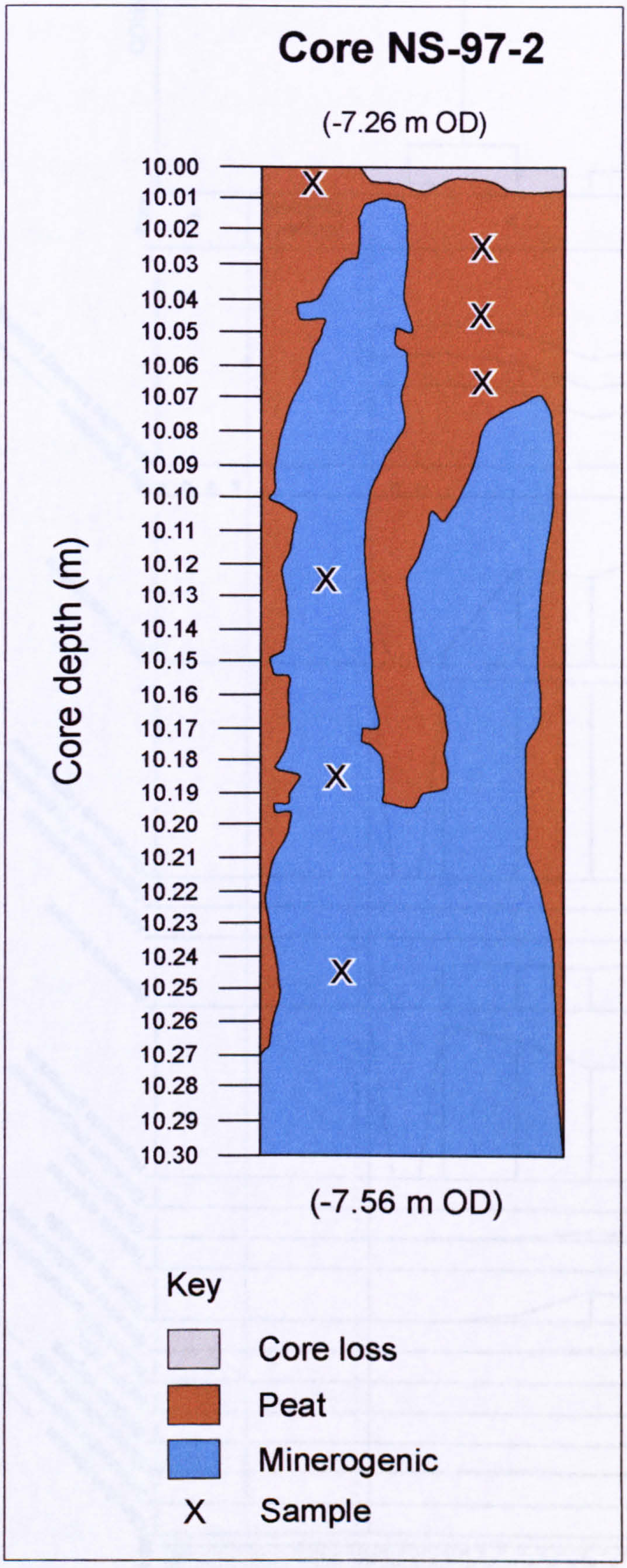


Figure 5.16. Profile of the upper contact in core NS-97-2. Crosses indicate sampling strategy.

Figure 5.17. Foraminifera stratigraphy (%) of core NS-97-2: in-core (upper contact) section.

5.4.3 Core NS-97-3

Basal (lower contact) to in-core (upper contact) section (Figure 5.18)

Sixty-three samples between 11.17 and 9.40 m contain foraminifera and nine samples contain no foraminifera. Bagged sediment from 9.50 to 9.40 m is not analysed, six samples contain no foraminifera between 9.39 and 9.25 m and eight distinct biozones are identified. The base of the core contains a few plant macrophytes but no foraminifera. Zone A (11.16 to 11.14 m) comprises very low numbers (≤9 per 5 ml) of predominantly agglutinated foraminifera (60–80%). Zone B (11.14 to 11.11 m) contains shell fragments (bivalves) and low numbers of foraminifera but calcareous species are dominant, e.g., *H. germanica* reaches 70%. Zone C (11.11 to 10.66 m) is dominated by *J. macrescens* and only a single calcareous species, *A. beccarii*, is found. This zone is devoid of shells but contains many plant remains, *Chironomidae* and *Acari*. A marked change in the faunal assemblage occurs in zone D (10.66 to 10.50 m) where calcareous foraminifera reach 85–100% dominance and shells, ostracods and diatoms are present. Above this zone 0.24 m core loss is recorded.

The uppermost part of the core section reveals a striking similarity to the lowermost. Zone E (10.26 to 10.07 m) consists of 100% agglutinated foraminifera, predominantly *J. macrescens*, except in very gravelly samples. Microscopic plant macrophytes, *Chironomidae* and *Acari* are found but no shells. In zone F (10.07 to 9.50 m), similar to zone D, calcareous foraminifera comprise 80–100% of the total and shells, ostracods, diatoms, plant macrophytes, *Chironomidae* and *Acari* are present. Counts reach 3700 specimens per 5 ml in this zone. The bagged sediment above zone F probably contains foraminifera because a sample at 9.39 m (zone G) contains *J. macrescens* and *H. germanica*. The uppermost six samples (zone H, 9.39 to 9.25 m) contain diatoms, plant macrophytes and species of *Acari* but no foraminifera.

Figure 5.18. Foraminifera stratigraphy (%) of core NS-97-3: basal (lower contact) to in-core (upper contact) section.

Figure 5.19. Foraminifera stratigraphy (%) of core NS-97-4: basal (lower contact) to in-core (upper contact) section.

5.4.4 Core NS-97-4

Basal (lower contact) to in-core (upper contact) section (Figure 5.19)

Five samples between 7.31 and 6.75 m contain foraminifera and twenty-five samples contain no foraminifera. Three biozones, based on the presence or absence of foraminifera, are identified. Zone A (7.31 to 7.12 m) contains very few plant macrophytes and no foraminifera. Two samples in zone B (7.12 to 6.95 m) contain only one animal (*A. beccarii* at 7.11 m and *H. germanica* at 7.02 m) and three samples contain these species plus *B. variabilis* and *E. williamsoni*, and numbers increase to 73 animals per 5 ml (6.98 and 6.95 m). Plant macrophytes, including *Phragmites* remains, and high numbers of *Acari* are also found in zones B and C. Zone C (6.94 to 6.75 m) contains no foraminifera.

5.5 Carbon analyses

Core NS-97-3 provides litho- (Figure 5.6) and biostratigraphical (Figure 5.18) results that contrast markedly. The basal and in-core sequence from 11.35 to 9.40 m reveals no discernible organic facies, unlike cores NS-97-1, 2 and 4 (Figures 5.4, 5.5 and 5.7 respectively). However, foraminiferal results suggest that high marsh conditions were well established at the site on at least two separate occasions (zones C and E, Figure 5.18). Total and organic carbon results are shown alongside chronological, litho- and biostratigraphical results (Figure 5.20) in an attempt to provide an alternative method to the foraminiferal sampling strategy for identifying potential salt-marsh facies (see section 2.3.2). The hypothesis proposed here is that salt-marsh facies will contain high carbon percentages. Carbon analysis also provides justification for the AMS ^{14}C dating of core NS-97-3.

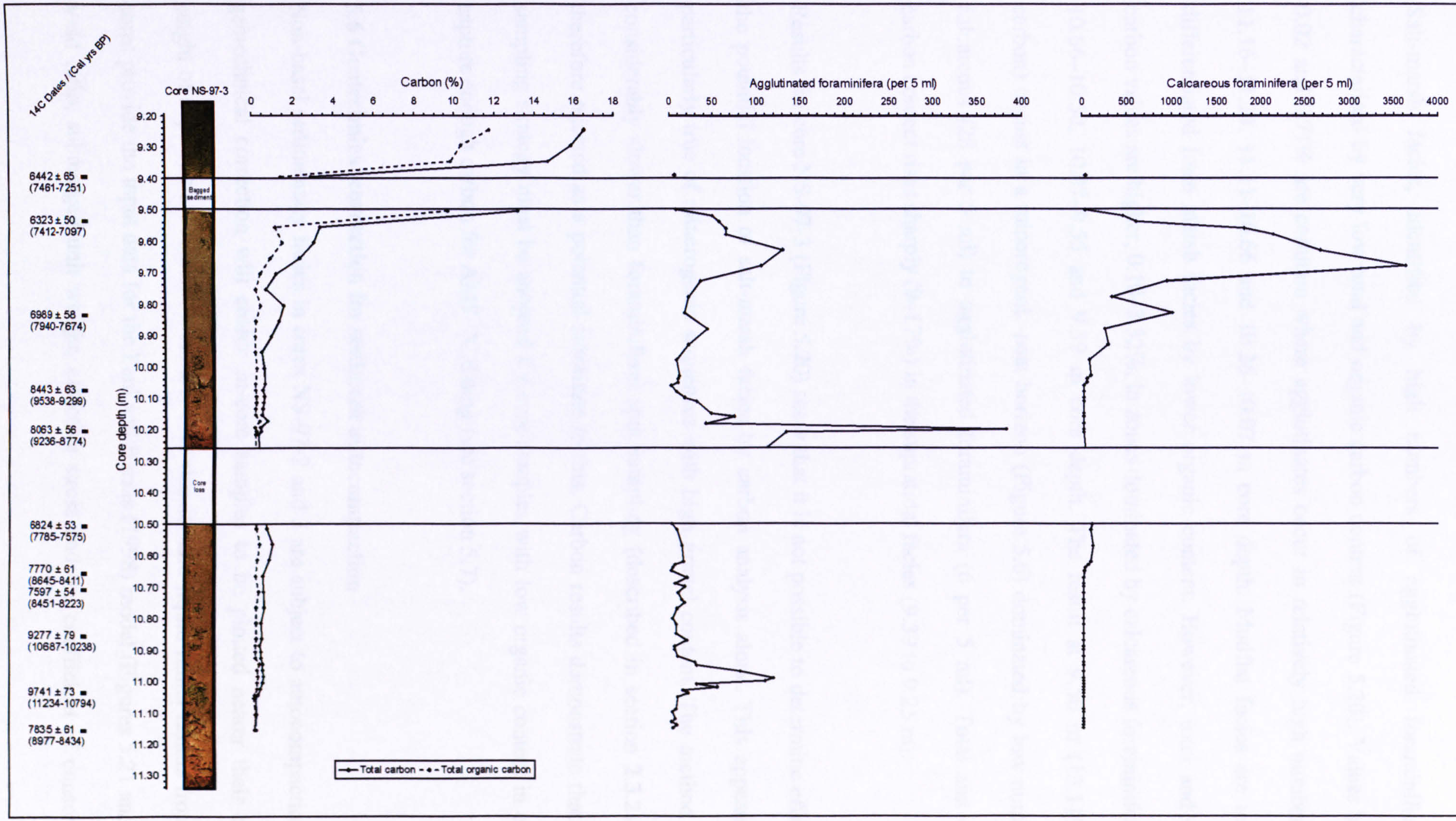


Figure 5.20. Total and organic carbon content (%) of core NS-97-3. Foraminiferal results are given in raw counts per 5 ml of sediment.

5.5.1 Total and organic carbon content of core NS-97-3

Salt-marsh facies, identified by high numbers of agglutinated foraminifera, are characterised by very low total and organic carbon content (Figure 5.20). Values between 0.02 and 0.71% are common where agglutinates occur in relatively high numbers, e.g., 11.16–11.14, 11.11–10.66 and 10.26–10.07 m core depth. Mudflat facies are routinely differentiated from marsh facies by lower organic content. However, total and organic carbon values are higher, 0.10–3.32%, in zones dominated by calcareous foraminifera, e.g., 10.66–10.50, 10.07–9.55 and 9.39 m core depth. The result at 9.50 m (13.11% total carbon) occurs in a minerogenic-peat horizon (Figure 5.6) dominated by low numbers of calcareous (28 per 5 ml) to agglutinated foraminifera (6 per 5 ml). Total and organic carbon content rises sharply (9–17%) in the supratidal facies (9.39 to 9.25 m).

Results for core NS-97-3 (Figure 5.20) show that it is not possible to determine effectively the potential location of salt-marsh facies by carbon analysis alone. This appears to be particularly true of minerogenic sequences with high gravel content. The method is also considerably slower than foraminiferal spot sampling (described in section 2.3.2) and is therefore rejected as a potential substitute to this. Carbon results demonstrate that a bulk sampling strategy must be adopted for core samples with low organic content in order to capture enough carbon for AMS ^{14}C dating (see section 5.7).

5.6 Geotechnical correction for sediment autocompaction

Non-basal sedimentary facies in cores NS-97-2 and 3 are subject to autocompaction but a geotechnical correction will enable in-core samples to be plotted nearer their original height of deposition. Sediment moisture, bulk density and liquid limits results from these cores provide the input data for the Paul and Barras (1998) model (Figures 5.21 and 5.24). Void ratio, submerged unit weight, effective stress and the coefficient of consolidation

(Figures 5.22 and 5.25) are used to determine the total correction for each core (Figures 5.23 and 5.26).

5.6.1 Core NS-97-2

Input data (Figure 5.21)

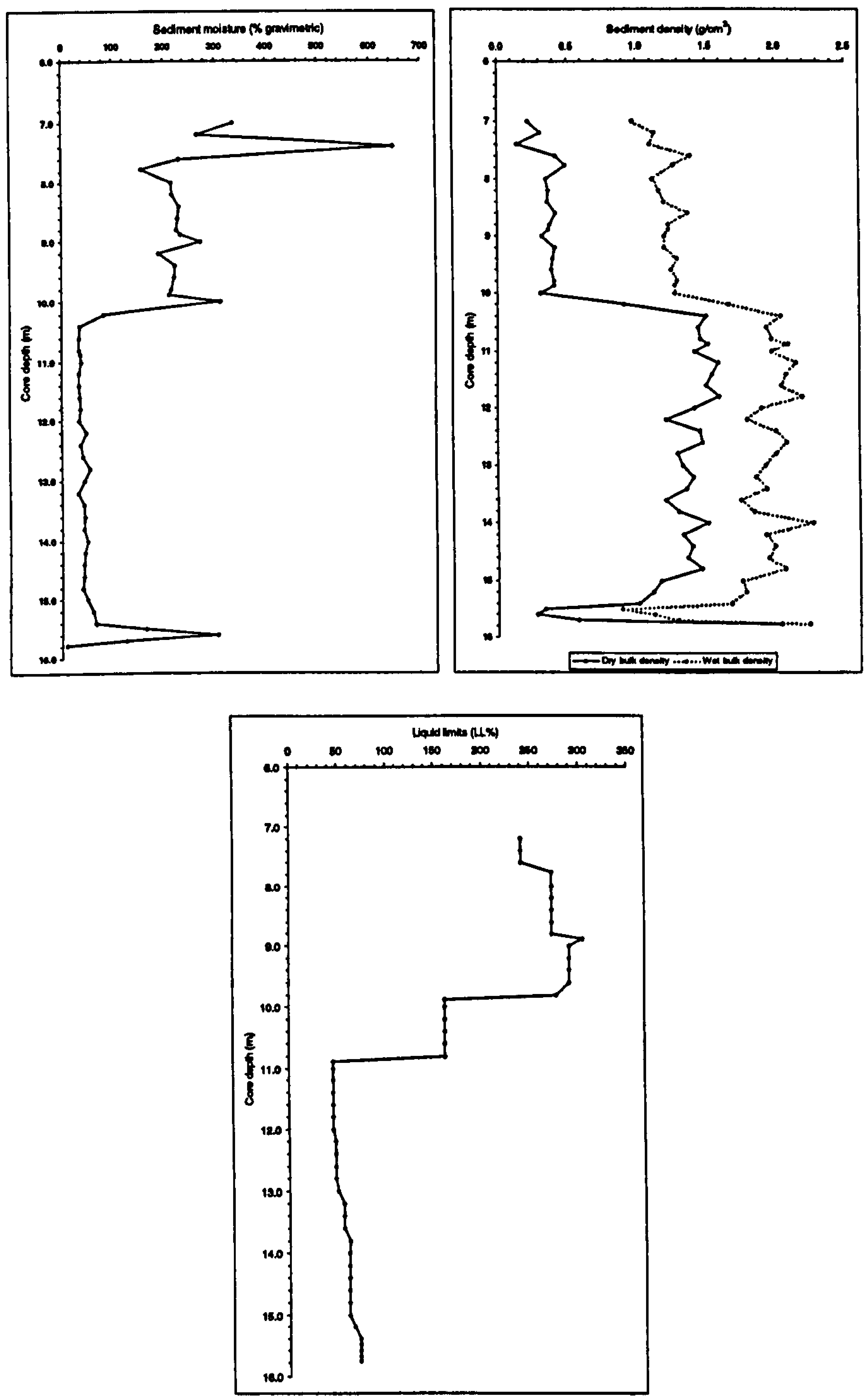


Figure 5.21. Input geotechnical data from core NS-97-2. 50 layers sampled between 0.07 and 0.23 m thick with the majority set at 0.20 m. Sediment moisture (%), dry (solid line) and wet (broken line) bulk density of the sediment (g/cm^3), and liquid limits (LL, %) results. LL results are interpolated. Changes in LL are not matched to changes in sediment type from the core log. Groundwater level = 1.0 m below ground level.

This core contains saturated organic sediment, e.g., >300% sediment moisture in basal peat at 15.6 m and 150 – 650% in the uppermost peat sections. By contrast, in-core minerogenic sediment ranges from 30 – 80% moisture content. Gravel facies have high dry bulk density, e.g., 2 g/cm^3 in the basal gravel compared with 0.3 g/cm^3 in peat. Saturated organic sediments record higher liquid limits, e.g., >240% in peat, in contrast to 160% in minerogenic-peat and 45 – 75% in minerogenic sediment.

Additional data (Figure 5.22)

Void ratio is greatest in saturated organic sediment, e.g., 17 in the overlying peat compared with 0.3 in the basal gravel. Submerged unit weight is greater in minerogenic (0.9 – 2.5 kN/m^3) than organic (peat) sediment (0.1 – 0.8 kN/m^3). Effective stress increases with depth and an inflection occurs at 10 m at the minerogenic-peat contact. Post-depositional compression (Cc) is greater in peat (1.4 – 2.6) than minerogenic sediment (0.3 – 0.6).

Total geotechnical correction for core NS-97-2 (Figure 5.23)

Autocompaction in basal sediment is quite small, e.g., 7 cm at 15.7 m core depth, and no correction was possible above the basal peat layer (15.4 m) because the wet bulk density at 15.5 m is <1 g/cm^3, resulting in a negative value for the submerged unit weight. However, over a metre of vertical displacement is recorded above the minerogenic-peat contact at 10 m and over 2 metres at 7.6 m core depth. This is due in part to the increase in liquid limits (>200%) and the coefficient of consolidation (>2) of the peat. On the whole, the rate of change is greatest around 10.8 m core depth, above which all minero-organic and organic layers have been considerably compressed. Table 5.3 contains the total correction for SLIPs obtained from core NS-97-2.

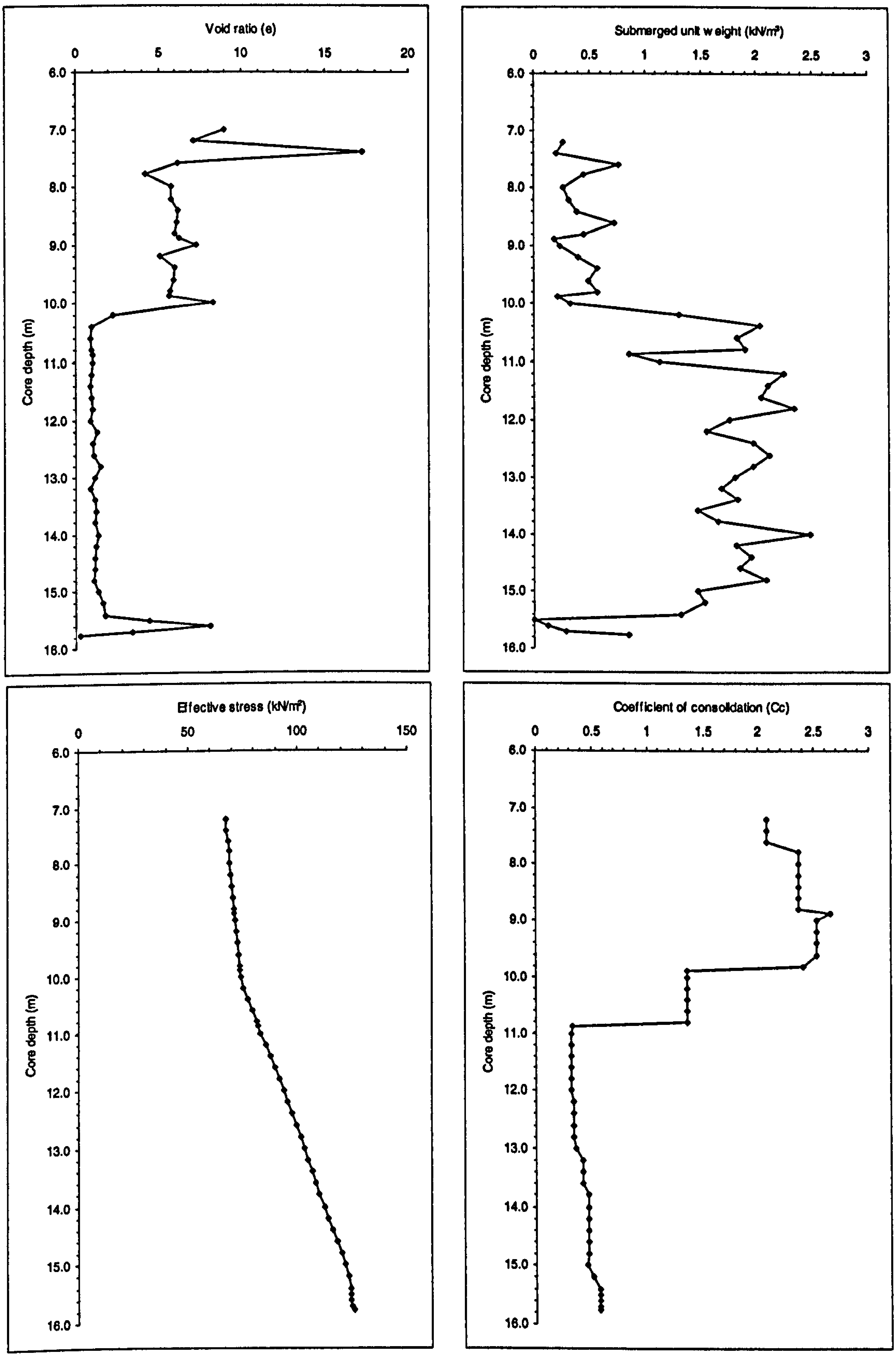


Figure 5.22. Additional geotechnical data from core NS-97-2. Void ratio (e = wG) where w = water content or soil moisture and G = grain density of 2.67. A correction cannot be calculated for partly saturated sediment located above the water table. Submerged unit weight = wet bulk density x layer thickness x 9.81. Effective stress = running sum of the submerged unit weights plus any total weight from soil above the water table. The coefficient of consolidation using Skempton's (1944) correlation (Cc = 0.009*(LL – 10) where Cc = coefficient of consolidation and LL = liquid limits.

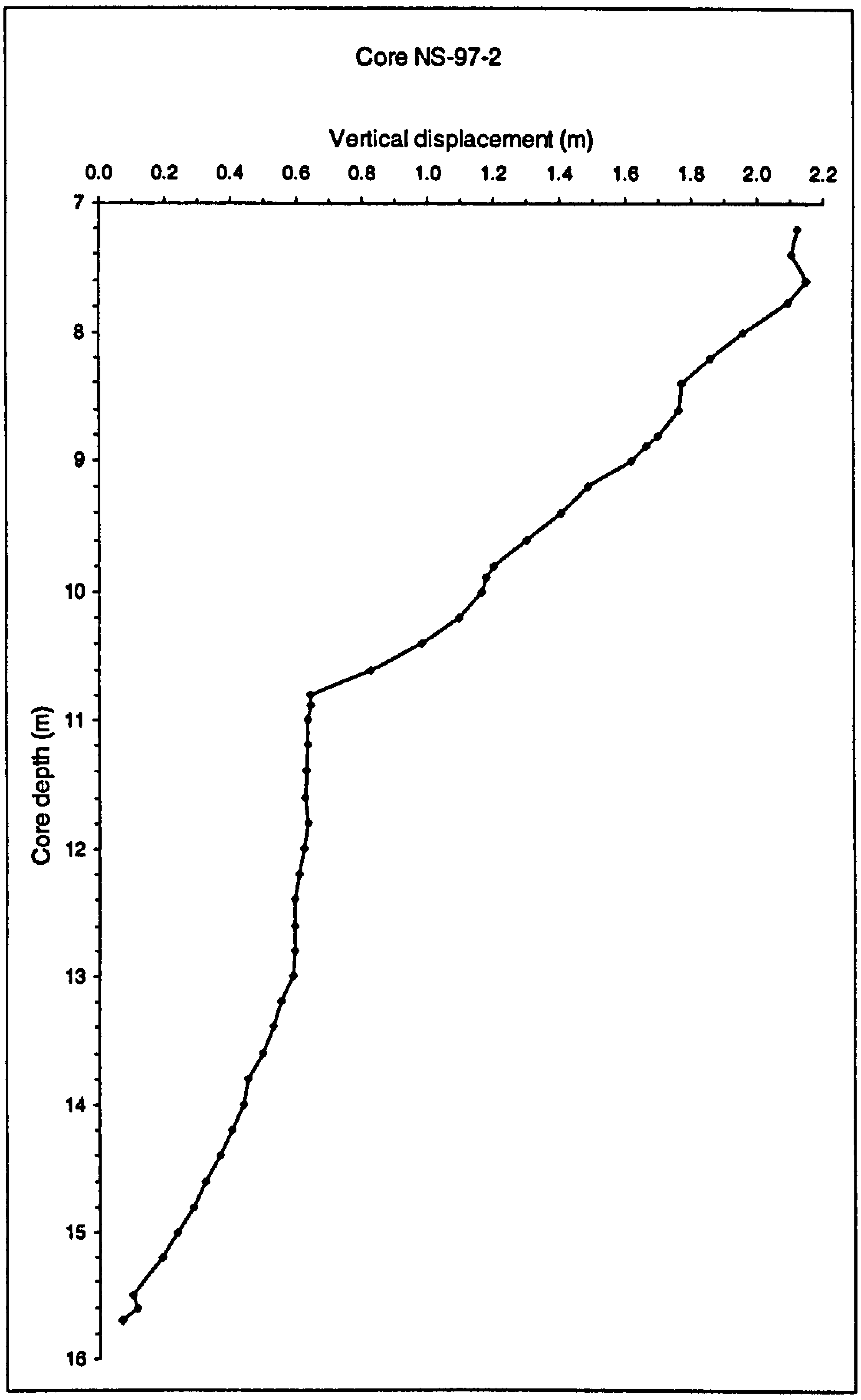


Figure 5.23. Total geotechnical correction for core NS-97-2 (ΔH in Figure 2.5). This graphs the total downward movement after the deposition of each sediment layer in the core.

5.6.2 Core NS-97-3

Input data (Figure 5.24)

Moisture content is low in the basal gravel, e.g., 15 – 50% from 11.3 – 9.7 m, and very high in the overlying peat (150 – 845%). Dry bulk density is highest in the basal gravel (1.8 g/cm^3) and lowest in the overlying peat (0.1 g/cm^3). The liquid variability of the overlying peat was immeasurable (drop-cone penetration values were often >25 – 30 mm) and therefore inadmissible under BS 1377 (1975). The liquid limits of basal minerogenic sediment are around 30 – 60% whilst minerogenic-peat reaches 85%.

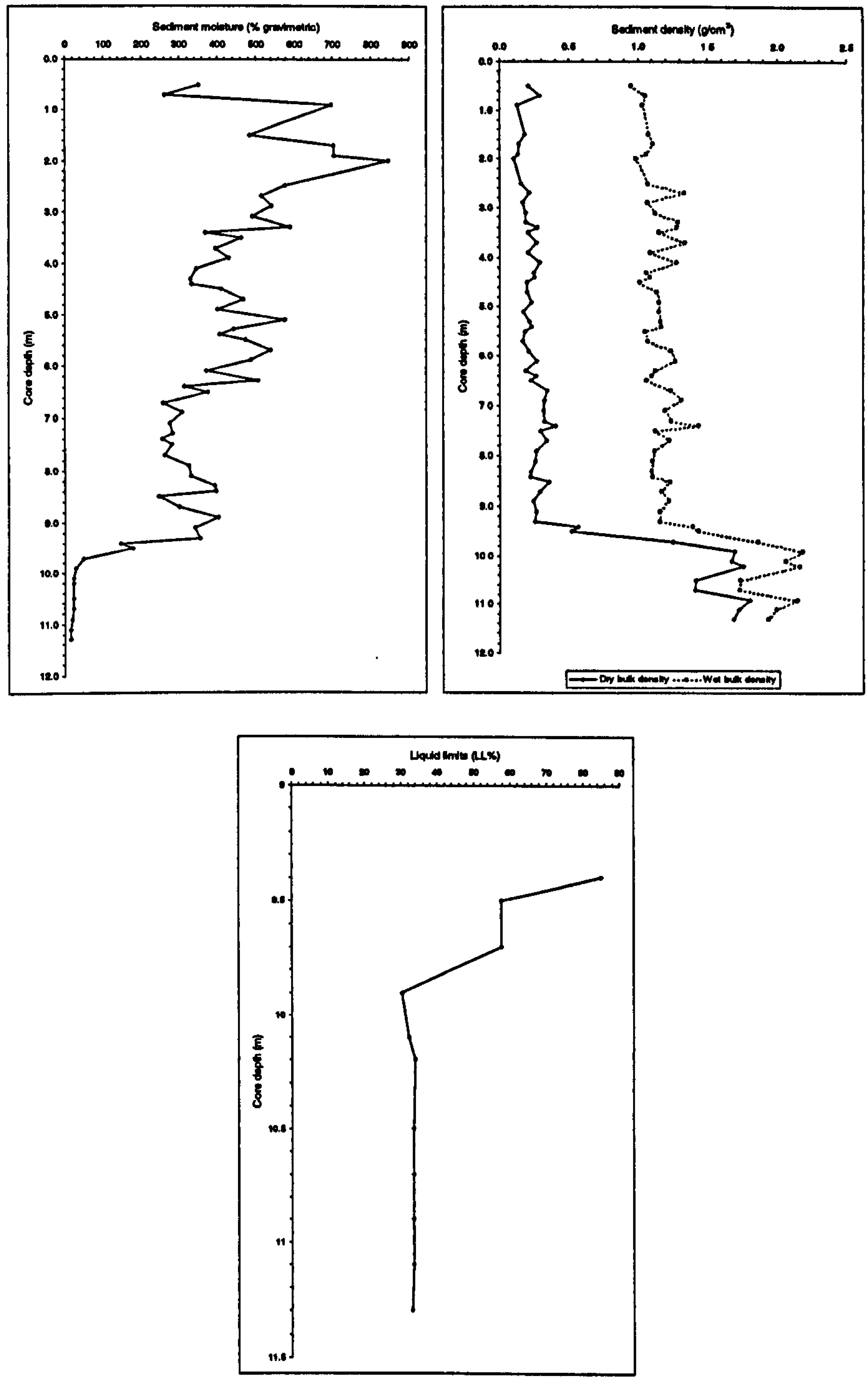


Figure 5.24. Input geotechnical data from core NS-97-3. 59 layers sampled between 0.1 and 0.6 m thick, with the majority set at 0.20 m. See Figure 5.21 caption for explanation.

Additional data (Figure 5.25)

Void ratio is again much greater in peat (23 at 2 m core depth) than minerogenic sediment (0.4 at 11.3 m). Although values of submerged unit weight below 7 m core depth are similar to NS-97-2 (Figure 5.22), the results above ground water level (1 m) are much higher (2.0 – 4.7 kN/m^3). The effective stress profile has two inflections around 1 m (ground water level) and 9.4 m (minerogenic-peat contact). Post-depositional compression (Cc) is greater in minerogenic-peat (0.4 – 0.7) than minerogenic sediment (~0.2). Cc could not be calculated above 9.4 m due to the inadmissibility of the liquid limits data.

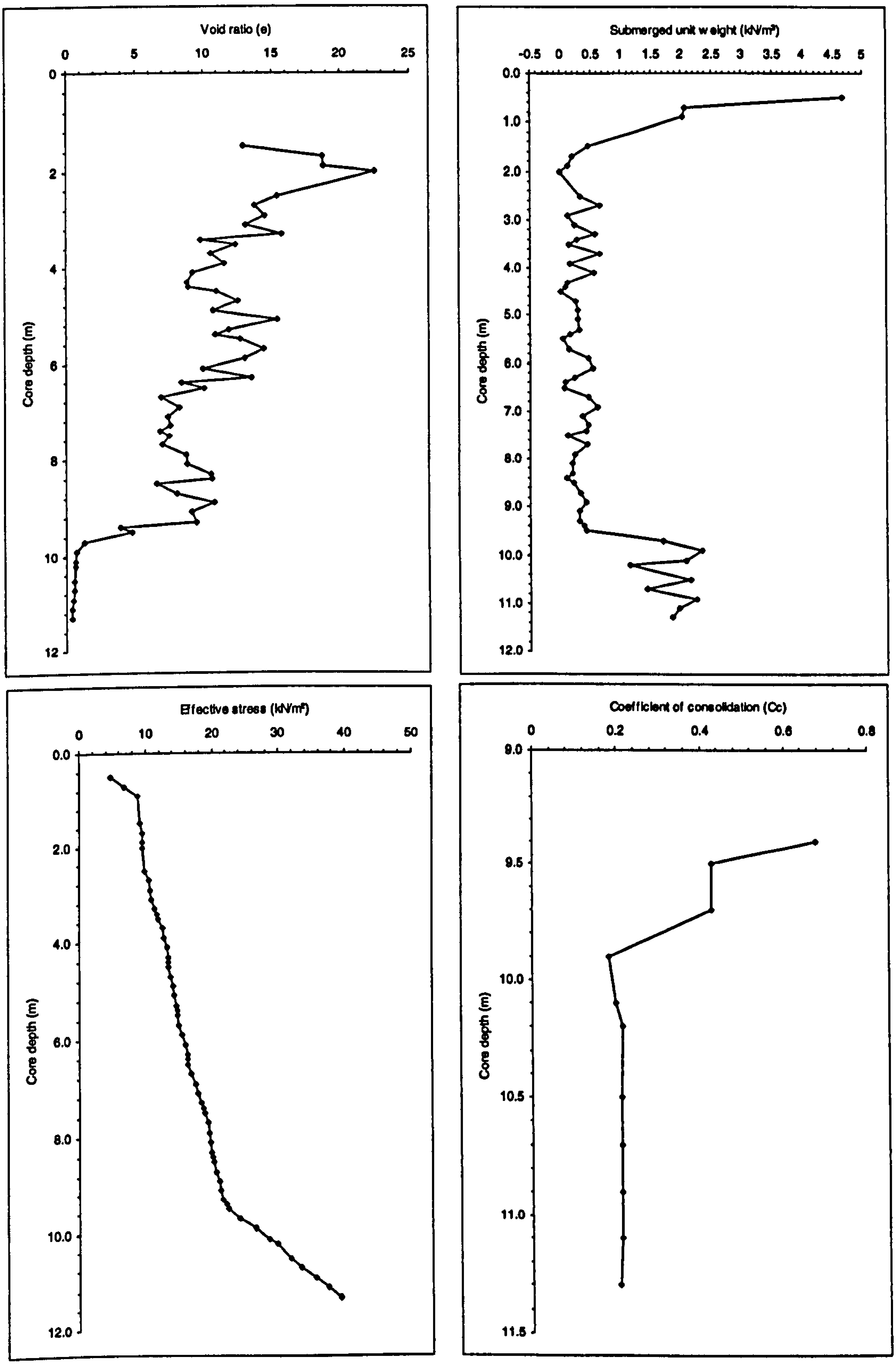


Figure 5.25. Additional geotechnical data from core NS-97-3. See Figure 5.22 caption for explanation.

Total geotechnical correction for core NS-97-3 (Figure 5.26)

The minerogenic facies of core NS-97-3 contains large gravel particles (Figure 5.6), hence autocompaction at the base is very small, e.g., 4 cm at 11.10 m core depth. This increases to 16 cm directly below the minerogenic-peat contact at 9.40 m core depth. No correction was possible above 9.40 m because liquid limits and the coefficient of consolidation of peat layers could not be determined. Table 5.3 contains the total correction for SLIPs obtained from core NS-97-3.

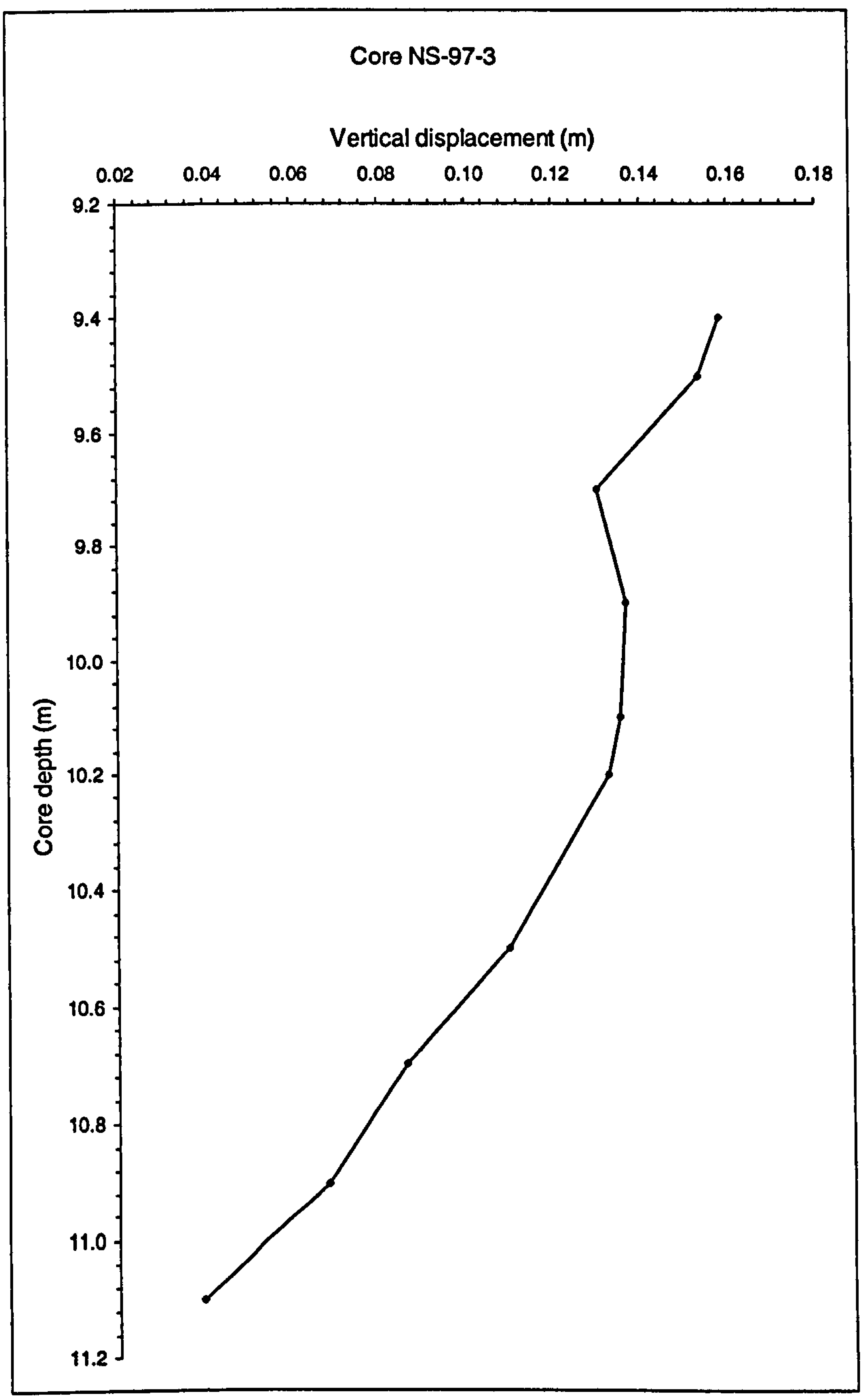


Figure 5.26. Total geotechnical correction for core NS-97-3 (ΔH in Figure 2.5). See Figure 5.23 caption for explanation.

5.6.3 Summary and limitations of geotechnical correction

The total geotechnical correction for North Sands, in particular core NS-97-2, has important implications for the elevation of SLIPs. However, the validity of the correction is entirely dependent on the quality of both the input and model-generated data. Liquid limits in North Sands cores reach very high values (>200%) that are outside the usual range for the coefficient of consolidation index (Skempton, 1944). This may result in corrections being too low for specific layers. The core loss recorded in boreholes results in a lack of input data and therefore the total correction may again be seriously underestimated. Similarly, if the groundwater level has always been greater than 1 m below ground level then effective stresses and compressions will be greater. The results shown are therefore considered a minimum correction for North Sands cores.

5.7 Chronology

5.7.1 North Sands sea-level index points

North Sands chronology is determined from fifteen age measurements by AMS ^{14}C dating the basal and in-core sediment of cores NS-97-1, 2, 3 and 4 (Table 5.3 and Figure 5.27). Basal SLIPs generated from cores NS-97-1 and 2 date the onset of marine conditions at North Sands *ca.* 8400 to 7800 cal years BP (Figures 5.11 and 5.15). The age determination from the base of core NS-97-4 dates the cessation of marine influence at the site *ca.* 4800 to 4400 cal years BP (Figure 5.19). However, concern is raised with regard to the indicative meaning of the sample (–1.17 m MTL, Table 5.3) and that it represents an inter-tidal mudflat. Three contiguous samples contain low numbers and concentrations of calcareous foraminifera (21, 15 and 73 animals per 5 ml, Figure 5.19). The flora is indicative of higher marsh, e.g., *Juncus* spp., *Phragmites* spp., and freshwater bryophytes in large quantities (>100 specimens per 5 ml sample, Appendix 2a). It is possible that the foraminifera have been re-deposited in a freshwater marsh setting at the edge of the marsh.

If the depositional environment was higher or a freshwater marsh *ca.* 4600 cal years BP then SLIP 17 would plot lower than its current position (Figure 5.27).

Index number	3	4	5	6
Radiocarbon laboratory number	AA-38822	AA-38823	AA-38824	AA-38825
Core	NS-97-1	NS-97-2	NS-97-2	NS-97-3
Material	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)	Bulk sediment (minerogenic-peat)	Bulk sediment (minerogenic)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	41.22 ± 0.32	39.77 ± 0.29	40.01 ± 0.29	37.71 ± 0.29
^{14}C age (years BP $\pm 1\sigma$)	7119 ± 63	7408 ± 59	7359 ± 59	7835 ± 61
Calibrated BP $\pm 1\sigma$ age ranges	(7974-7925) (7899-7866)	(8329-8256) (8251-8232) (8219-8169)	(8192-8150) (8140-8125) (8123-8110) (8063-8043)	(8675-8674) (8645-8585) (8579-8542)
Calibrated BP $\pm 2\sigma$ age ranges	(8108-8093) (8031-7814) (7813-7790)	(8355-8146) (8143-8108) (8088-8037)	(8333-8013)	(8977-8967) (8951-8938) (8933-8917) (8898-8882) (8864-8830) (8779-8449) (8435-8433)
Median calibrated age (cal yrs BP)	7885	8183	8172	8597
Carbon content (% by weight)	6.1	3.2	1.8	0.05
$\delta^{13}C \pm 0.1$ (‰)	-28.60	-28.40	-28.10	-24.50#
MTL sample height (m)	-13.75	-13.08	-12.78	-8.54
Miliammina fusca	0	0	8	0
Eggerella scabra	0	0	0	2
Haplophragmoides spp.	0	1	0	0
Trochammina inflata	0	0	5	0
Jadammina macrescens	0	38	15	1
Elphidium spp.	0	0	0	1
Elphidium williamsoni	1	0	0	0
Haynesina germanica	13	0	3	1
Ammonia beccarii	2	1	0	0
Indicative meaning (m MTL)	-0.91 ± 0.29	2.23 ± 0.29	1.82 ± 0.29	0.57 ± 0.29
Autocompaction (m)	0	0.07	0.15*	0.04
SLIP (m MTL)	-12.84 ± 0.29	-15.24 ± 0.29	-14.45 ± 0.29	-9.07 ± 0.29

Table 5.3. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 5.27, 5.28, 5.30 and 5.31. # estimated $\delta^{13}C$ value – insufficient sample material for an independent $\delta^{13}C$ measurement. * denotes an absolute minimum due to the unavailability of data for the sediment layer at 15.40 m (see section 5.6.1). See Table 5.1 for MTL datum.

Index number	7	8	9	10
Radiocarbon laboratory number	AA-38826	AA-38827	AA-38828	AA-38829
Core	NS-97-3	NS-97-3	NS-97-3	NS-97-3
Material	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	29.74 ± 0.27	31.51 ± 0.31	38.84 ± 0.26	38.01 ± 0.29
^{14}C age (years BP $\pm 1\sigma$)	9741 ± 73	9277 ± 79	7597 ± 54	7770 ± 61
Calibrated BP $\pm 1\sigma$ age ranges	(11202-11156) (11148-11140)	(10577-10569) (10561-10360) (10344-10328) (10320-10306) (10302-10285)	(8412-8366)	(8597-8449) (8436-8433)
Calibrated BP $\pm 2\sigma$ age ranges	(11234-11066) (10940-10860) (10823-10807) (10796-10793)	(10687-10649) (10642-10605) (10603-10237)	(8451-8333) (8226-8222)	(8645-8410)
Median calibrated age (cal yrs BP)	11173	10441	8389	8582
Carbon content (% by weight)	0.1	0.3	0.4	0.3
$\delta^{13}C \pm 0.1$ (‰)	-24.50	-24.70	-25.20	-24.90
MTL sample height (m)	-8.42	-8.24	-8.09	-8.04
Miliammina fusca	1	0	0	0
Trochammina inflata	2	0	0	0
Jadammina macrescens	4	3	8	0
Elphidium spp.	0	0	0	2
Haynesina germanica	0	0	0	1
Indicative meaning (m MTL)	2.33 ± 0.29	2.20 ± 0.29	2.20 ± 0.29	0.84 ± 0.29
Autocompaction (m)	0.05	0.07	0.09	0.09
SLIP (m MTL)	-10.70 ± 0.29	-10.37 ± 0.29	-10.20 ± 0.29	-8.79 ± 0.29

Table 5.3 continued. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 5.27, 5.28, 5.30 and 5.31. See Table 5.1 for MTL datum.

Index number	11	12	13	14
Radiocarbon laboratory number	AA-38830	AA-38831	AA-38832	AA-38833
Core	NS-97-3	NS-97-3	NS-97-3	NS-97-3
Material	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	42.76 ± 0.28	36.65 ± 0.26	34.96 ± 0.27	41.90 ± 0.30
^{14}C age (years BP $\pm 1\sigma$)	6824 ± 53	8063 ± 56	8443 ± 63	6989 ± 58
Calibrated BP $\pm 1\sigma$ age ranges	(7683-7610) (7596-7592)	(9028-8989) (8824-8823)	(9527-9465) (9453-9429) (9335-9333)	(7926-7900) (7865-7742)
Calibrated BP $\pm 2\sigma$ age ranges	(7785-7782) (7751-7574)	(9236-9221) (9127-9101) (9090-8927) (8921-8849) (8832-8773)	(9538-9398) (9382-9372) (9358-9347) (9342-9298)	(7940-7886) (7881-7673)
Median calibrated age (cal yrs BP)	7667	9008	9481	7814
Carbon content (% by weight)	0.5	0.4	0.3	0.6
$\delta^{13}C \pm 0.1$ (‰)	-25.40	-24.60	-24.80	-25.30
MTL sample height (m)	-7.89	-7.59	-7.47	-7.23
Miliammina fusca	0	8	0	0
Trochammina inflata	0	5	0	0
Jadammina macrescens	0	177	3	2
Brizalina spp.	3	0	0	5
Elphidium spp.	8	0	0	8
Elphidium oceanensis	2	0	0	0
Elphidium williamsoni	2	0	0	2
Haynesina germanica	26	0	0	104
Ammonia beccarii	4	0	0	5
Asterigerinata mamilla	2	0	0	0
Indicative meaning (m MTL)	-0.85 ± 0.29	2.20 ± 0.29	2.20 ± 0.29	-1.03 ± 0.29
Autocompaction (m)	0.11	0.13	0.14	0.14
SLIP (m MTL)	-6.93 ± 0.29	-9.66 ± 0.29	-9.53 ± 0.29	-6.06 ± 0.29

Table 5.3 continued. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 5.27, 5.28, 5.30 and 5.31. See Table 5.1 for MTL datum.

Index number	15	16	17
Radiocarbon laboratory number	AA-38834	AA-38835	AA-38836
Core	NS-97-3	NS-97-3	NS-97-4
Material	Bulk sediment (minerogenic)	Bulk sediment (minerogenic)	Bulk sediment (minerogenic-peat)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	45.52 ± 0.28	44.85 ± 0.36	59.81 ± 0.44
^{14}C age (years BP $\pm 1\sigma$)	6323 ± 50	6442 ± 65	4129 ± 59
Calibrated BP $\pm 1\sigma$ age ranges	(7305-7302) (7288-7280) (7271-7233) (7219-7211)	(7427-7305) (7303-7287) (7281-7272)	(4822-4749) (4733-4566) (4562-4529)
Calibrated BP $\pm 2\sigma$ age ranges	(7412-7397) (7367-7352) (7325-7159) (7111-7097)	(7461-7448) (7433-7250)	(4833-4505) (4487-4443)
Median calibrated age (cal yrs BP)	7252	7369	4615
Carbon content (% by weight)	2.9	4.8	44
$\delta^{13}C \pm 0.1$ (‰)	-26.30	-27.20	-29.30
MTL sample height (m)	-6.93	-6.78	-4.31
Trochammina ochracea	2	0	0
Jadammina macrescens	5	5	0
Brizalina spp.	1	0	4
Haynesina germanica	56	32	6
Ammonia beccarii	18	0	3
Unidentified calcareous	0	0	2
Indicative meaning (m MTL)	-0.69 ± 0.29	-0.81 ± 0.29	-1.17 ± 0.29
Autocompaction (m)	0.15	0.16	0
SLIP (m MTL)	-6.09 ± 0.29	-5.81 ± 0.29	-3.14 ± 0.29

Table 5.3 continued. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 5.27, 5.28, 5.30 and 5.31. See Table 5.1 for MTL datum.

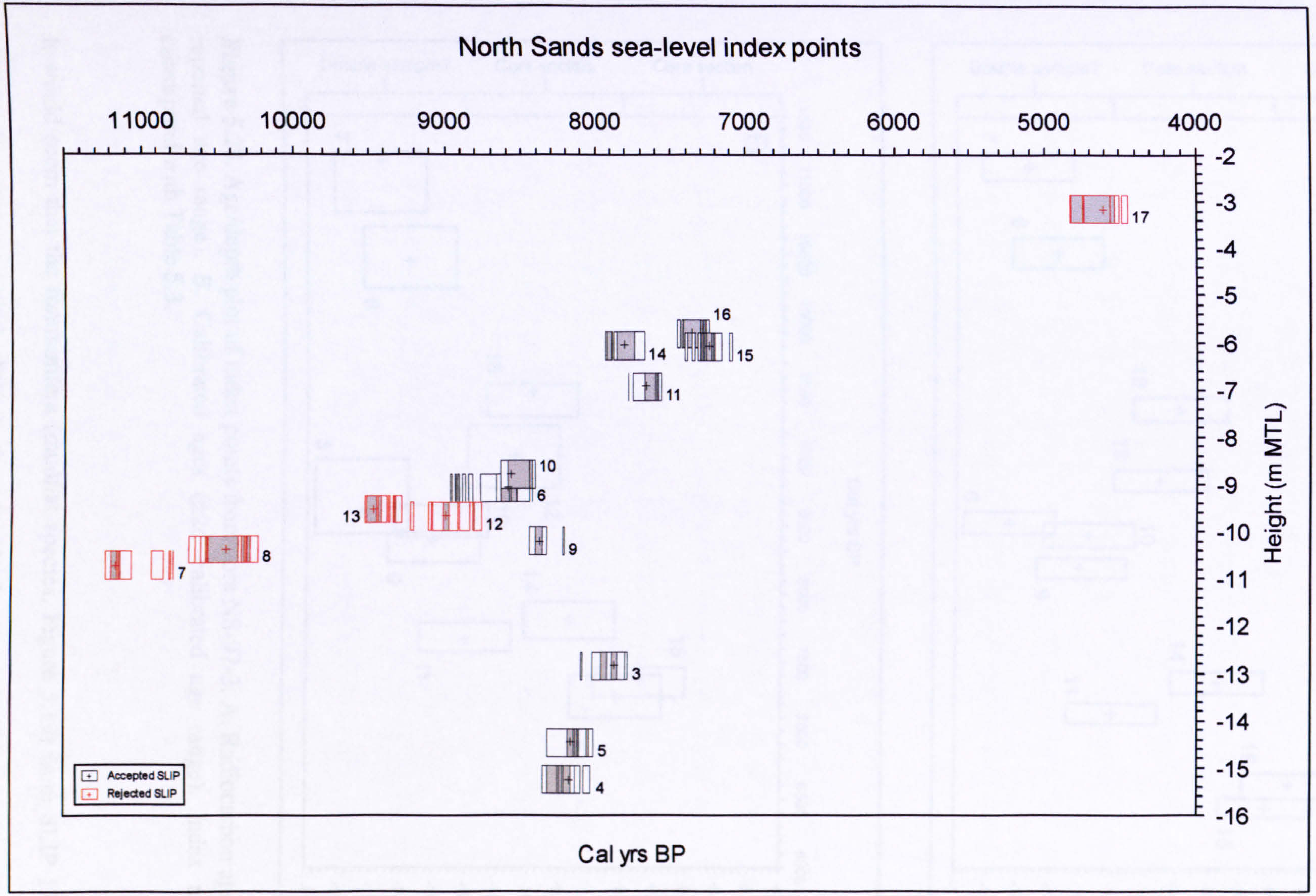
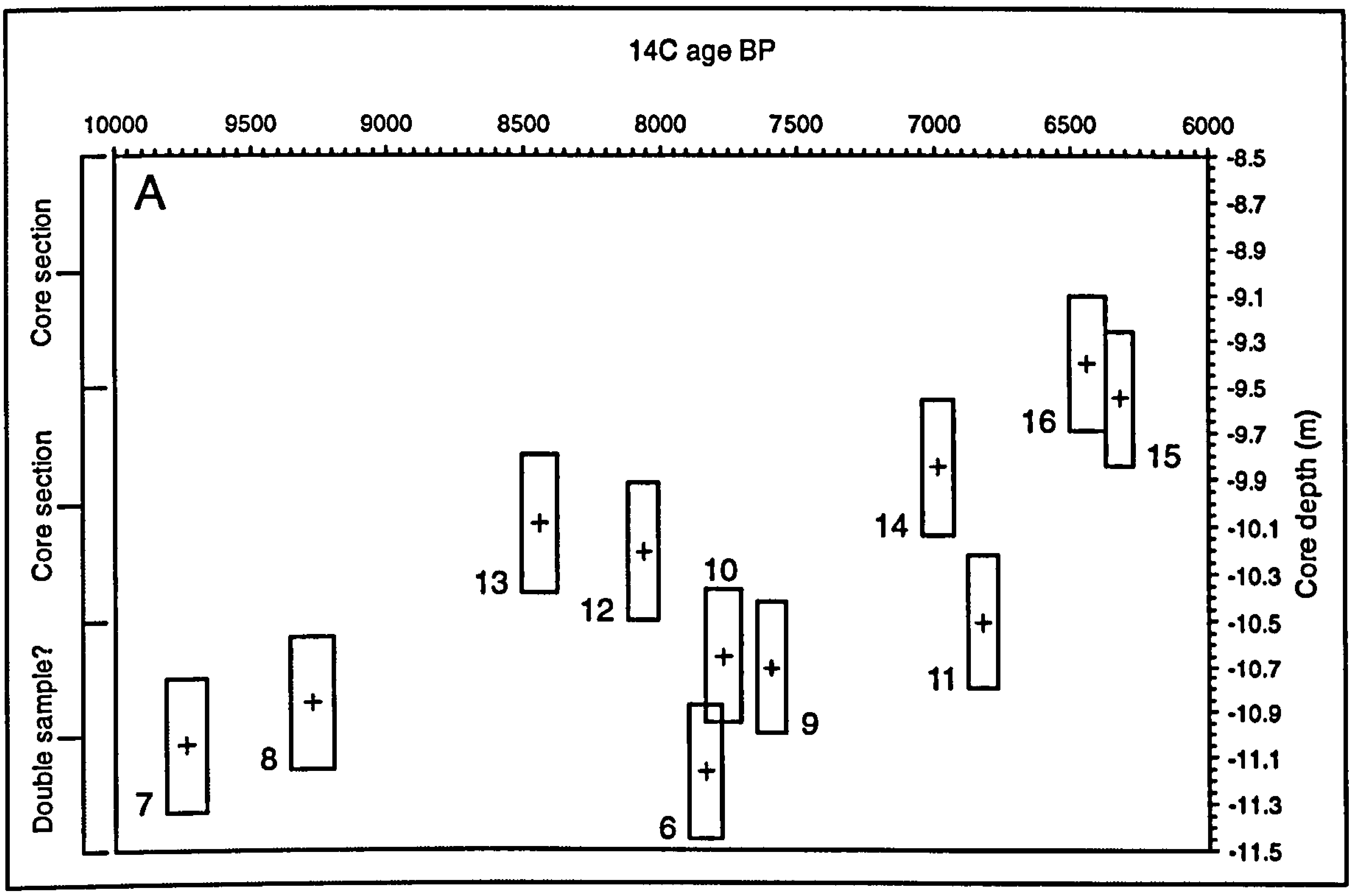


Figure 5.27. North Sands SLIPs. Core NS-97-1 (SLIP 3), core NS-97-2 (SLIPs 4 and 5), core NS-97-3 (SLIPs 6 to 16), core NS-97-4 (SLIP 17). Index numbers correspond with Table 5.3. The rationale for borderline inclusions, i.e., SLIPs 6, 9, 10, 11, 14, 15 and 16, is explained in section 5.7.2.



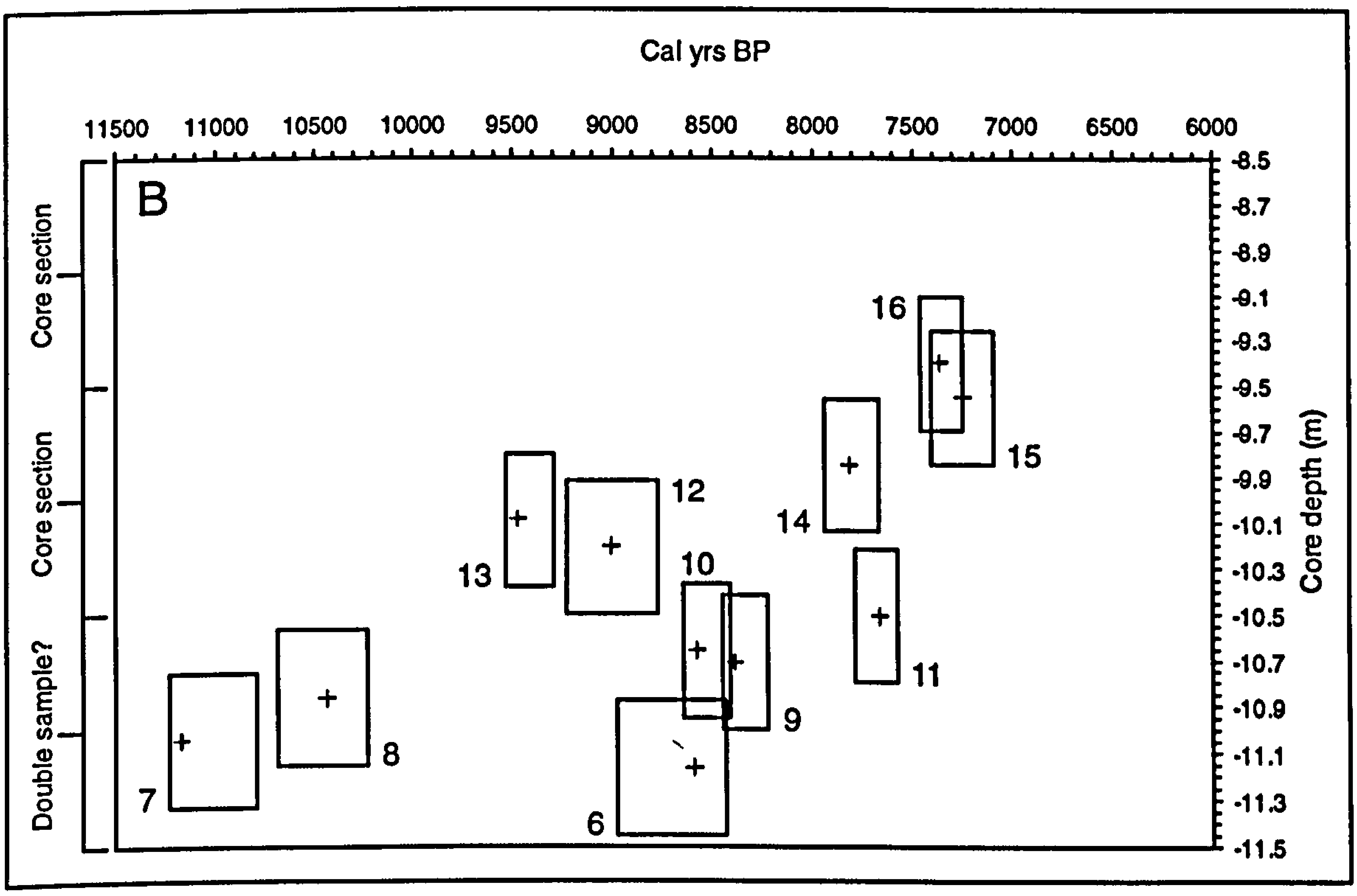


Figure 5.28. Age/depth plot of index points from core NS-97-3. A. Radiocarbon ages (±1σ reported age range). B. Calibrated ages (±2σ calibrated age range). Index numbers correspond with Table 5.3.

It would seem that the foraminifera (mudflat species, Figure 5.19) from SLIP 17 are in obvious disagreement with the lithostratigraphy (freshwater peat, Figure 5.7) and that it is not possible to determine reliably the indicative meaning of the sample. This index point is therefore rejected. Core NS-97-3 also contains problems with its litho- and biostratigraphy and more obviously its chronology (Table 5.3 and Figure 5.27). Lithofacies in the basal 2 metres of the core contain predominantly large gravel clasts in a matrix of smaller minerogenic particles (Figure 5.6). Geotechnical results, e.g., plasticity indices and mean particle size (>2mm fraction), are very similar in the two core sections (Appendix 5b). Biostratigraphical analysis reveals a striking similarity between fossil biozones C and E (salt-marsh facies) and D and F (mudflat facies) in the two lowermost basal sections (Figure 5.18 and section 5.4.3). Age/depth plots of SLIPs acquired from the core sediments show that some index points, particularly 7, 8, 12 and 13, are evidently too old and age reversals occur in the core (Figure 5.28a,b). Concerns are also raised with regard to the ages and heights of SLIPs 6, 9, 10 and 11 from the basal section of the core. These results raise the possibility that the basal portion of core NS-97-3 (11.50 – 10.50 m) is a "double sample" collected in error during coring. Pollen analysis was carried out to assess the possibility of sediment reworking in the lower core sections and further evaluate SLIPs 6, 9 and 10 in the gravel sediment and 11, 14, 15 and 16 in the finer minerogenic facies (see section 5.7.2). The analysis is also intended to shed further light on the "double sampling" hypothesis and ascertain if these SLIPs may be acceptable for sea-level studies.

5.7.2 Pollen analyses of core NS-97-3

Forty-five samples from 11.35 to 8.51 m are investigated for the presence of pollen and a CONISS cluster analysis (Grimm, 1987) identifies three main biozones (Figure 5.29). Zone A (11.35 to 9.98 m) contains a fairly mixed assemblage representing oak woodland (up to 80% *Quercus* pollen) and hazel scrub (up to 45% *Corylus* pollen). This zone is post-alder rise and, assuming conventional interpretation of pollen changes in the UK, must date post-8000 to 7500 cal years BP (Birks, 1989). Oak declines in zone B (9.98 to 9.18 m) and alder, hazel and elm increase suggesting a more mixed woodland composition (9.98 and

9.71 m). In zone C (9.18 to 8.50 m) alder dominates, oak pollen continues to decrease, and hazel pollen remains stable. Of particular importance is the possible elm decline at 8.75 m suggesting an age of *ca.* 5300 ^{14}C years BP for this horizon (Birks, 1989; Peglar, 1993; Parker *et al.*, 2002). However, only five peat samples were analysed for the presence of pollen from 9.20 – 8.50 m and only two above 8.75 m core depth (Figure 5.29 and Appendix 4a). This sampling resolution is clearly insufficient to identify confidently the stratigraphic position (core depth) of the elm decline. The uppermost sample containing elm pollen cannot therefore be reliably employed as a biostratigraphic marker here. The result is thus considered inconclusive.

The pollen stratigraphy reveals a remarkable similarity between assemblages in the basal and overlying section. For example, the pattern of change in trees and shrubs, in particular *Quercus* and *Corylus* pollen, around 10.85 – 10.50 m is mirrored in the facies from 10.25 – 9.90 m (Figure 5.29). The result is similar to that found in the foraminifera stratigraphy (Figure 5.18 and section 5.4.3) and further supports the earlier hypothesis that the basal section is a "double sample" (section 5.7.1). Some significant fluctuations in the assemblages, e.g., *Alnus* and many of the rare pollen types, show that reworking of pollen has probably taken place below 9.40 m (Figure 5.29). The basal record is similar to pollen assemblages in cores from Somerset where the dominant arboreal taxon are *Corylus*, *Quercus* and subordinate *Pinus* (Jennings *et al.*, 1998). Facies above and below SLIP 11 are dated pre-8200 cal years BP and the sequence contains 10 – 50% *Alnus* pollen but Birks (1989), Chambers and Elliot (1989) and Bennett and Birks (1990) state that this period pre-dates the expansion of *Alnus* in south-west England (Jennings *et al.*, 1998). The reliability of SLIPs 6 to 13 (*ca.* 11,200 – 8200 cal years BP) is therefore questionable although the younger 2σ cal ages of SLIPs 6, 9 and 10 are within 250 years of 8200 cal years BP (Table 5.3). Pollen from sedimentary facies containing SLIPs 7, 8, 12 and 13 are most likely reworked (Figure 5.29) and they are clearly not in chronostratigraphic order

(Figure 5.28). These index points cannot therefore be accepted as a reliable contribution to the early Holocene RSL history of south Devon.

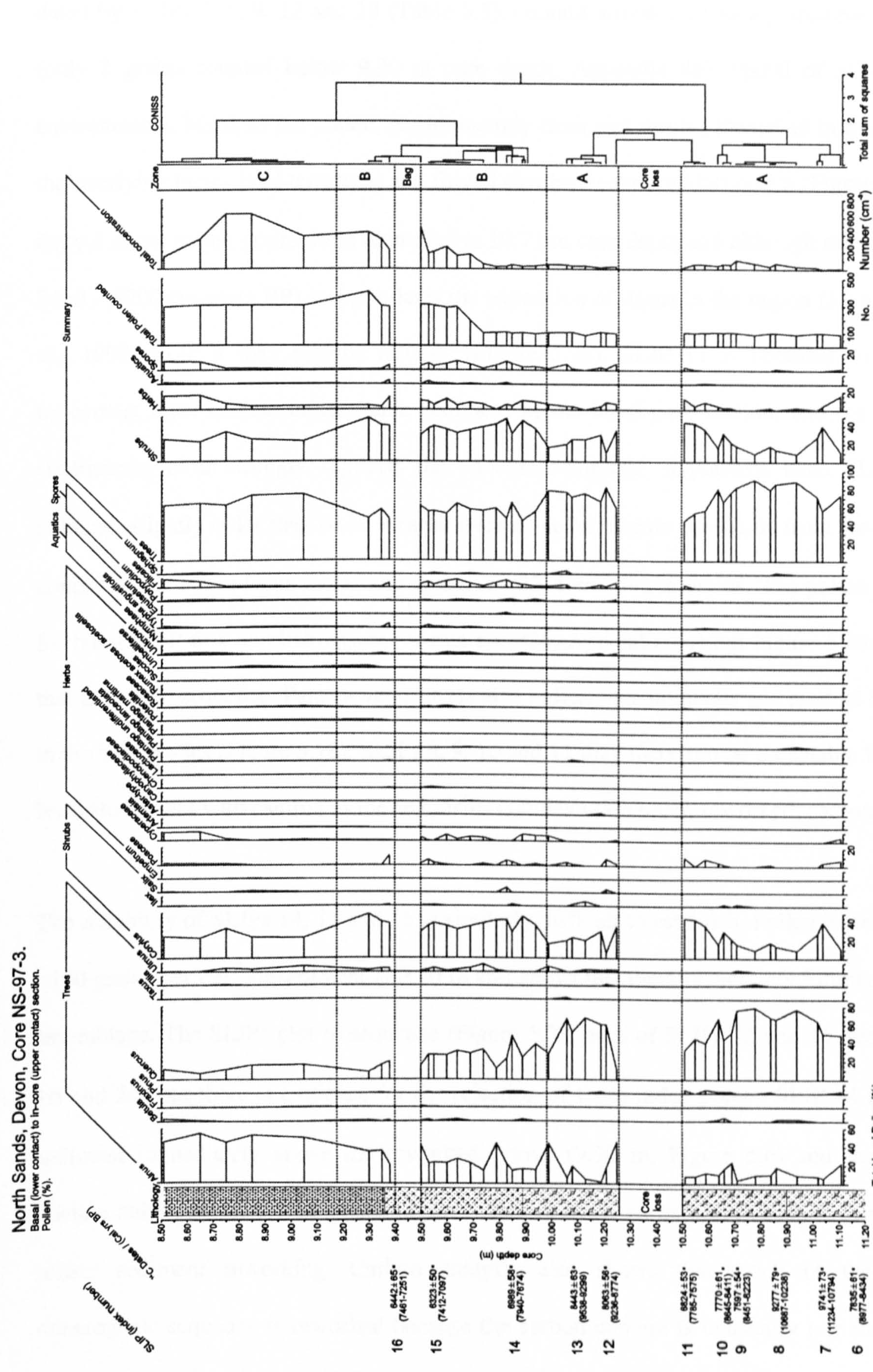


Figure 5.29. Pollen stratigraphy (%) of core NS-97-3: basal (lower contact) to in-core (upper contact) section. Samples at 11.35, 11.24 and 11.16 m core depth yielded <25 pollen grains and no pollen was found at 11.14 m (samples omitted from analysis). Frequencies calculated as % Total Land Pollen plus Aquatics and Spores.

Salt-marsh facies are identified in core NS-97-3 by the predominance of agglutinated foraminifera (Figure 5.18). It is interesting to note that the high marsh (+2 m MTL) facies, dated by SLIPs 7, 8, 9, 12 and 13 (Table 5.3), contain almost no *Chenopodiaceae* pollen (only 2 grains counted below 9.90 m core depth, Appendix 4a) typical of salt-marsh environments. Much of the pollen, predominantly trees and shrubs, identified in these and the overlying facies is of terrestrial and fluvial (freshwater/supratidal) origin (Figure 5.29). Only 4 *Alnus* pollen grains were identified at 10.71 m core depth and although the age (*ca.* 8450 – 8200 cal years BP) just pre-dates the expansion of *Alnus* in the region (Jennings *et al.*, 1998) SLIP 9 may still be reliable (Figure 5.28). SLIP 11 is obtained from the uppermost layer of fine minerogenic sediment in the basal core section, and the pollen stratigraphy most strongly supports the "double sampling" hypothesis here. SLIP 10, although slightly older than SLIP 9 (ages overlap at 2σ, Figure 5.28b), is from the gravel contact at the base of this facies and contains <25 *Alnus* pollen grains. The pollen record for basal SLIP 6 is unreliable (total counts are <25, Appendix 4a) but Figure 5.28 shows that it plots in sequence. Palaeo-water levels also indicate a continuous and gradual change in the basal facies (Figure 5.18). SLIPs 6, 9, 10 and 11 may therefore be acceptable for sea-level studies but with caution as the sediments could contain older, reworked, carbon.

The reliability of SLIPs 14, 15 and 16 is similarly difficult to establish. Pollen increases to >100 grains per cm^{-3} of sediment at 9.70 m and above this there is more conformity in the assemblage. The SLIPs plot in sequence (Figure 5.28, ages of SLIPs 15 and 16 overlap at 1σ and 2σ) but there is evidence for the rejection of these index points. SLIP 14 is from sediment immediately above the reworked gravel (9.90 m, Figure 5.6) and may also contain older carbon. Pollen assemblages in zone A and B show some variability and may reflect sediment reworking. Carbon analysis also shows that much of the basal minerogenic sequence is reworked because the carbon content is unusually low in marsh facies identified from foraminiferal analyses (Figure 5.20). The 'spike' in *Alnus* around

10.00 and 9.80 m suggests that reworking has taken place there. SLIPs 15 and 16 are from the top and bottom, respectively, of two (1 m) core sections and may contain older sediment. Furthermore, the ages of these SLIPs are reversed and, although estimates overlap at 1σ and 2σ, this could be interpreted as further evidence that the entire minerogenic sequence of core NS-97-3 has been reworked.

The greatest conformity occurs in the uppermost assemblages of zone B and especially the overlying peat sequence of zone C where the possible *Ulmus* decline is recorded. If further analysis was carried out and subsequently improved the reliability of this proposed biostratigraphic marker then SLIPs 11, 14, 15 and 16 could only be constrained by the pollen analysis because they must be >5300 (^{14}C years) and <8000 years old. These SLIPs may have been reworked only minimally and remain at their original elevation despite any reworking but it could be argued that it is difficult to accept them on this notion. It could also be argued that SLIPs much older than 8000 cal years BP are likely to be too old as the pollen is post-alder rise (Jennings *et al.*, 1998).

The litho- and biostratigraphical evidence presented here (Figures 5.6, 5.18 and 5.29) appears to show that the drill head has been deflected in the gravel facies and re-sampled the borehole at approximately 9.91 m core depth. Index points from the basal section in core NS-97-3 may therefore plot 0.59 m higher (Figure 5.30a,b). Non-parametric analyses (Mann-Whitney Test) of the difference between the median counts of carbon, foraminifera and pollen between facies 9.90 – 10.25 m and 10.50 – 10.85 m, on the whole, show no difference statistically between the two biofacies (Table 5.4). For example, there is a 22% (carbon) and 81% (shrub pollen and *H. germanica*) likelihood that the difference in median counts of these variables occurred by chance and the null hypothesis can therefore be accepted. However, *J. macrescens* (test result = 5.04%) and total counts of agglutinated foraminifera (test result = 3.03%) produce much lower probabilities. This is most likely due to the significant rise in *J. macrescens* (177 animals at 10.20 m, Appendix 2a) not being recorded anywhere in the lower biofacies (<10 animals per sample).

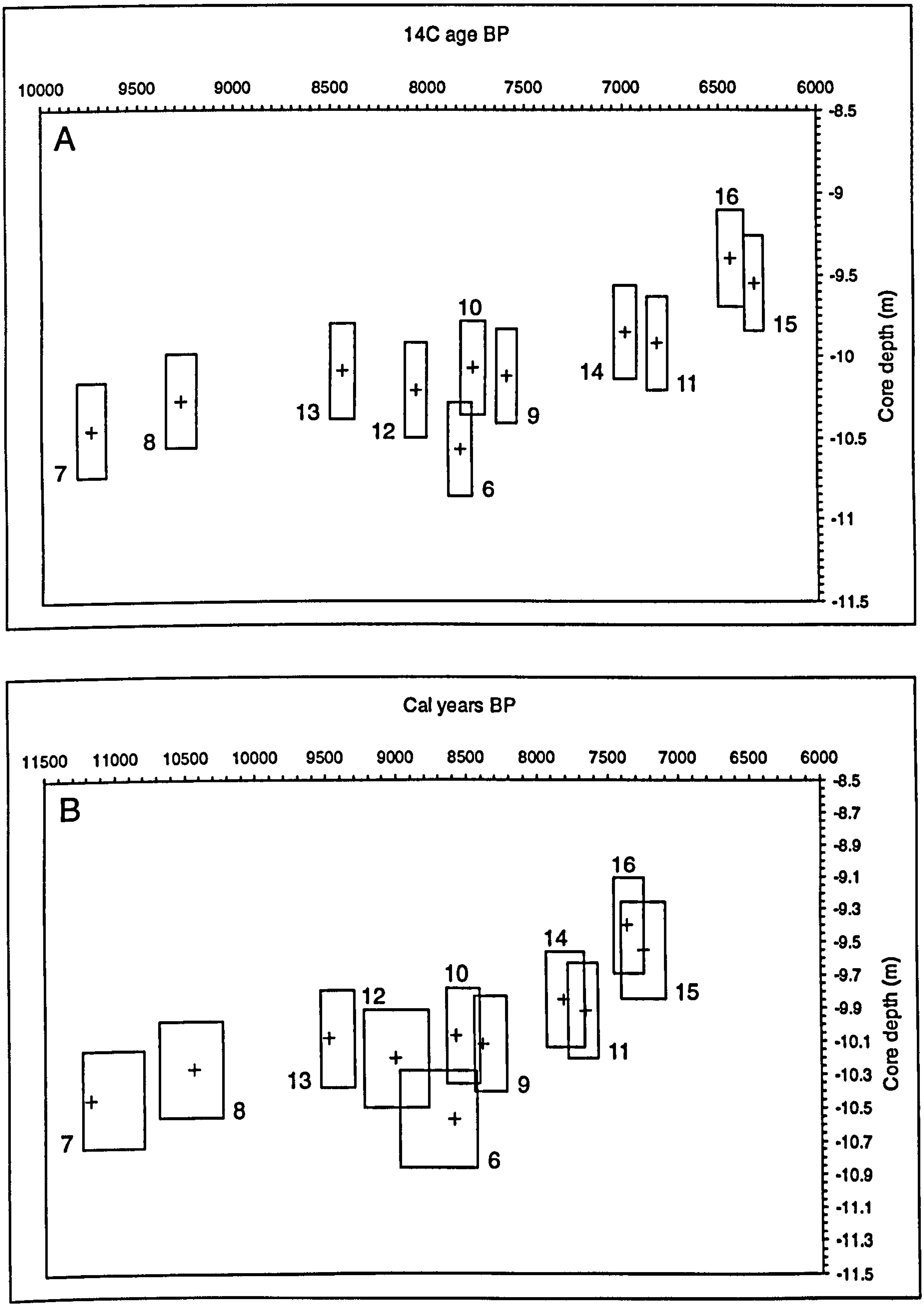


Figure 5.30. Corrected age/depth plot of index points from core NS-97-3. A. Radiocarbon ages (±1σ reported age range). B. Calibrated ages (±2σ calibrated age range). Index numbers correspond with Table 5.3.

On the whole, the statistical analyses support the earlier hypothesis that the basal section of core NS-97-3 is a "double sample" taken erroneously during coring at the site. Foraminifera (Figure 5.18) and pollen results (Figure 5.29) clearly show duplication of the 9.90 – 10.25 m biofacies at 10.50 – 10.85 m core depth. Pollen and carbon (Figure 5.20) results also show that reworking of older carbon is likely to have occurred at depths below 9.40 m, but age/depth plots (Figure 5.28a,b) question the validity of rejecting all SLIPs from core NS-97-3 on this basis. In particular, an age/depth plot (Figure 5.30a,b) that takes into account the possible "double sampling" at the base of the core provides further evidence for accepting the hypothesis. The possible elm decline at 8.75 m core depth is not accepted as a biostratigraphic marker (further biostratigraphical analysis is required) and SLIPs 6, 9, 10, 11, 14, 15 and 16 are accepted as borderline inclusions to the early Holocene relative sea-level history of south Devon.

Variable	Biofacies depth (m)	N (obs)	Median count	Point estimate (ETA1-ETA2)	95% C.I. for ETA1-ETA2	Test statistic (w)	Accept H_0 at alpha = 0.05? (95%)
Total Carbon	9.90 – 10.25	13	0.4710	-0.0790	-0.2790, 0.0481	120.5	Yes at 0.2192
	10.50 – 10.85	7	0.5220				
J. macrescens	9.90 – 10.25	13	9	6	0.00, 15.01	214.0	Yes at 0.0504
	10.50 – 10.85	13	6				
H. germanica	9.90 – 10.25	5	16	-3	-17.00, 91.01	30.5	Yes at 0.8072
	10.50 – 10.85	7	24				
Agglutinated	9.90 – 10.25	13	10	6	1.01, 18.00	227.0	No at 0.0303
Foraminifera	10.50 – 10.85	14	6				
Calcareous	9.90 – 10.25	6	24.5	-7.5	-30.02, 82.01	39.0	Yes at 0.7210
Foraminifera	10.50 – 10.85	7	38.0				
Quercus	9.90 – 10.25	8	54	-8	-30.00, 14.01	63.5	Yes at 0.4380
Pollen	10.50 – 10.85	9	52				
Corylus	9.90 – 10.25	8	22	-2	-16.00, 12.00	67.0	Yes at 0.6629
Pollen	10.50 – 10.85	9	22				
Tree Pollen	9.90 – 10.25	8	72	-1	-16.00, 15.99	68.5	Yes at 0.7708
	10.50 – 10.85	9	78				
Shrub Pollen	9.90 – 10.25	8	23	-2	-16.00, 12.00	69.0	Yes at 0.8093
	10.50 – 10.85	9	22				

Table 5.4. Non-parametric (Mann-Whitney Test) analyses of biofacies in core NS-97-3. Variables with the highest counts are chosen. Shaw and Wheeler (1994) recommend this test for use with very small samples. H_0 : There is no difference between biofacies at 9.90 – 10.25 m and 10.50 – 10.85 m core depth.

5.7.3 Summary of North Sands chronology

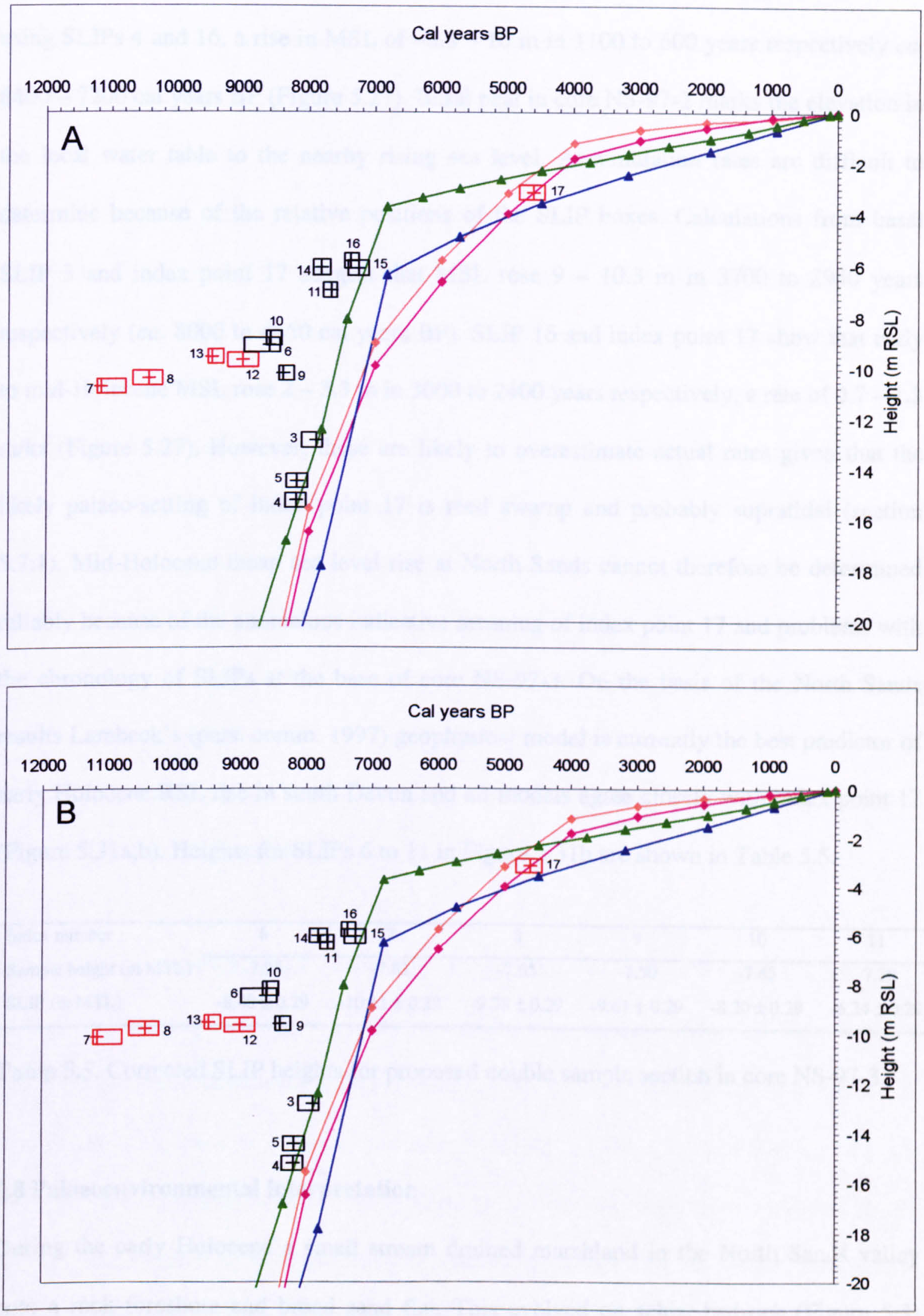


Figure 5.31. North Sands SLIPs and geophysical modelling results. A. Including uncorrected SLIPs from core NS-97-3. B. Including corrected SLIPs from core NS-97-3 (see section 5.7.2 and Figure 5.30). Index numbers correspond with Tables 5.3 and 5.5.

The data show, using basal SLIPs 3 and 4, that mean sea level rose locally by ~1.8 – 3 m in almost 600 years during the early Holocene, a rate of 3 – 5 m/ka. Rates are 8 – 17 m/ka using SLIPs 4 and 16, a rise in MSL of ~8.8 – 10 m in 1100 to 600 years respectively *ca.* 8400 – 7200 cal years BP (Figure 5.27). Basal peat in core NS-97-2 marks the elevation in the local water table to the nearby rising sea level. Accumulation rates are difficult to determine because of the relative positions of the SLIP boxes. Calculations from basal SLIP 3 and index point 17 suggest that MSL rose 9 – 10.3 m in 3700 to 2950 years respectively (*ca.* 8000 to 4450 cal years BP). SLIP 16 and index point 17 show that early to mid-Holocene MSL rose 2 – 3.3 m in 3000 to 2400 years respectively, a rate of 0.7 – 1.3 m/ka (Figure 5.27). However, these are likely to overestimate actual rates given that the likely palaeo-setting of index point 17 is reed swamp and probably supratidal (section 5.7.1). Mid-Holocene mean sea level rise at North Sands cannot therefore be determined reliably because of the anomalous indicative meaning of index point 17 and problems with the chronology of SLIPs at the base of core NS-97-3. On the basis of the North Sands results Lambeck's (pers. comm. 1997) geophysical model is currently the best predictor of early Holocene RSL rise in south Devon and all models agree closely with index point 17 (Figure 5.31a,b). Heights for SLIPs 6 to 11 in Figure 5.31b are shown in Table 5.5.

Index number	6	7	8	9	10	11
Sample height (m MTL)	-7.95	-7.83	-7.65	-7.50	-7.45	-7.30
SLIP (m MTL)	-8.48 ± 0.29	-10.11 ± 0.29	-9.78 ± 0.29	-9.61 ± 0.29	-8.20 ± 0.29	-6.34 ± 0.29

Table 5.5. Corrected SLIP heights for proposed double sample section in core NS-97-3.

5.8 Palaeoenvironmental interpretation

During the early Holocene a small stream drained marshland in the North Sands valley onto a rock foreshore and broad sand flat. This evolved on schist bedrock (Figure 5.3) covered by weathered Lower Devonian slates (Figure 5.8). A deep channel, possibly fluvial in origin, may exist in the weathered rock basin approximately –15 m OD north-east of core NS-97-1 (Figure 5.9). A supratidal environment (Figure 5.11, zone A) and nearby wood or marsh (Figure 5.15, zone B) existed at the core site before 8500 cal years BP (Figures 5.4 and 5.5). The presence of a nearby barrier-lagoon complex, similar to that proposed in Start Bay (Hails, 1975a,b) is also likely (Figures 5.6 and 5.18).

The earliest sign of a post-glacial marine transgression of North Sands is *ca.* 8400 cal years BP (SLIP 4, Figure 5.27) but the possible deep channel (Figure 5.9) may contain older marine deposits. It is interesting to note that as early as 11,000 BP salt marshes must have existed offshore of present-day North Sands. Older material (*ca.* 11,200 to 10,300 cal years BP) has probably been carried shoreward by migrating barriers under the influence of rising sea level (Figure 5.6). These barriers were possibly subtidal bar(s) that later grew in height above the tidal frame as they reached the present day shoreline. Shallow tidal mudflat or lagoonal conditions became established at the core sites between *ca.* 8400 and 4400 cal years BP when a small creek probably existed prior to barrier closure (Figure 5.8). Fringing marshes developed above fluvial deposits and weathered rock but were soon inundated by the rising sea as North Sands evolved into a complex shallow lagoon with a tidal inlet (Figures 5.11 and 5.15).

Six to seven metre thick facies (around –14 to –7 m OD, core NS-97-1 and 2, Figure 5.8) of fine and gravelly minerogenic sediment, containing calcareous foraminifera, indicate that lagoonal, tidal conditions continued throughout the early Holocene. No intercalated peat facies are present in the cores but biostratigraphical results suggest that fringing salt marshes may have developed and then disappeared under barrier-lagoon sediment during the early Holocene (Figure 5.18). Small quantities of *Chenopodiaceae* pollen are found in samples dating *ca.* 7500 to 7000 cal years BP (Figure 5.29) supporting the proposed existence of nearby salt marsh conditions. It is also possible that salt-marsh foraminifera found in the sequence could have been washed in from fringing marshes under rapidly rising seas during the early Holocene. This evidence is based on core NS-97-3 (Figure 5.6) and is tentatively proposed given the problems of chronology in that core (see sections 5.7.1 and 5.7.2).

As the beach barrier extended south-west to north-east across the bay (Figures 5.1 and 5.2) it began to restrict tidal influence and closure of the lagoon seems to have occurred suddenly. This abrupt closure is proposed because no salt marsh sediments are found in the uppermost facies of cores NS-97-1 to 4. Results (see upper contacts in section 5.4) show that shallow marine foraminifera are present in minerogenic sediment overlain by freshwater peat containing plant macrophytes (e.g., *Chara* oospores) indicative of standing water in a marsh setting. Barrier closure probably began around 5000 cal years BP and the remaining narrow creek closed by *ca.* 4400 cal years BP (Figure 5.7). Mudflat foraminifera are found re-deposited at the edge of a marsh in a supratidal palaeo-setting (Figure 5.19). Following the cessation of marine influence a back-barrier environment of reed marsh and alder carr developed (above –6.6 m OD, Figure 5.29). No washover facies are present in the lithostratigraphy but brief storm events are known to have inundated the back-barrier marsh (see section 6.1.3) and it must be assumed that similar events have occurred previously.

5.9 Summary

The lithostratigraphy at North Sands suggests that a system of subtidal barriers migrated shoreward on top of the Lower Devonian slate and fluvial deposits during the early Holocene marine transgression. Foraminiferal analysis shows that basal salt-marsh peat developed *ca.* 8400 cal years BP and was subsequently buried by seven metres of minerogenic sediment. Local MSL rose at a rate of ~3 – 5 m/ka *ca.* 8000 – 7000 cal years BP (using SLIPs 3, 4 and 5) or 8 – 17 m/ka *ca.* 8400 – 7200 cal years BP (using SLIPs 3, 4, 5, 6, 9, 10, 11, 14, 15 and 16). A probable "double sample" collected in error during coring is identified at the base of core NS-97-3 and SLIPs 6, 9, 10, 11, 14, 15 and 16 from this core are borderline inclusions to the RSL history of south Devon. Geophysical models, particularly Lambeck (pers. comm. 1997), agree very closely with SLIPs 3, 4, 5, 15 and 16. North Sands valley gradually became choked with fine minerogenic sediment and coarse gravel as the pace of sea-level rise slowed until supratidal barriers closed the narrow tidal inlet by *ca.* 4400 cal years BP (index point 17). This initiated the start of the back-barrier fresh-water reed marsh and alder carr system. The present-day back-barrier environment is protected from further sea-level rise and storm surges by the Cliff Road artificial barrier.

Chapter 6

Slapton Sands

6.1 Site description

Slapton Sands is located between SX 823 419 and SX 843 465 and takes its name from the nearby village of Slapton (Figure 6.1) meaning "slippery place" (Stanes, 1983). The site occupies the eastern shore of a coastal lake, Slapton Ley, and the western shore of Start Bay, and is in the lower reaches of several drainage systems surrounding the Ley. The upland slopes of Slapton Parish are predominantly mixed deciduous woodland and permanent grassland. The lower slopes consist mainly of pastoral and arable farmland. Archaeological research suggests that upland south-east Devon was quite densely populated throughout the early Holocene (Greig and Rankine, 1953; Smith, 1956; Anon, 1957; Masson Phillips, 1958, 1966; Rosenfeld, 1964; Simmons, 1964, 1969; Miles, 1976) but relatively smaller numbers of Mesolithic humans lived a nomadic lifestyle along the Slapton shoreline (Palmer, 1977). A significant amount of research has provided information on post-glacial sea level history (Table 6.1) and evolution of this coastline (e.g., Pengelly, 1866a; Worth, 1904, 1909, 1923; Orme, 1960; Dineley, 1961; Clarke, 1970; Morey, 1974, 1976, 1980, 1983a,b; Hails, 1975a,b; Lees, 1975; O'Sullivan, 1994).

6.1.1 Geomorphology

Slapton Sands is a gravel barrier about 5 km in length from Pilchard Cove below Strete to Torcross Point (Figure 6.2), extending a further 4 km to Lobster Rock below Hallsands (Figure 1.12). The back-barrier system is extensive coastal wetland separated into two basins. The Higher Ley is reedswamp dominated by *Phragmites communis* and the Lower Ley is a freshwater lake around 1.5 – 2.8 m deep, formed by the damming of several pre-Holocene valleys (Morey, 1983a). This is situated in the lower reaches of four main drainage basins, the River Gara, Slapton Wood, Start Stream and Stokeley Barton. The low scrub along the eastern shore of the Ley lies approximately 80 to 130 m inland from MHWST at +4 to +6 m OD and the entire wetland Reserve covers an area of 116 ha.

Radiocarbon laboratory number	Location (OSGR)	^{14}C age (years BP)	Max (Median) Min Cal age (BP) (± 2σ range)	Sample Height (m OD)	Altitude (m RSL)	Reference
SRR317[$]	N. Hallsands SX818389	1683 ± 45	1706 (1564) 1516	-1.0	-3.17 ± 0.66	Hails (1975b)
SRR164[$]	Beesands SX821410	4302 ± 45	4968 (4852) 4826	-4.32	-5.99 ± 0.73	Hails (1975b)
SRR165[$]	Beesands SX821410	4767 ± 45	5595 (5516) 5327	-4.62	-7.09 ± 0.29	Hails (1975b)
SRR237[†]	Start Bay SX825390	8108 ± 60	9261 (9025) 9006	-16 to -17	-11.89 ± 1.31	Hails (1975a)
SRR491[#]	Slapton Ley SX819440	1540 ± 40	1527 (1412) 1330	+2.0	+2.0 ± 0.50*	Heyworth and Kidson (1982)
SRR492[$]	Slapton Ley SX823439	1813 ± 40	1864 (1716) 1613	-0.19	-2.16 ± 0.56	Morey (1983a)
SRR493[$]	Slapton Ley SX823439	2889 ± 50	3207 (2996) 2868	-1.79	-4.06 ± 0.27	Morey (1983a)

Table 6.1. Radiocarbon dates and associated sample information from the Slapton coastline: Start Point (SX 830 370) to Pilchard Cove, north of Slapton Sands (SX 843 465) including Start Bay data between northings 37 and 46. [$]Index points recorded in the Holocene SLIP database (2003). [†]Index point rejected by Durham (top of basal peat possibly eroded). [#]Not included in the database. Calibrations carried out by the author of this project. *Altitude data from Heyworth and Kidson (1982) not adjusted to RSL. However, applying MSL data from Table 6.2, index point SRR491 may plot approximately –1.1 ± 0.50 m RSL.

Start Bay is a macro-tidal asymmetrical embayment covering roughly 60 km^2 and lies to the east of a former inter-tidal barrier-lagoon complex (Hails, 1975a; Morey, 1983a, Figures 6.2 and 1.12). A rocky coastal slope rises 100 m above MHWST north and south of the 9 km gravel barrier completing a Start Bay shoreline of over 27 km in length (from Start Point SX 831 370 to Berry Head SX 947 565, Figure 1.12). The neighbouring coastline to Slapton Sands consists predominantly of barrier beaches that have sealed off post-glacially drowned river valleys (Hails, 1975a,b; Morey, 1976). Shingle is banked up parallel to the Ipswichian cliffline (Orme, 1960) except where early-Holocene lagoons have now become freshwater lakes or filled with peat and silt (e.g., Blackpool Sands, chapter 7). Other local examples of back-barrier sites exist, most notably above Beesands to the west of Widdicombe Ley and at Hallsands, 4 and 6 km south of Slapton respectively (Figure 1.12). Pengelly (1869) discovered evidence for forests that once occupied these valleys and were subsequently drowned by rising seas. In particular, Blackpool Sands has revealed relict forest following the removal of sand and shingle by storm events.

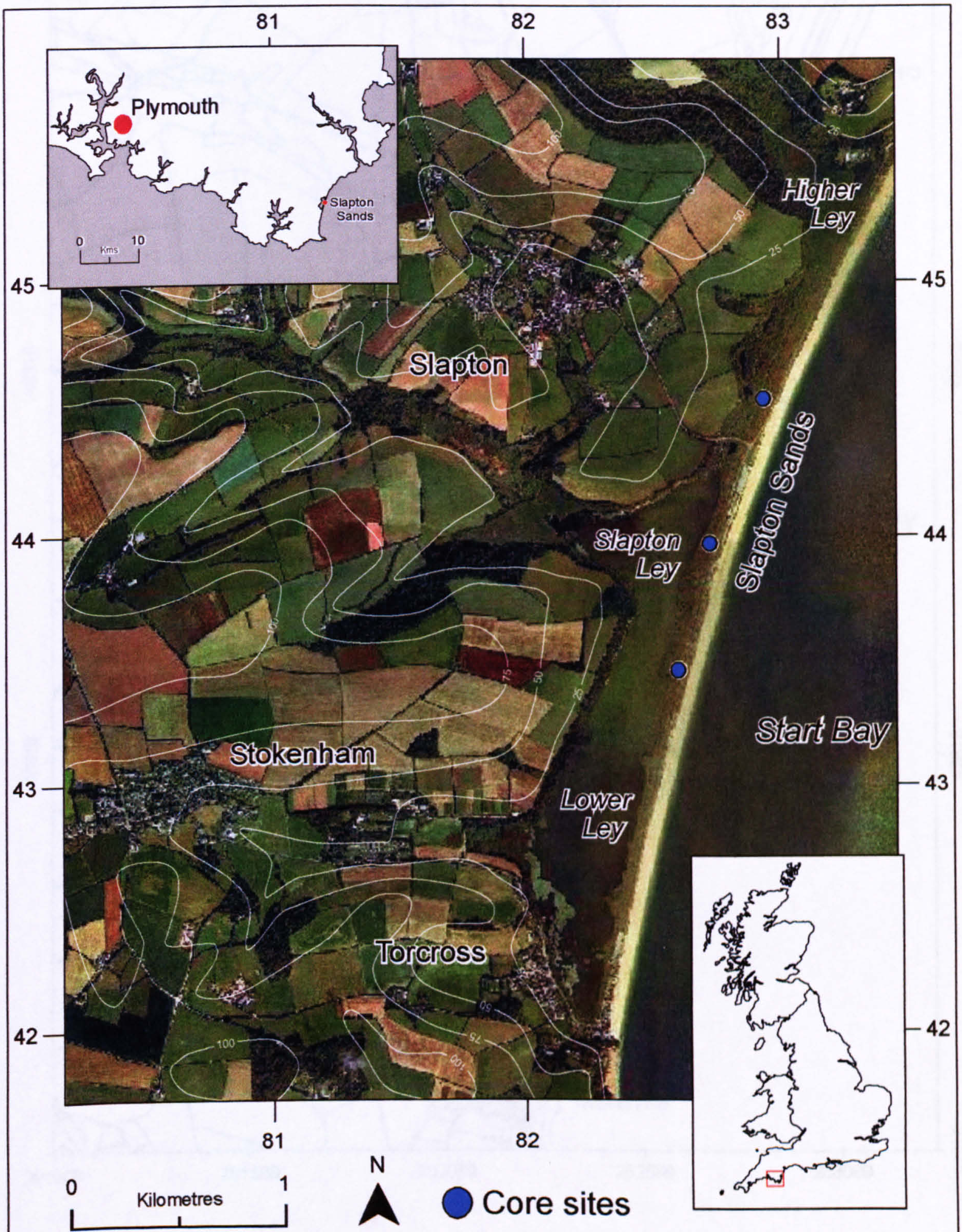


Figure 6.1. Aerial view of Slapton Sands. Source: http://www.multimap.com. Extract from Great Britain map and aerial photograph of South Hams District at scale 1:25 000 (2002). Heights in metres above MSL.

Figure 6.2. Slapton Sands back-barrier system. Borehole locations: cores SS-99-1, 2 and SS-01-3. Electrical resistivity survey lines: Sla 1 and Sla 2. Source: http://digimap.edina.ac.uk/service/advanced/carto_welcome.html. Map extract at scale 1:2500 (2002).

6.1.2 Geology

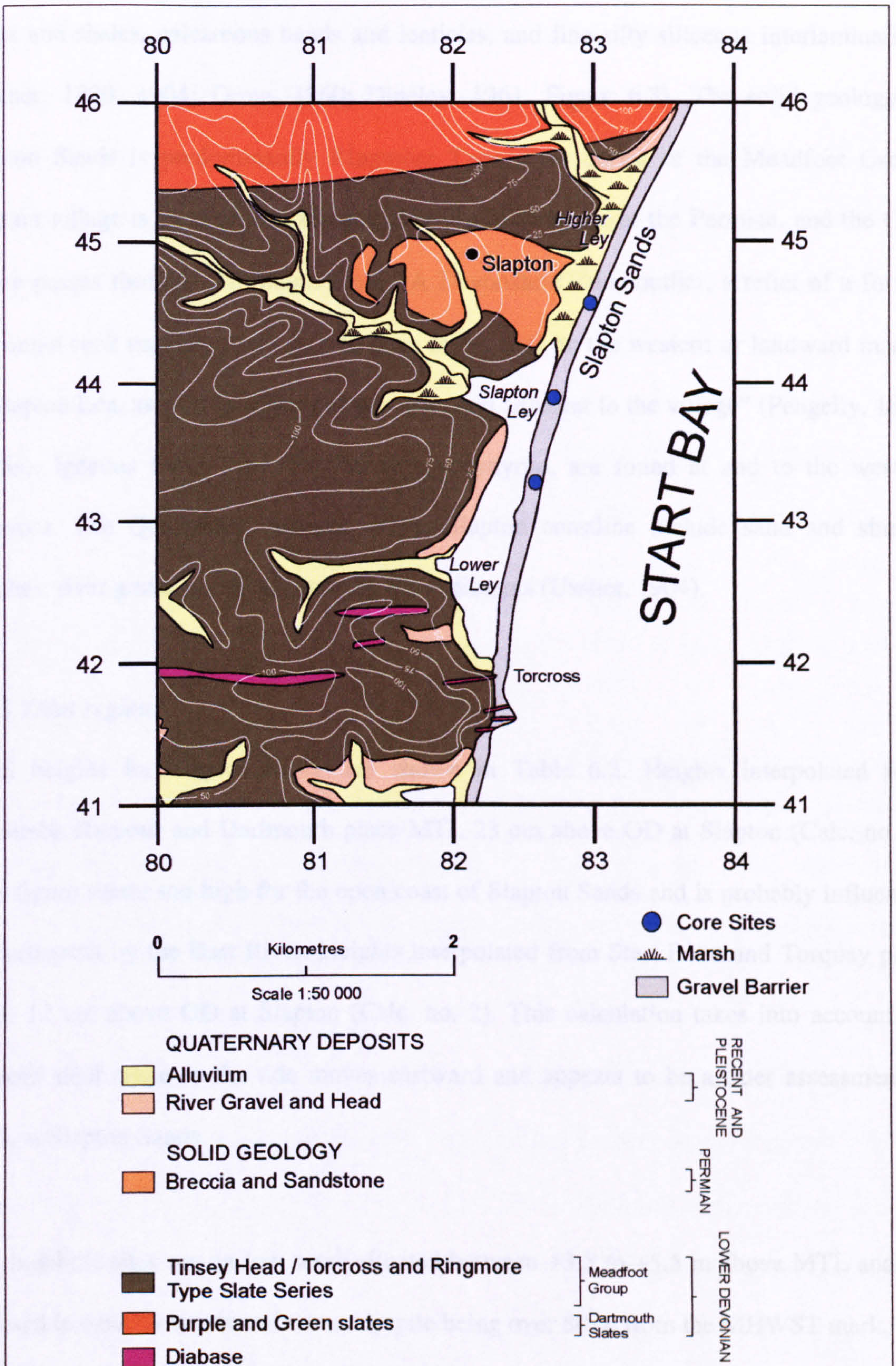


Figure 6.3. The solid geology and Quaternary deposits of Slapton Sands and the surrounding country. Source: Geological Survey of Great Britain (England and Wales) Sheets 355 and 356, Kingsbridge and Start Point, 1:50 000 (1975). Heights in metres above MSL.

Slapton Sands lies in the northernmost region of the Start Point Complex and is formed of Lower Devonian rocks comprising mainly slates with fine-grained grit beds, silty grey slates and shales, calcareous bands and lenticles, and fine silty siliceous interlaminations (Ussher, 1890, 1904; Orme, 1960; Dineley, 1961, Figure 6.3). The solid geology of Slapton Sands is predominantly Ringmore Type Slate Series of the Meadfoot Group. Slapton village is sited on Breccia and New Red Sandstone of the Permian, and the Gara valley passes through Dartmouth Slates. A Devonshire Trias outlier, a relict of a former Devonian rock surface, "may well be seen in the cliff on the western or landward margin of Slapton Lea, as well as at various places in and adjacent to the village" (Pengelly, 1864, 1866a). Igneous rocks, including Diabase porphyrite, are found at and to the west of Torcross. The Quaternary deposits of the Slapton coastline include sand and shingle beaches, river gravel and head, and alluvial sediments (Ussher, 1904).

6.1.3 Tidal regime

Tidal heights for Slapton Sands are shown in Table 6.2. Heights interpolated from Salcombe Harbour and Dartmouth place MTL 23 cm above OD at Slapton (Calc. no. 1). This figure seems too high for the open coast of Slapton Sands and is probably influenced at Dartmouth by the Dart River. Heights interpolated from Start Point and Torquay place MTL 12 cm above OD at Slapton (Calc. no. 2). This calculation takes into account the reduced tidal range as the tide moves eastward and appears to be a truer assessment of MTL at Slapton Sands.

The borehole sites are on low scrub situated between +3.5 to +5.5 m above MTL and are exposed to easterly winds and waves despite being over 80 m from the MHWST mark. The effects of wave and tidal conditions on beach and sandbank stability and morphology in Start Bay is well documented (Robinson, 1961; Acton and Dyer, 1975; Gleason *et al.*, 1975; Holmes, 1975; Huntley and Bowen, 1975; McManus, 1975). During 1995, the crest of Slapton Sands came within 5 m of the A379 road as storm waves reduced beach level by 2 m (Slapton Ley National Nature Reserve, 2002). After 5 years of natural replenishment strong easterly winds in December 2000 deposited about 50 tons of shingle onto the road, reducing beach level by 2 m again, and destroying about 600 m^2 of crest vegetation. The road was breached on the 11th January 2001 when easterly winds coincided with heavy rain and high spring tides causing seawater to enter the Ley. Approximately 3500 tons of rock armour was deposited on the beach to protect over 250 m of the A379 road (SX 831 447), or "line" (Mercer, 1966), prior to repairs.

Tidal Station / Site	Calc. no.	Latitude N	Longitude W	Distance (m)	MHWST (m)	MLWST (m)	Tidal Range (m)	Chart Datum (m)	MSL (m)	MTL (m)
Salcombe Harbour	1	50° 13'	3° 47'	10 000	2.25	-2.35	4.6	-3.05	3.14	0.09
Dartmouth (Standard Port)	1	50° 21'	3° 34'	5625	2.28	-2.02	4.3	-2.62	2.93	0.31
Slapton Sands, Start Bay	1	50° 16'.57	3° 38'.48	-	2.27	-2.14	4.41	-2.78	3.01	0.23
Start Point, Start Bay	2	50° 13'	3° 39'	6850	2.35	-2.05	4.4	-3.05*	3.20	0.15
Torquay (Standard Port)	2	50° 28'	3° 31'	17625	2.10	-2.10	4.2	-2.80	2.84	0.04
Slapton Sands, Start Bay	2	50° 16'.57	3° 38'.48	-	2.28	-2.06	4.34	-2.98	3.10	0.12

Table 6.2. Tidal heights for Slapton Sands interpolated from the nearest tidal stations. All heights are relative to the UK Ordnance Datum. Distance – approximate distance between Slapton Sands and Tidal Station using eastings for calculation no. 1 and northings for calculation no. 2 (coast turns due north at Start Point). *Start Point Chart Datum is not given in Admiralty Tide Tables and is assumed to be the same as Salcombe Harbour entrance. Source: Hydrographic Office (2002).

The Shoreline Management Plan (Lyme Bay and South Devon Coastline Group, 1998) discussed "holding or retreating the line" and English Nature assessed further possible threat to the site (W. S. Atkins Consultants Ltd, 2001). Further overtopping will threaten the line but landward migration of the barrier will not occur whilst the road remains in place, thus preserving the Reserve and its subsurface marine sediments. The Torcross outlet from the Ley allows the passage of storm water to Start Bay. Flooding of the wetlands is normally caused by rising groundwater levels after heavy rainfall.

6.2 Lithostratigraphy

Three cores were extracted approximately 80 to 130 m west of the MHWST mark (Figure 6.2). Detailed results from cores SS-99-1, SS-99-2 and SS-01-3 are shown in Figures 6.4 to 6.6 and are correlated in Figure 6.7. Further stratigraphical reconstruction using geophysical survey is illustrated in Figures 6.8 to 6.11.

6.2.1 Core SS-99-1

Core SS-99-1 (Figure 6.4) is located at SX 82698 43880 approximately 45 m west-north-west of the A379 road opposite Ireland Bay (Figure 6.2) and is 19.70 m long. Nine sections suffered core loss and one section included bagged sediment. The base comprises 20 cm of lithified reddish-brown slate overlain by fractured slates in a reddish-brown silty-clay (19.70 to 19.30 m). Wood and peat in silty-clay (19.30 to 19.16 m) is covered by reddish-pink-brown laminated mud (19.16 to 18.00 m) and silty-sandy-clay and shells (18.79 and 18.70 m). The overlying lithofacies consists of dark grey silty-sandy-clay, shell fragments, wood and burrows (18.00 to 9.10 m). The uppermost lithofacies is predominantly reddish-brown gravel, wood, roots and fibrous peat remains (9.10 m to ground level). This is interrupted by dark grey silty-sandy-clay with shells (8.22 and 8.17 m), and very dark greyish-brown sandy-silty clay with herbaceous plant remains (7.91 to 7.70 m).

6.2.2 Core SS-99-2

Core SS-99-2 (Figure 6.5) is located at SX 82576 43318 about 600 m south-south-west of core SS-99-1 and 15 m west-north-west of the A379 road near the shore of the Lower Ley (Figure 6.2) and is 16.50 m long. The uppermost 7.5 metres of gravel were not retrieved (see section 2.2.1) and three sections suffered core loss. The base (16.50 to 16.29 m) comprises fractured reddish-brown slates in sandy-silty clay with plant remains and the overlying organic rich clayey-silty-gravel contains wood fragments (16.29 and 16.23 m). This is overlain by a mixture of dark grey sandy-silty clay with shells (16.23 to 16.08 and

15.85 to 15.68 m) and light grey and reddish-brown slates and shales in a sandy-silty-clay with shells (16.08 to 15.85 and 15.68 to 15.48 m). The overlying sequence comprises dark grey sandy-silty clay, shells and organic remains (15.48 to 7.77 m), interrupted by dark greyish-brown laminated mud (13.80 to 13.65 m) and coarse sand and gravel (12.42 to 10.75 m) facies. Dark grey organic sandy-silty clay (7.77 to 7.65 m) is overlain by the uppermost lithofacies of reddish-brown gravel with roots, wood and fibrous peat remains (7.65 m to ground level).

6.2.3 Core SS-01-3

Core SS-01-3 (Figure 6.6) is located at SX 82987 44563 approximately 750 m north-north-east of core SS-99-1, 30 m west-north-west of the A379 road near the Higher Ley reedswamp perimeter (Figure 6.2) and is 15.44 m long (Massey, 2001). Four sections suffered core loss and the section between 7.3 and 6 m was not recovered. The basal unit of lithified and fractured light greyish-brown and weak red slate (15.44 to 13.90 m) is overlain by red sandy-silty clay with plant detritus (13.90 to 13.00 m) and fractured red slate in silty-sandy clay (13.00 to 12.56 m). Dark grey sandy-silty clay with shells lies at 13.05 to 13.00 metres. The overlying lithofacies consists predominantly of dark grey gravelly-sandy-silty clay with shells and organic detritus (12.56 to 9.60 m) interrupted by silty-clayey-sandy gravel containing shell fragments (10.17 to 10.00 m). This is overlain by a similar unit between 9.60 and 9.42 m, but devoid of shells and weak red in colour. The uppermost sequence consists of greyish and pale brown coarse sand and gravel with roots, wood and humified peat remains (9.42 m to ground level).

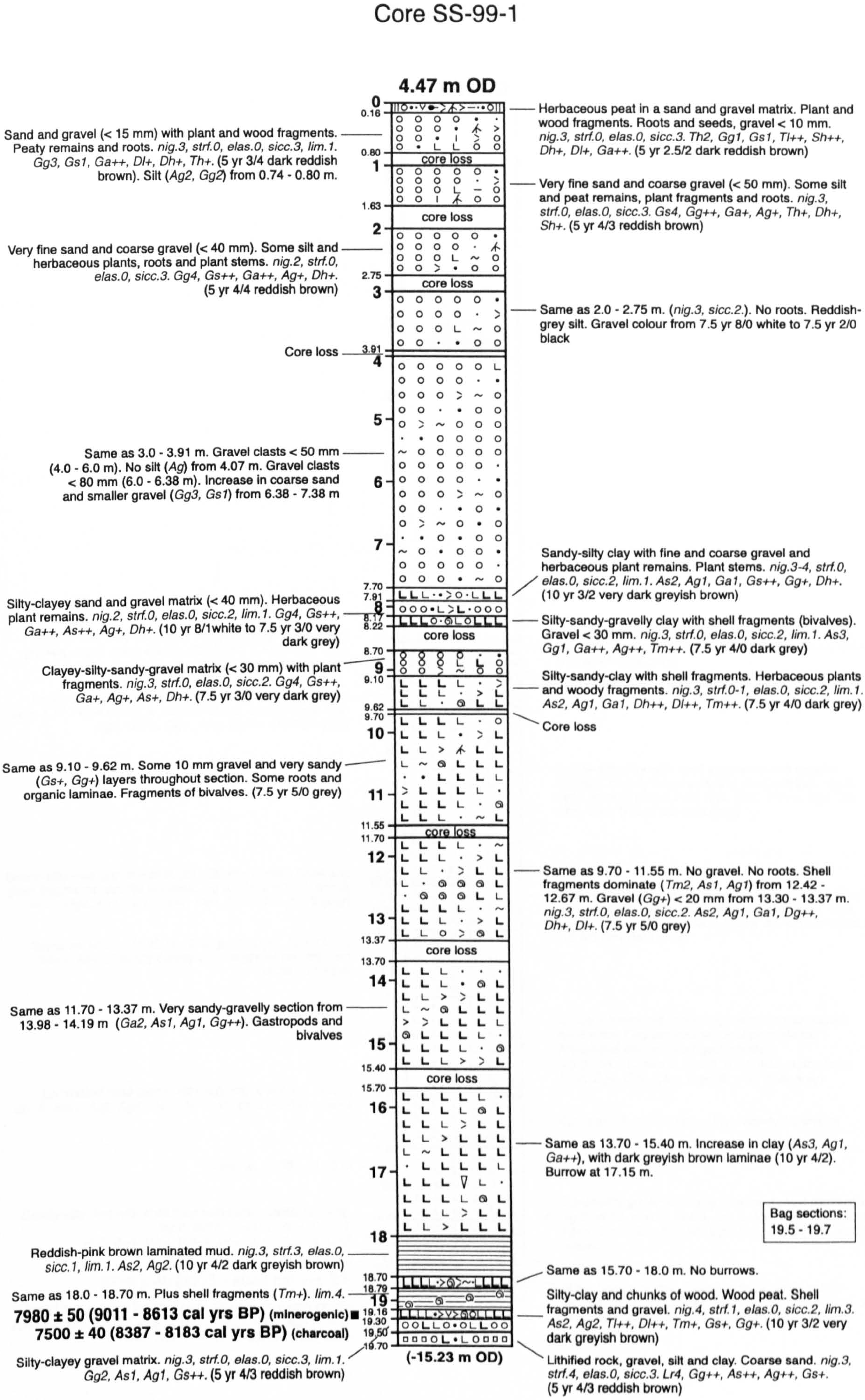


Figure 6.4. Lithostratigraphy of core SS-99-1.

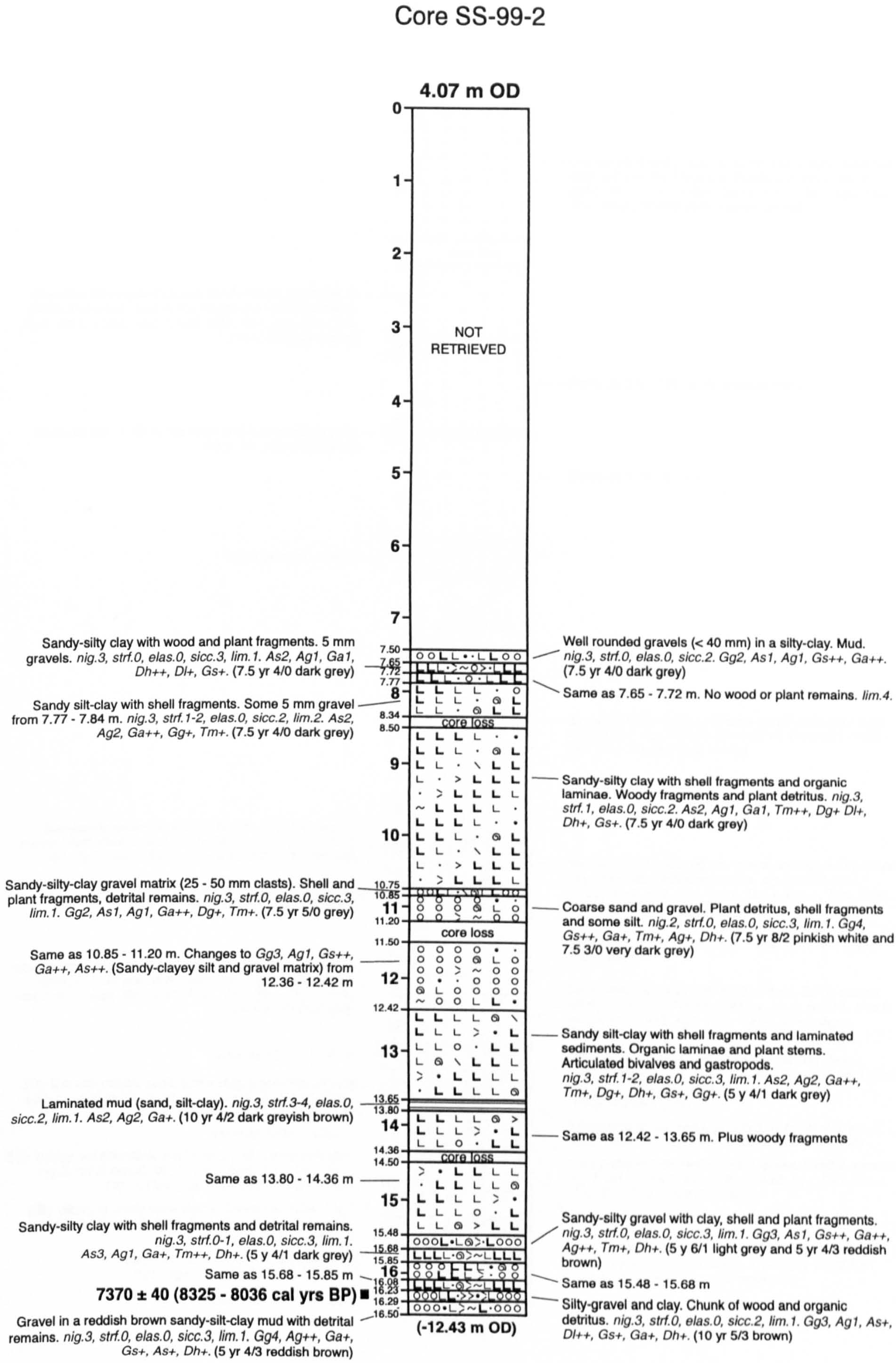


Figure 6.5. Lithostratigraphy of core SS-99-2.

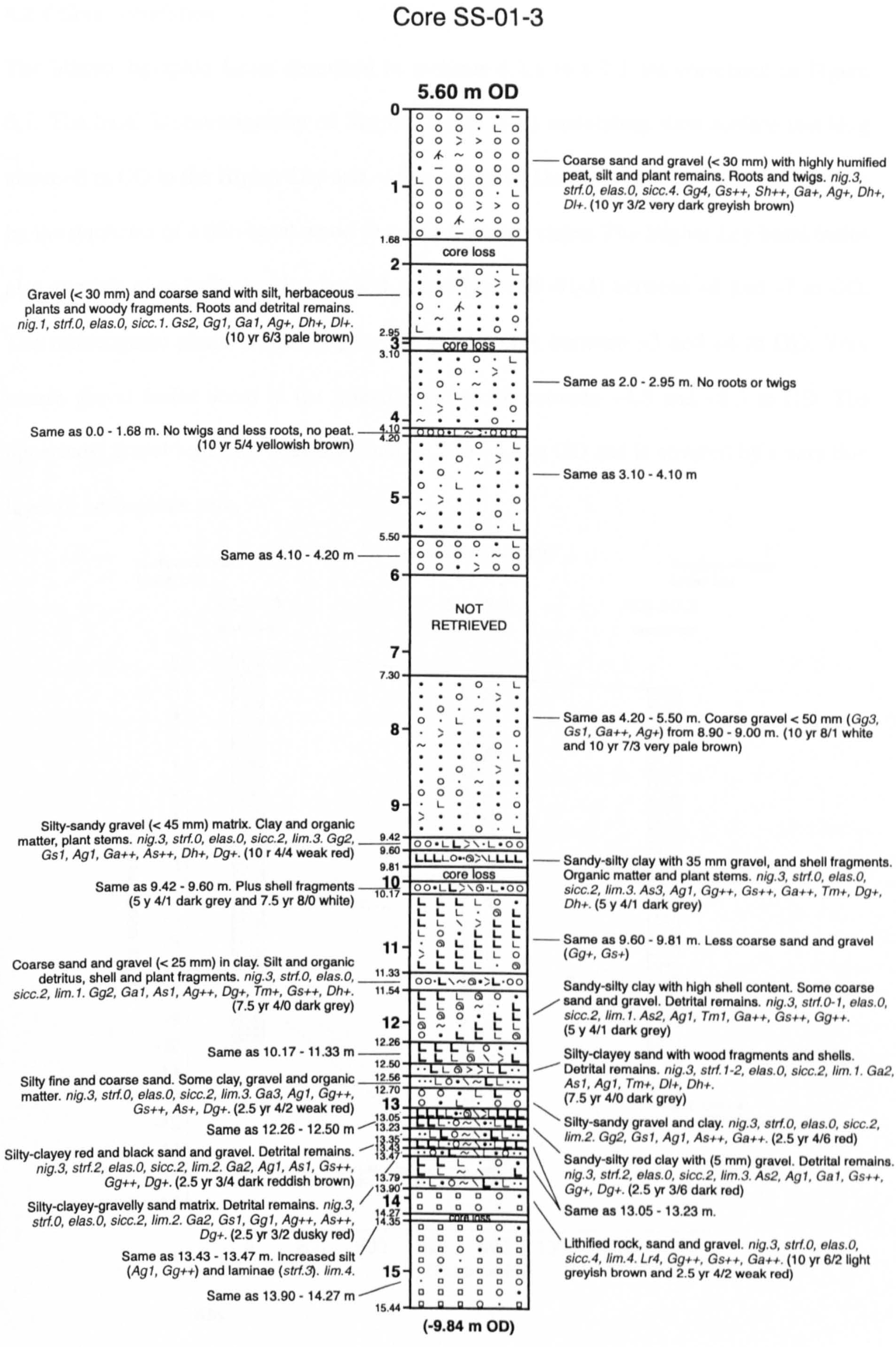


Figure 6.6. Lithostratigraphy of core SS-01-3.

6.2.4 Core correlation

The lithostratigraphic facies described in sections 6.2.1 to 6.2.3 are correlated in Figure 6.7. The basal lithostratigraphy of Slapton Sands is an undulating slate surface reaching about –8 m OD in the Higher Ley and –15 to –12 m OD in the Lower Ley. This is overlain by the remnants of a thin basal wood peat and fractured slates. The Higher Ley basal facies also consists of red silty-sandy clay and gravel (core SS-01-3) between –8 and –7 m OD. The minerogenic phase is mainly grey sandy-silty-clay between –3 and –4 m OD. Very coarse gravel facies occur in the minerogenic phase between –4.5 and –8.5 m OD. The uppermost gravel sequence begins around –3.2 to –3.8 m OD and is covered by a very thin layer of herbaceous peat.

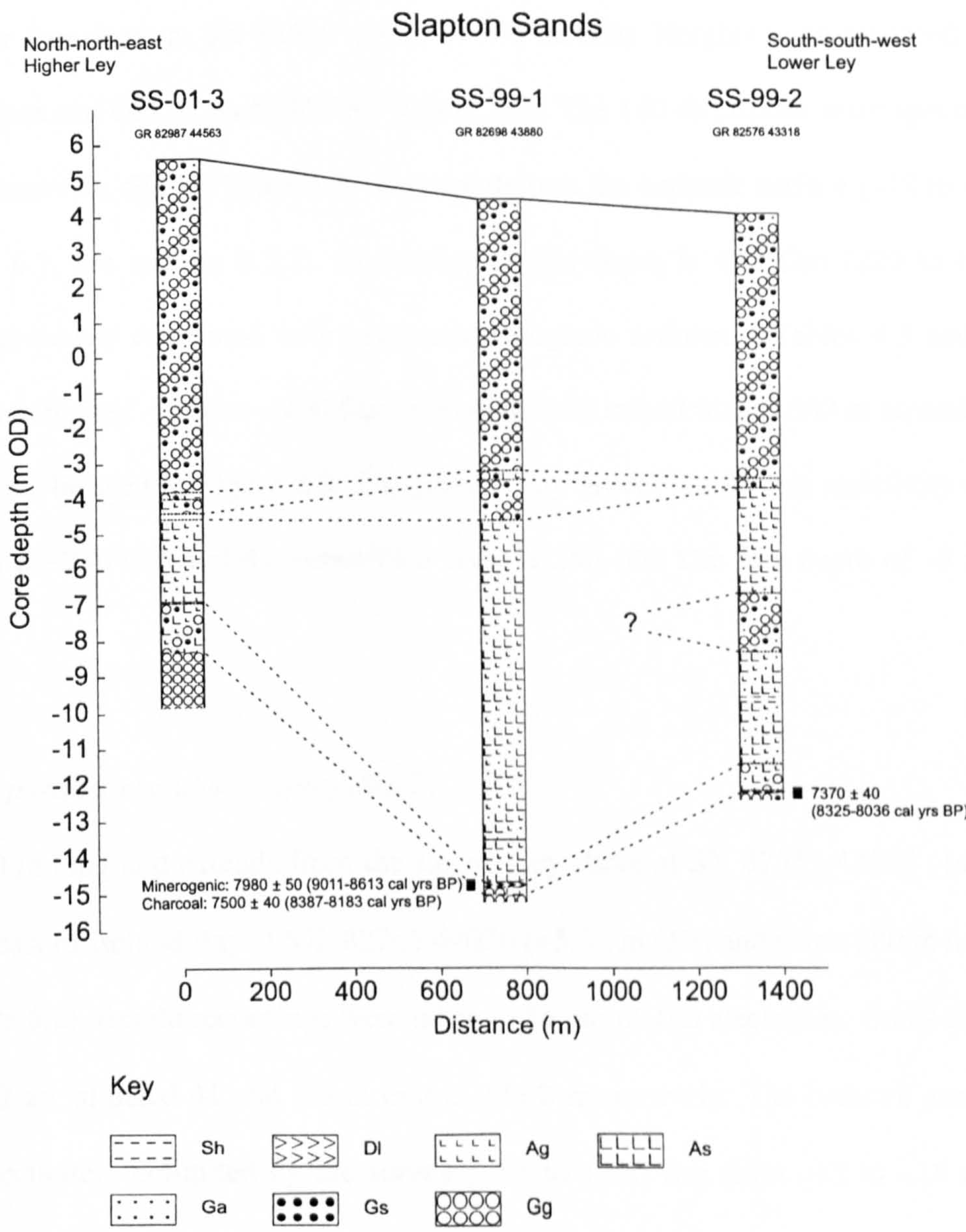


Figure 6.7. Slapton Sands core correlation.

6.3 Electrical resistivity survey

Two shore-parallel (Figures 6.8 to 6.11) electrical resistivity surveys were carried out along the Slapton barrier. Cores SS-99-1 and 2 are used as ground-truths for the survey. The geographical locations of the survey lines are shown in Figure 6.2. Both survey lines were rolled out along the footpath about 15 (near core SS-99-2) to 50 m (near core SS-99-1) east of Slapton Ley to reduce interference from the lake. A summary of the electrical resistivity of sediments at Slapton Sands is given in Table 6.3.

6.3.1 Shore-parallel resistivity survey (Sla 1)

Sla 1 is 700 m long and extends from Torcross gatepost at SX 82365 42400 (+5.43 m OD) to the Lower Ley shore at SX 82510 43075 (+4.71 m OD). Heights were recorded at the start, mid-point and end of each (100 m) survey line. The 140 electrodes were spaced at 5 m standard intervals. Sla 1 (Figure 6.8) does not delimit the bedrock surface (–12 to –15 m OD, Figure 6.7, see section 8.5.3). Resistivity at this depth is <32 Ωm (220 to 600 m distance), previously correlated with peat or minerogenic sediment (Tables 4.3 and 5.2). Two large 'anomalies' of 50 to +200 Ωm from 40 to 220 m and 600 to 660 m around –8 to –20 m OD can be seen in Figure 6.8. The present day gravel barrier has resistivity values in the range 63–250 Ωm and the subsurface layer is 250–320 Ωm to a depth of –3 m OD (Figure 6.7).

6.3.2 Shore-parallel resistivity survey (Sla 2)

Sla 2 is 800 m long and extends from the Lower Ley shore at SX 82555 43250 (+4.11 m OD) to the east of Ireland Bay at SX 82765 44020 (+5.31 m OD) and starts 180 m north of Sla 1 (Figure 6.2). Height recordings were taken at 86 out of 160 electrodes. Cores SS-99-1 and SS-99-2 are situated 41 and 5.5 m east of Sla 2 respectively. The bedrock surface is again not accurately delimited by the survey (<32 to 1500 Ωm from –12 to –15 m OD, Figures 6.9, 6.10, 6.11 and section 8.5.3) and another anomaly (220 to 1500 Ωm) occurs at

160 m around –30 m OD. Between 300 and 540 m at –7 m OD, resistivity values >68 Ωm are recorded coinciding with a gravel facies in core SS-99-2. This is not shown in the enlarged view (Figure 6.11). The present-day gravel barrier is again delimited by resistivity values >68 Ωm between –5 and –3 m OD to ground level (Figures 6.10 and 6.11).

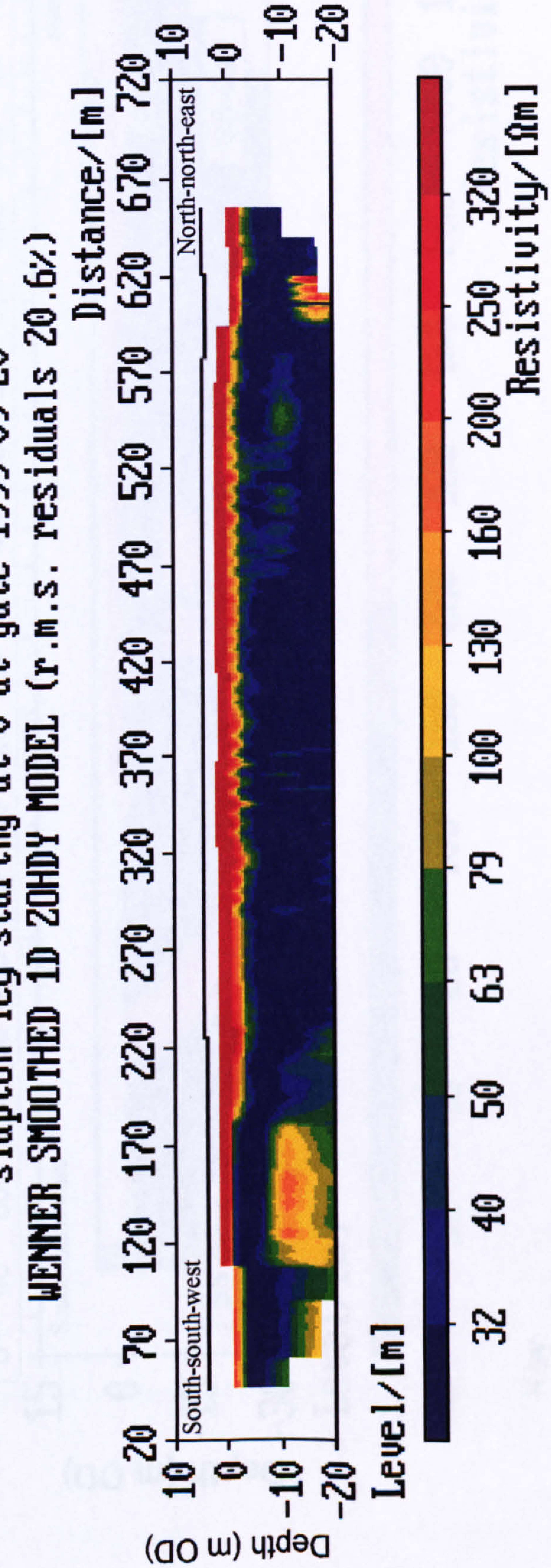


Figure 6.8. Shore-parallel resistivity profile (Sla 1).

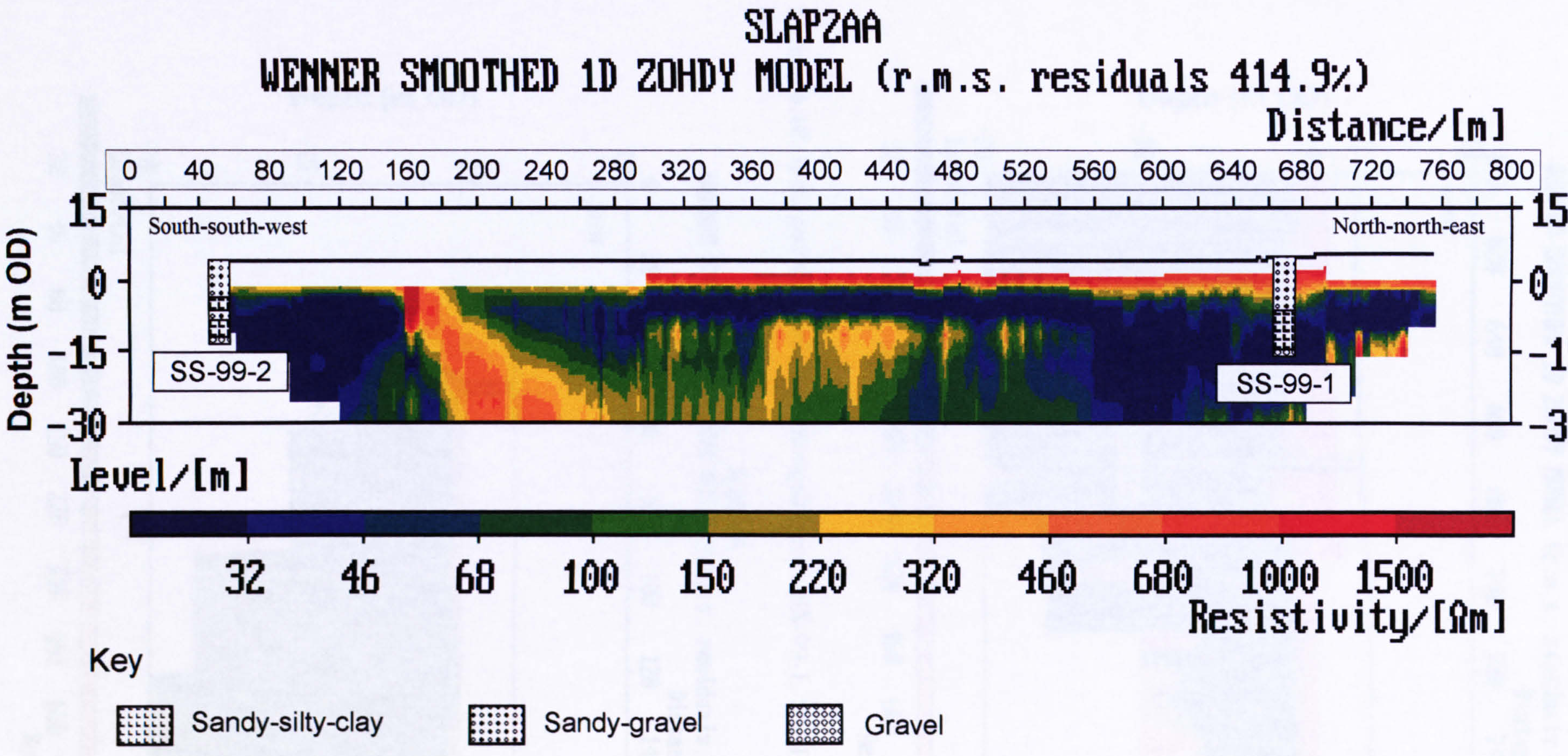


Figure 6.9. Shore-parallel resistivity profile (Sla 2).

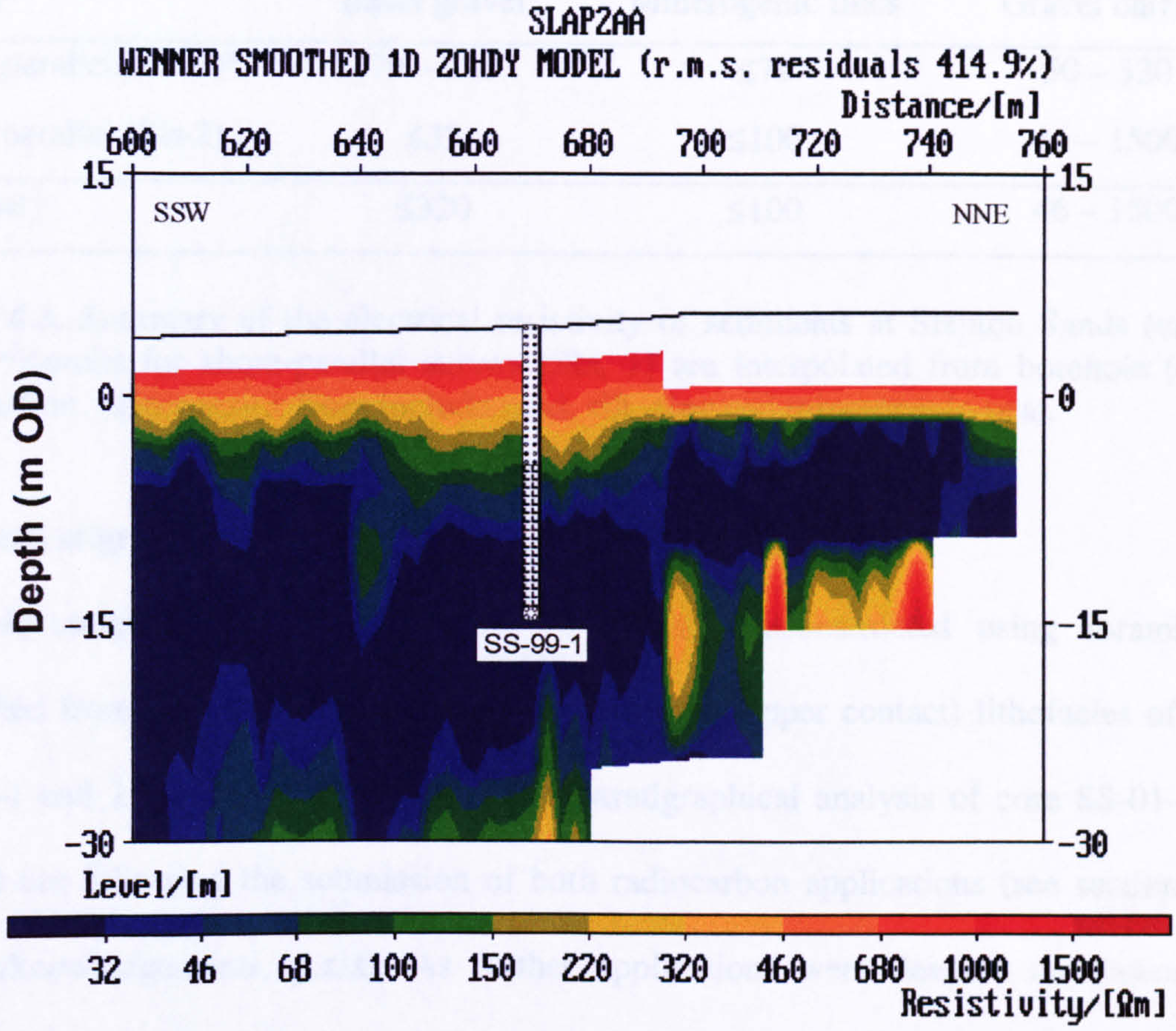


Figure 6.10. Enlarged view of Sla 2 alongside core SS-99-1. See Figure 6.9 for borehole key.

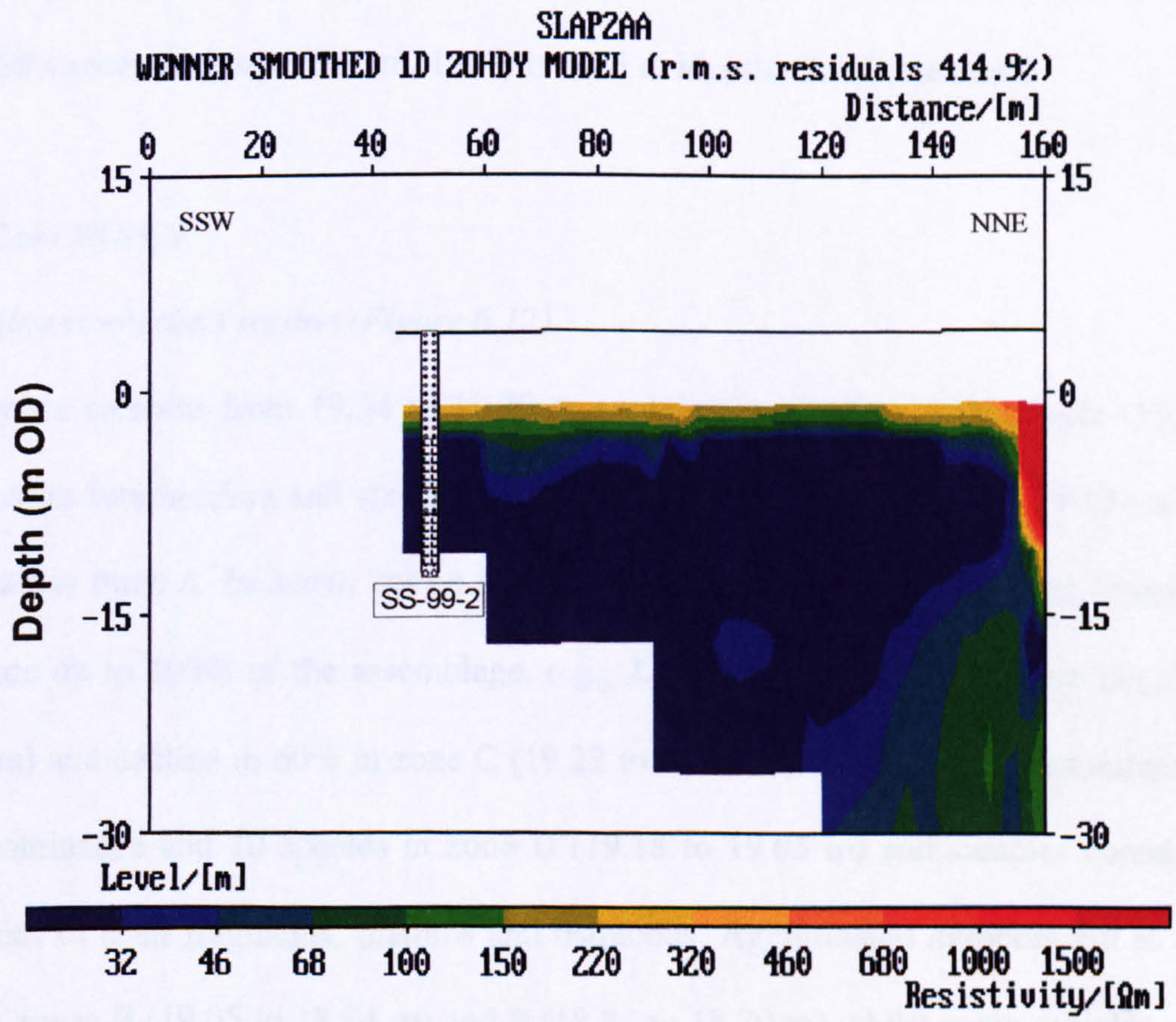


Figure 6.11. Enlarged view of Sla 2 alongside core SS-99-2. See Figure 6.9 for borehole key.

Survey	Basal gravel	Minerogenic fines	Gravel barrier
Shore-parallel (Sla 1)*	79 – 320	≤79	50 – 320
Shore-parallel (Sla 2)	≤32	≤100	46 – 1500
Summary	≤320	≤100	46 – 1500

Table 6.3. Summary of the electrical resistivity of sediments at Slapton Sands (units in Ωm). *Results for shore-parallel survey (Sla, 1) are interpolated from borehole (depth) information. Minerogenic fines include sand, silt, clay and minerogenic-peat.

6.4 Biostratigraphy

The biostratigraphy of Slapton Sands is partially reconstructed using foraminifera identified from the basal (lower contact) and in-core (upper contact) lithofacies of cores SS-99-1 and 2 (Figures 6.12 to 6.15). Lithostratigraphical analysis of core SS-01-3 was carried out following the submission of both radiocarbon applications (see section 2.3.4 and acknowledgements, p.xix). As further applications were deemed unnecessary no biostratigraphical analysis was carried out on core SS-01-3. The indicative meaning of each sample is calculated based on the transfer function described in chapter 3 and the CONISS cluster analyses (Grimm, 1987) is used to identify fossil biozones.

6.4.1 Core SS-99-1

Basal (lower contact) section (Figure 6.12)

Twenty-six samples from 19.34 to 18.70 m contain foraminifera, one sample (19.31 m) contains no foraminifera and six distinct biozones are identified. Zone A (19.34 to 19.31 m) contains three *A. beccarii*, but no salt- or freshwater shells. Agglutinated foraminifera dominate up to 100% of the assemblage, e.g., *J. macrescens* (80%), in zone B (19.31 to 19.22 m) and decline to 60% in zone C (19.22 to 19.18 m). Calcareous foraminifera reach 65% dominance and 10 species in zone D (19.18 to 19.05 m) and samples contain high quantities of shell fragments, diatoms and ostracods. Agglutinated numbers fall to almost zero in zones E (19.05 to 18.84 m) and F (18.84 to 18.70 m) whilst some samples contain fifteen species of calcareous foraminifera.

Figure 6.12. Foraminifera stratigraphy (%) of core SS-99-1: basal (lower contact) section.

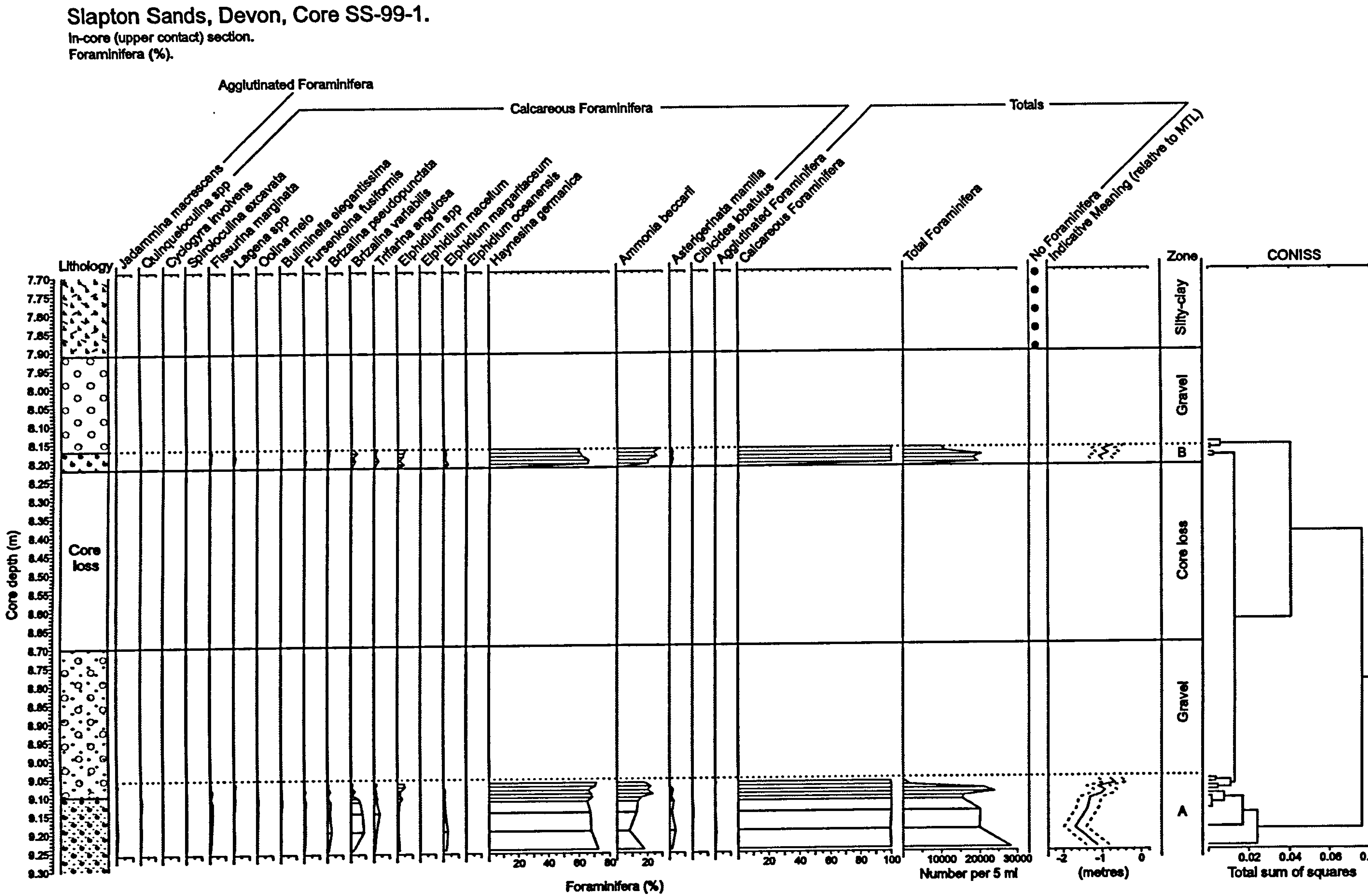


Figure 6.13. Foraminifera stratigraphy (%) of core SS-99-1: in-core (upper contact) section.

In-core (upper contact) section (Figure 6.13)

Samples were selected above and below the coarse gravel facies (Figure 6.4). Fourteen samples from 9.26 to 8.17 m contain foraminifera. Five samples from 7.91 to 7.70 m contain no foraminifera and two distinct zones are identified. Zone A (9.25 to 9.06 m) contains high numbers (>28000 per 5 ml) of predominantly calcareous foraminifera (*J. macrescens* is present in small numbers). Numbers decrease to 300 specimens, and shells, diatoms and ostracods are also present in the fine gravel facies at 9.07 m. A similar trend is found in zone B (8.22 to 8.17 m) where *H. germanica* again dominates (~70%). Samples above the gravel facies (7.91 m) contain significant quantities of *Chara* oospores (216 specimens per 5 ml at 7.75 m) and fauna, e.g., *Acari, Chironomidae, Coleoptera,* and flora, e.g., moss leaves, *Phragmites* remains and a hazel nut (*Corylus avellana*) are also present.

6.4.2 Core SS-99-2

Basal (lower contact) section (Figure 6.14)

Twenty-two samples from 16.25 to 15.51 m contain foraminifera. Seven samples between 16.33 and 16.26 m contain no foraminifera and four distinct zones are identified. Basal samples (16.33 to 16.26 m) contain seeds, bryophytes, *Phragmites* remains and large fragments of alder in peat around 16.30 metres. Zone A (16.26 to 16.23 m) is dominated initially (100%) by low numbers (10 to 80 per 5 ml) of agglutinated foraminifera reducing to 35% within a few samples. The total (around 4000 specimens) and diversity (up to 12 species) of calcareous specimens increase significantly in zone B (16.23 to 15.97 m) but numbers fall sharply in the gravel facies (zone C, 15.97 to 15.84 m). For example, 12 damaged calcareous tests were found at 15.90 metres. Numbers increase to >3000 per 5 ml in zone D (15.84 to 15.50 m) and the most dominant species are *A. beccarii* and *E. margaritaceum*. Zones B, C and D contain significant numbers of ostracods, shells (including articulated gastropods and bivalves) and diatoms.

Figure 6.14. Foraminifera stratigraphy (%) of core SS-99-2: basal (lower contact) section.

Figure 6.15. Foraminifera stratigraphy (%) of core SS-99-2: in-core (upper contact) section.

In-core (upper contact) section (Figure 6.15)

Samples were selected above and below the coarse gravel facies (Figure 6.5). Thirty-one samples from 12.50 to 7.71 m contain foraminifera. Five samples from 7.71 to 7.50 m contain no foraminifera and three distinct zones are identified. Zone A (12.50 to 12.38 m) contains up to 4000 specimens of around 19 calcareous species dominated by *H. germanica* and *A. beccarii*. Totals decrease significantly (e.g., 60 specimens at 12.38 m) in the fine gravel and the zone contains many shells, diatoms and ostracods. Zone B (11.20 to 10.65 m) is similar to zone A but the gravel (11.19 to 10.87 m) was too coarse to be analysed. Numbers in zone C (7.86 to 7.71 m) initially reach over 10,000 per 5 ml, falling to 16 at 7.71 metres. Samples above 7.75 m also contain *Chara* oospores, moss leaves, *Acari*, *Coleoptera*, and *Chironomidae*. Samples between 7.71 and 7.50 m contain aquatic flora and fauna but no foraminifera or other calcareous fauna.

6.5 Chronology

Three age measurements by AMS ^{14}C of the basal sediment of cores SS-99-1 and 2 (Table 6.4) provide SLIPs (Figure 6.16). Selecting suitable material for AMS ^{14}C from the base of core SS-99-1 proved problematic. A 1.5 cm slice of sandy-silty-clay extracted from –19.24 m (–14.89 m MTL) lacked plant macro remains, but contained charcoal fragments that may be old carbon or a remnant of the original Holocene basal peat. Charcoal weighing 3.55 gms and minerogenic sediment weighing 16.55 gms, were therefore dated separately. Similarly, in core SS-99-2, a very thin salt-marsh facies exists at –16.26 m (–12.31 m MTL) containing low numbers of foraminifera (10 per 5 ml). Numbers increase to 2776 at a core depth of –16.23 m but the assemblage and indicative meaning indicate mudflat conditions (Figure 6.14) which is not the preferred facies for dating. A large (2 cm long) chunk of alder is present at –16.28 m (–12.33 m MTL). This has clearly been transported but the environment in which it was deposited is an inter-tidal one, i.e., the overlying salt marsh. The wood fragment, weighing 1.86 gms, was therefore selected for AMS ^{14}C dating.

Index number	18	19	20
Radiocarbon laboratory number	CAMS-72401	CAMS-72402	CAMS-75531
Core	SS-99-1	SS-99-1	SS-99-2
Material	Bulk sediment (minerogenic)	Bulk sediment (charcoal)	Wood (twig)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	37.05 ± 0.19	39.32 ± 0.18	39.98 ± 0.18
^{14}C age (years BP $\pm 1\sigma$)	7980 ± 50	7500 ± 40	7370 ± 40
Calibrated BP $\pm 1\sigma$ age ranges	(9001-8749)	(8367-8325)	(8187-8162)
	(8739-8719)	(8310-8296)	(8134-8133)
	(8704-8704)	(8260-8247)	(8117-8112)
	(8664-8664)	(8240-8211)	(8077-8061)
Calibrated BP $\pm 2\sigma$ age ranges	(9011-8633)	(8387-8183)	(8325-8295)
	(8615-8613)		(8294-8260)
			(8247-8239)
			(8214-8146)
			(8143-8107)
			(8089-8036)
Median calibrated age (cal yrs BP)	8868	8341	8175
Carbon content (% by weight)	1.2	37	54
$\delta^{13}C \pm 0.1$ (‰)	-29.1	-26.7	-29.7
MTL sample height (m)	-14.89	-14.89	-12.33
Miliammina fusca	16	16	0
Polysaccammina ipohalina	32	32	0
Trochammina inflata	552	552	4
Trochammina ochracea	72	72	0
Jadammina macrescens	1792	1792	6
Indicative meaning (m MTL)	2.43 ± 0.29	2.43 ± 0.29	2.51 ± 0.29
SLIP (m MTL)	-17.32 ± 0.29	-17.32 ± 0.29	-14.84 ± 0.29

Table 6.4. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 6.16 and 6.17. See Table 6.2 for MTL datum.

The results for core SS-99-1 are unusual in that charcoal is expected to produce an older age than minerogenic sediment (Table 6.4). However, the minerogenic sample could contain older carbon washed in from 'offshore' barriers (see sections 5.7 and 5.8). The charcoal may also be remnants of charred plant material deposited on the former marsh surface, or it could be re-worked and therefore older than the in-situ sediment, but is unlikely to be too young. This argument suggests that the minerogenic age is likely to be too old and that the charcoal date could be an accurate age for the palaeo-marsh surface. As it is unlikely to be younger than the early-Holocene salt marsh it therefore provides a maximum age estimate for this facies. On this basis SLIP 18 is considered a borderline inclusion.

The onset of marine conditions at Slapton Sands occurred *ca*. 9000 to 8000 cal years BP (Figures 6.12, 6.14 and 6.16). SLIP 18 shows that high marsh conditions (around +2.5 m MTL) were well established at Slapton Sands *ca*. 9000 – 8600 cal years BP (Figure 6.12). SLIP 20 shows that high marsh / supratidal conditions (around +2.5 to +3 m MTL) were established *ca*. 8300 – 8000 cal years BP (Figure 6.14). Using SLIP 19 (charcoal) mean sea level rose ~1.9 – 3.1 metres in 350 years, a rate of 5 to 9 m/ka. Lambeck's (pers. comm. 1997) geophysical model data is the closest fit of the SLIPs (Figure 6.17).

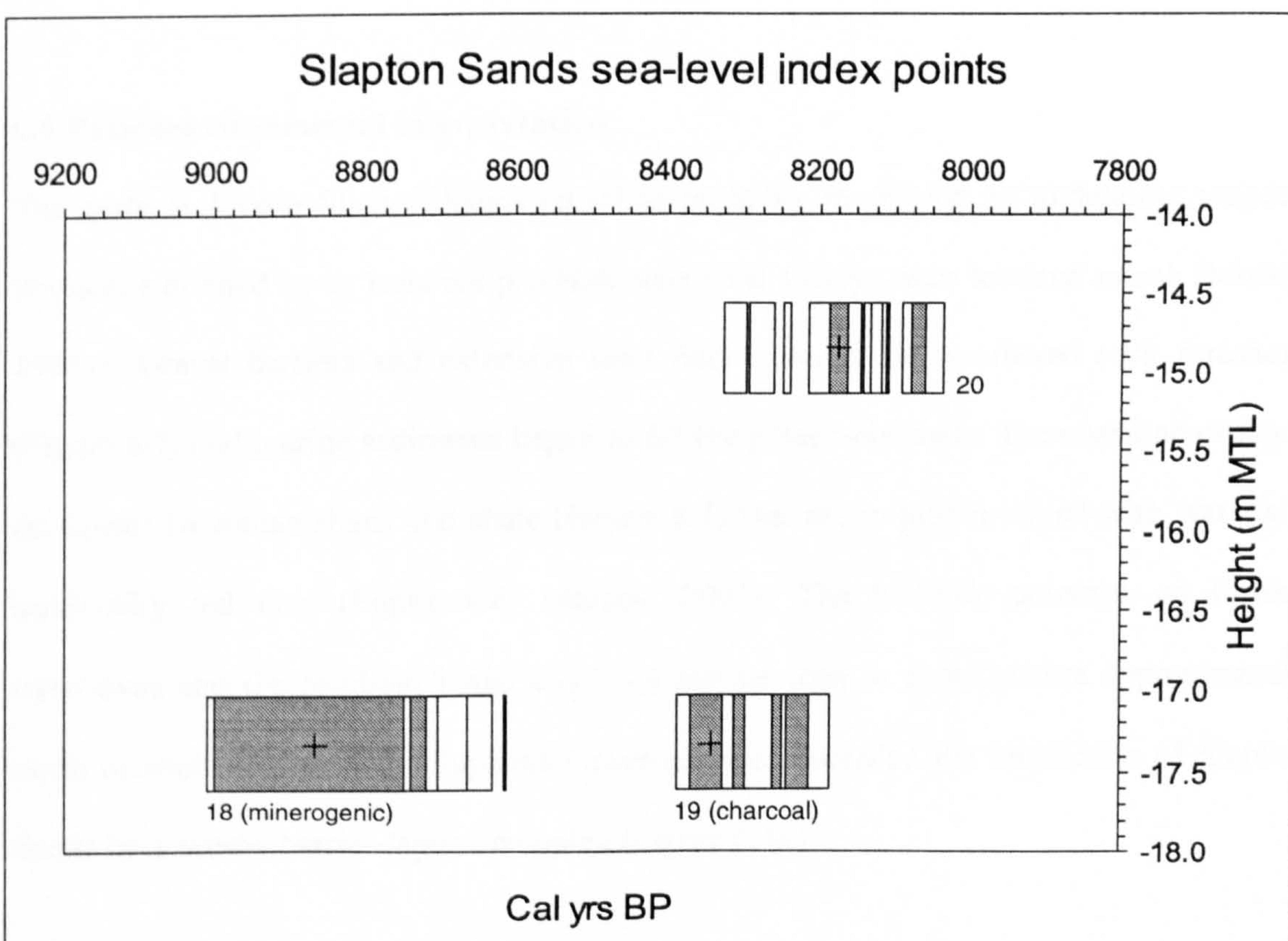


Figure 6.16. Slapton Sands SLIPs. Index numbers correspond with Table 6.4.

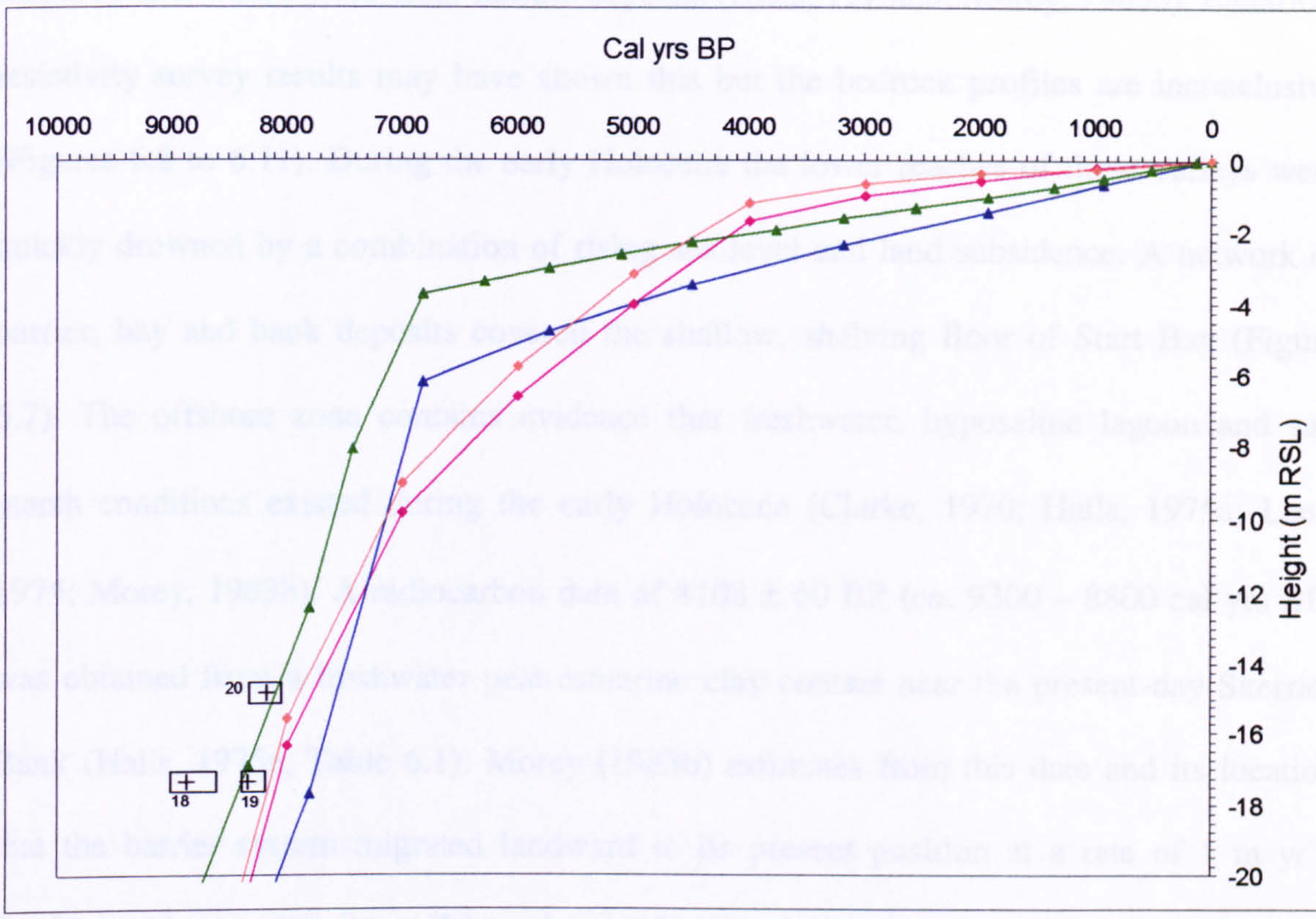


Figure 6.17. Slapton Sands SLIPs and geophysical modelling results. Index numbers correspond with Table 6.4.

6.6 Palaeoenvironmental interpretation

The early Holocene Slapton Sands coastline probably resembled an undulating wooded landscape drained by at least six pre-Holocene river valleys onto lowland marsh (Morey, 1983a). Gravel barriers and extensive sand flats covered the weathered rock foreshore (Figure 6.7) and marine sediments began to fill the palaeo-channels. The coastline evolved on Lower Devonian slates and shale (Figure 6.3) that are in places mixed with waterlain sandy-silty red clay (Figure 6.6, Massey, 2001). The bedrock outcrops on former interfluves and the resultant highs and lows can be seen in cross-section approximately north to south (Figure 6.7). Supratidal environments preceded the inundation of Slapton Sands by a nearby barrier-lagoon complex (Figure 6.14).

The post-glacial marine transgression of Slapton Sands occurred *ca.* 9000 – 8000 cal years BP (Figure 6.16), but deeper (28 m) palaeo-channels are thought to exist along the current shoreline that would yield older marine deposits (Hails, 1975a,b; Morey, 1983a). Electrical resistivity survey results may have shown this but the bedrock profiles are inconclusive (Figures 6.8 to 6.11). During the early Holocene the lower reaches of these valleys were quickly drowned by a combination of rising sea level and land subsidence. A network of barrier, bay and bank deposits covered the shallow, shelving floor of Start Bay (Figure 6.7). The offshore zone contains evidence that freshwater, hyposaline lagoon and salt marsh conditions existed during the early Holocene (Clarke, 1970; Hails, 1975a; Lees, 1975; Morey, 1983b). A radiocarbon date of 8108 ± 60 BP (*ca.* 9300 – 8800 cal yrs BP) was obtained from a freshwater peat-estuarine clay contact near the present-day Skerries Bank (Hails, 1975a, Table 6.1). Morey (1983b) estimates from this date and its location that the barrier system migrated landward to its present position at a rate of 1 m yr^{-1}. Further evidence of former-offshore lagoonal systems is difficult to find due to the likely cycle of proto-barrier formation, overwash and destruction that would have transgressed the floor of Start Bay during the early Holocene.

Salt marsh conditions existed at the core sites *ca.* 9000 to 8000 cal years BP (Figures 6.12 and 6.14). Shallow tidal mudflat or lagoonal conditions then replaced the brackish environment as local sea level rose at 5 to 9 m/ka. Around 11 – 12 metres of marine sediment was then deposited during the early to mid-Holocene punctuated by gravel facies indicating subtidal barrier movement at the site (Figure 6.7). Shallow lagoonal environments preceded and followed the establishment of gravel barriers at the core sites, and these barriers were not stable for long enough to allow brackish or freshwater back-barrier conditions to form (Figures 6.13 and 6.15).

Peat resting on grey clay with brackish fauna from Slapton Ley has been dated *ca.* 3200 – 2850 cal years BP (Morey, 1976, 1983b, Table 6.1) and this may be a reasonable approximation of when the lagoon was sealed by the gravel ridge. Dating of peat-clay contacts from Beesands suggest that such events occurred earlier, *ca.* 5600 – 4800 cal years BP (Morey, 1983b, Table 6.1). This event agrees with the approximate timing of barrier closure at North Sands, *ca.* 5000 – 4500 cal years BP (see section 5.8). A back-barrier wetland environment followed closure of the marine phase (around –3.5 m OD, Figure 6.7). This occurred very suddenly as salt-marsh biofacies are not well established in the record (Figures 6.13 and 6.15). A thin minerogenic-peat layer or freshwater *gyttja* is present in core SS-99-1 at –3.44 to –3.23 m OD containing plant macrophytes and fauna indicative of standing water in a marsh setting (e.g., 216 *Chara* oospores, see section 6.4.1 and Appendix 2a). This is situated between a subtidal or washover gravel facies at –3.70 m OD and the base of the present-day supratidal barrier at –3.23 m OD (Figure 6.4). The uppermost contacts of all three cores appear to contain washover facies similar to those identified in Slapton Ley by Morey (1976). Following the cessation of tidal influence a back-barrier environment of reed marsh (*Phragmites communis*), water lily (*Nymphaea*) and willow carr (*Salix*) developed in the Lower Ley. Reedswamp with rafts of bracken (*Pteridium aquilinum*), bramble (*Rubus fruticosus*), Ivy (*Hedera helix*) and Broad Buckler Fern (*Dryopteris dilatata*) developed in the Higher Ley (Benson-Evans *et al.*, 1966; Mercer, 1966; Morey, 1976).

6.7 Summary

Prior to this study at least seven index points had been produced to reconstruct Holocene sea-level change along the Slapton Sands coastline (Table 6.1). Holocene SLIPs in this project date specifically the onset of post-glacial sea-level rise at Slapton Sands. The lithostratigraphy shows that a complex barrier-lagoon system migrated landward on top of Lower Devonian slate and fluvial silts and clay during the early-Holocene marine transgression. Foraminiferal results show that basal salt-marsh peat developed along the fringes of barriers *ca.* 9000 to 8000 cal years BP. Marsh peat was then covered by up to 12 metres of marine sediment during the early- to mid-Holocene. This was punctuated, in places, by subtidal gravel barrier deposits. The rate of local sea-level rise, *ca.* 8000 BP, is about 5 to 9 m/ka. Geophysical models, particularly Lambeck's (pers. comm. 1997), agree very closely with SLIPs 19 and 20. During the mid-Holocene, palaeo-channels filled with sand, silt, clay and gravel under the influence of rising seas until the barrier resembled its current form and closed off all tidal inlets *ca.* 3000 cal years BP (Morey, 1983a). The back-barrier reed marsh and low scrub system then developed along the now impounded freshwater lake. The present-day back-barrier environment is still subject to washover events produced primarily by strong easterly winds and waves.

Chapter 7

Blackpool Sands

7.1 Site description

Blackpool Sands is located at SX 8545 4785 and derives its name from a range of sources (Figure 7.1). Black or Blag in Devon probably implied holy or sacred and was frequently prefixed to landmarks or boundaries. Blackpool Lake once marked the Stoke Fleming and Blackawton (now Strete, Stanes, 1983) parish boundary, but no lake exists now. The site lies in the lower reaches of a stream still known today as Blackpool Lake (Blackpool Sands Visitor Site, 2002) and the land-use is predominantly mixed woodland, temporary grass, grazing and arable farmland. The immediate back-barrier is managed grassland bordered by low scrub. Research on the occupation of Blackpool is limited but Neolithic, Iron Age and Roman sites and other archaeological finds have been discovered as close as the Dart Valley (Houlder, 1963; Masson Phillips, 1966; Fox, 1969, 1973; Hunt, 1873; Palmer, 1977). There are many references to the site in coastal studies (e.g., Pengelly, 1869; Hunt, 1881; Worth, 1904, 1909; Morey, 1974, 1976, 1980, 1983a; Hails, 1975a,b; Lees, 1975; Heyworth and Kidson, 1982). However, only a single index point from the submerged forest on Blackpool Sands has emerged from this research (Morey, 1974). Clarke (1970) has reconstructed early Holocene environments in the offshore region of Teignmouth, north of Blackpool Sands, but his index point has large age error estimates from the salt-marsh sample (Table 7.1).

7.1.1 Geomorphology

Blackpool Sands is a gravel barrier about 700 m in length from the rock spur north of Jenny Cole's Cove to the cliffs below Stoke Fleming (Figure 7.1) and is a continuation of the Slapton Sands barrier (Hails, 1975a,b). The back-barrier system is a wooded wetland valley drained by Blackpool Lake to the southernmost point of Blackpool Sands (Figures

7.1 and 7.2). The narrow valley once contained a dammed lake behind the gravel barrier that gradually filled in with organic silts and mud. It is now constrained by a rocky coastal slope rising 30 to 50 m above MHWST. The site lies south of the A379 about 50 m inland from MHWST at +4 to +6 m OD and occupies an area of 2 ha. It is similar to Hallsands where ridges of Lower Devonian slate are orientated east-west above steep valleys and enter Start Bay at right angles to the coastline (Robinson, 1961). Blackpool Sands lies in the lee of the prevailing south-westerly winds and waves, protected by the South Hams. However, this wave-dominated coastline (McManus, 1975) is exposed to the southerly and easterly weather systems off Start Bay.

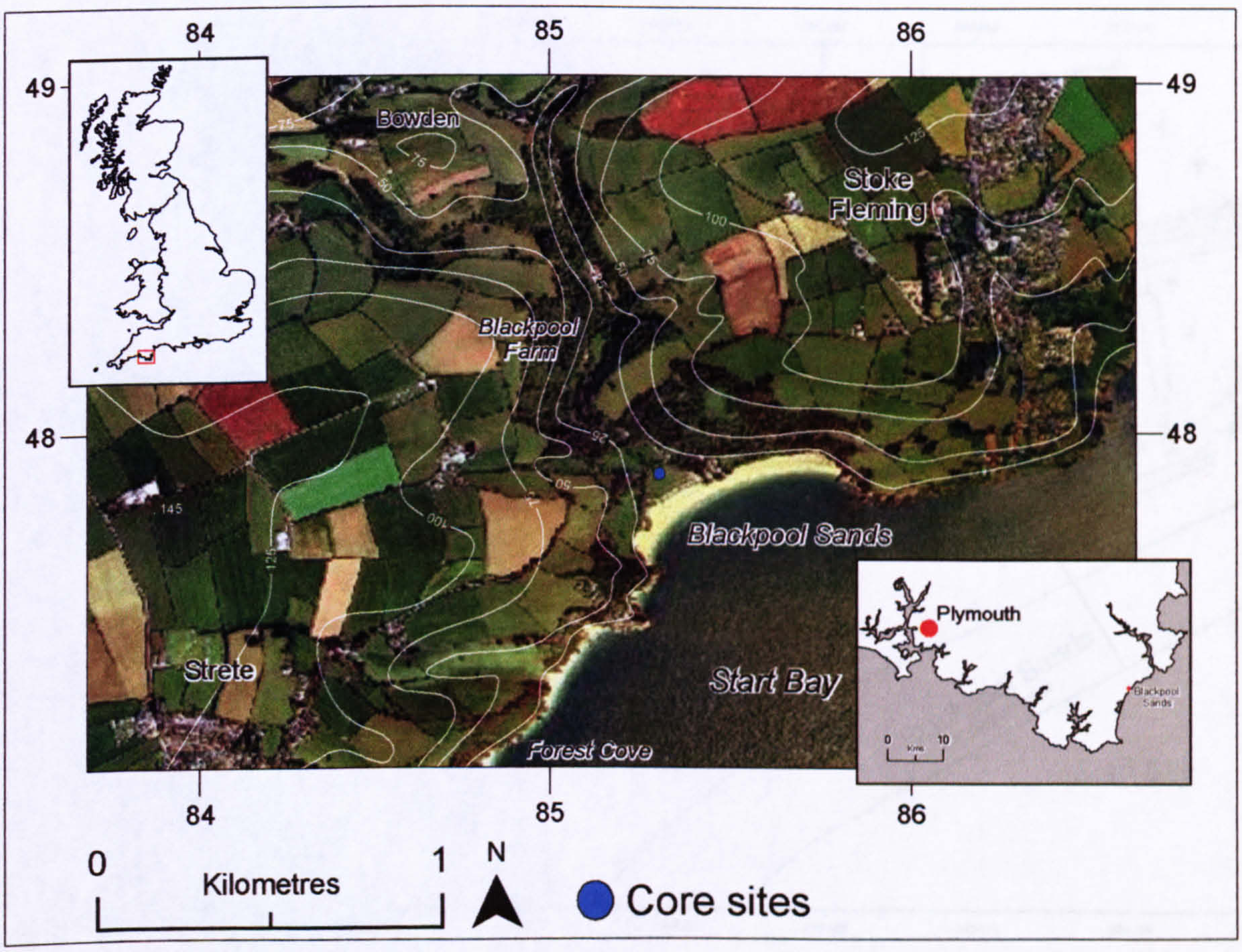


Figure 7.1. Aerial view of Blackpool Sands. Source: http://www.multimap.com. Extract from Great Britain map and aerial photograph of South Hams District at scale 1:25 000 (2002). Heights in metres above MSL.

Radiocarbon laboratory number	Location (OSGR)	^{14}C age (years BP)	Max (Median) Min Cal age (BP) (± 2σ range)	Sample Height (m OD)	Altitude (m RSL)	Source
SRR318	Blackpool Sands SX855478	2541 ± 67	2776 (2729) 2358$	+1.0 to -1.0	0 ± 0.75*	Morey (1974)
NPL86	Teignmouth SX940730	8580 (+830 –755)	11933 (9691) 7758	–24 ± 1.0	–25.2 ± 1.66	Clarke (1970)

Table 7.1. Sea-level index point data from the submerged forest on Blackpool Sands. SLIP SRR318 is not recorded in the Holocene SLIP database (2003) and SLIP NPL86 (from offshore Teignmouth) is a rejected limiting date (Horton, pers. comm. 2003). $Calibration carried out by the author of this project. *Altitude data from Heyworth and Kidson (1982) not adjusted to RSL. However, applying MSL data from Table 7.2, index point SRR318 may plot approximately –3.06 ± 0.75 m RSL.

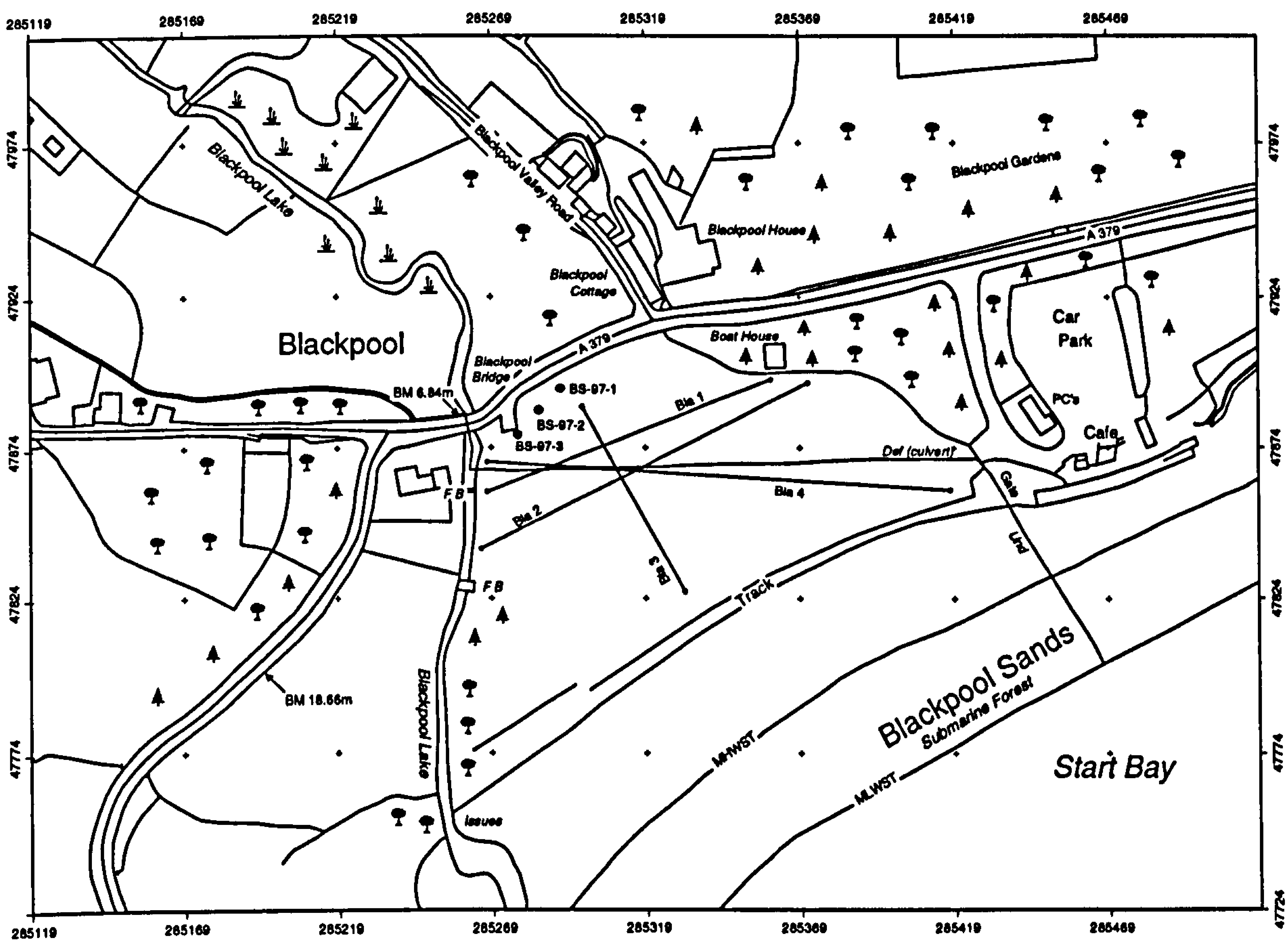


Figure 7.2. Blackpool Sands back-barrier system. Borehole locations: cores BS-97-1, 2 and 3. Electrical resistivity survey lines: Bla 1, 2, 3 and 4. Source: http://digimap.edina.ac.uk/service/advanced/carto_welcome.html. Map extract at scale 1:2500 (2002).

Morey (1974) suggested that beach movements along this stretch of the south-east Devon coastline are "sudden and catastrophic". Ancient peat marsh and forested land along the coastline appear to have been rapidly inundated by sub- and supratidal barrier beaches migrating landward under the influence of post-glacial sea-level rise (Morey, 1976). The submerged forest at Blackpool was first reported by Pengelly (1869) suggesting that it had been "concealed by the sand thrown up by the waves". He calculated that, based on local tidal conditions at the time, submergence of the forest bed had amounted to at least eighteen feet (approximately 5.5 m). Morey (1974) cited a radiocarbon date approximating the start of this event (*ca.* 2776 – 2358 cal years BP, Table 7.1). Tree growth at these sites are thought to have been terminated by "washover events" similar to those mapped in Slapton Ley (Morey, 1983a), occurring around the time of barrier closure and the establishment of back-barrier freshwater lake systems.

7.1.2 Geology

Blackpool Sands is on the northern margin of the Start Point Complex and is formed of Lower Devonian rocks with calcareous bands and basic igneous rocks occurring on certain horizons (Ussher, 1890, 1904; Orme, 1960; Dineley, 1961, Figure 7.3). The solid geology of Blackpool Sands and the Strete coast is partly siliceous slates with some hard grit and quartzite beds. Decomposed calcareous fossiliferous seams are also present in the Meadfoot Group slates and shales. Orme (1960) traced a 24 ft. (7.32 m) strandline cut into the Meadfoot Beds along the Strete coastline dating from the later stages of the last Interglacial. The emergent coastline at the time was more regular than the present submerging one. Igneous rocks, locally present in Dartmouth slates and Ringmore beds, are found at Blackpool, in particular the outcrop at Matthew's Point, and along the coast below Stoke Fleming (Orme, 1960). The Quaternary deposits of the Blackpool coastline include coarse sand and shingle beaches, and alluvial sediments (Ussher, 1904).

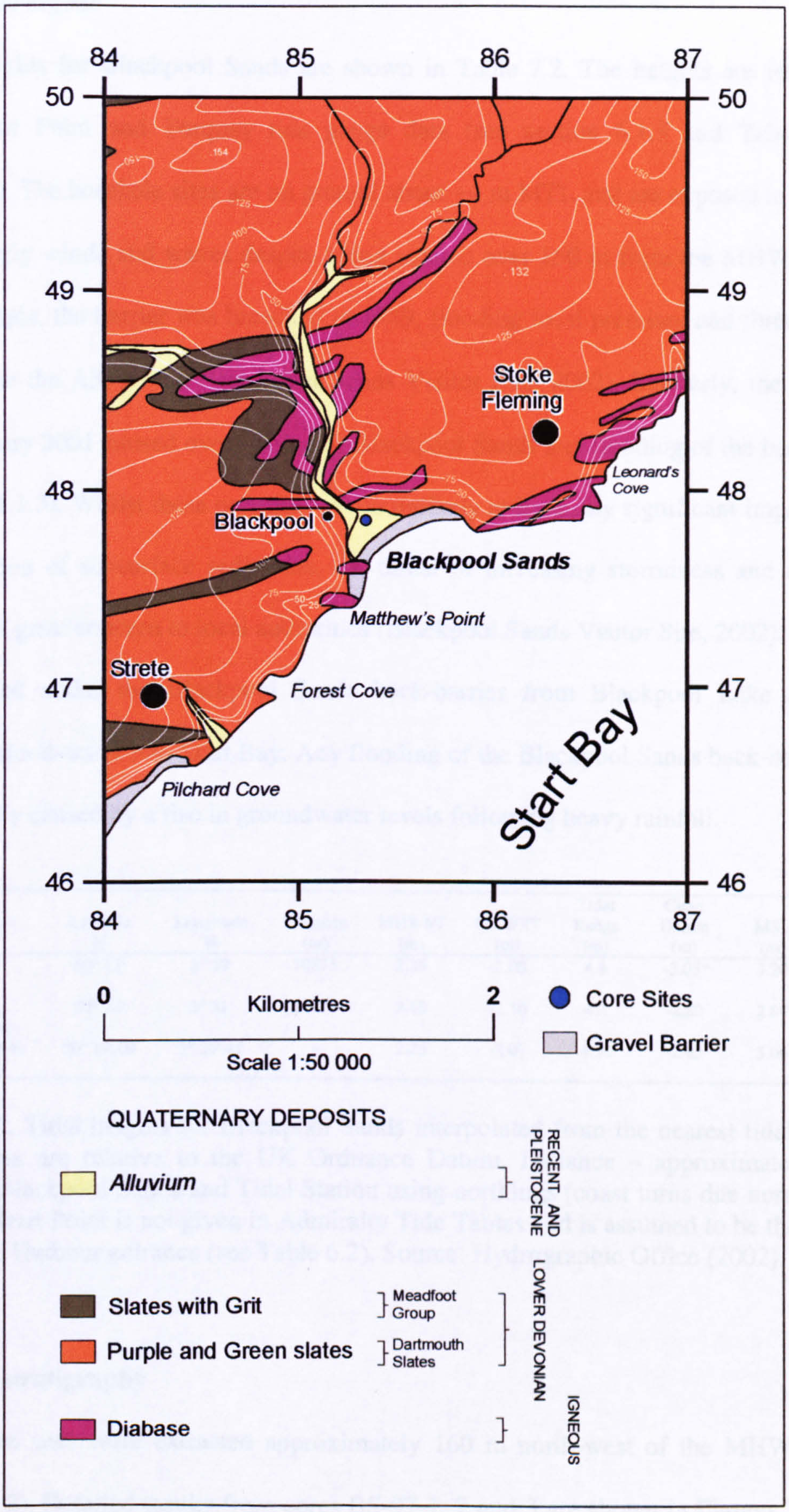


Figure 7.3. The solid geology and Quaternary deposits of Blackpool Sands and the surrounding country. Source: Geological Survey of Great Britain (England and Wales) Sheets 355 and 356, Kingsbridge and Start Point, 1:50 000 (1975). Heights in metres above MSL.

7.1.3 Tidal regime

Tidal heights for Blackpool Sands are shown in Table 7.2. The heights are interpolated from Start Point and Torquay tide gauge data (see section 6.1.3 and Table 6.2 for rationale). The borehole sites are on ground about +4 m MTL and are exposed to southerly and easterly winds and waves despite being located over 150 m from the MHWST mark. For example, the barrier was breached in 1989, flooding local premises and threatening to undermine the A379 road (Blackpool Sands Visitor Site, 2002). Similarly, the events of 11th January 2001 caused overtopping of Blackpool Sands and flooding of the back-barrier (section 6.1.3). Whilst these occurrences are unlikely to have any significant impact on the preservation of subsurface sediments, the threat of increasing storminess and rising sea level is of great concern to local authorities (Blackpool Sands Visitor Site, 2002). A culvert constructed under the Blackpool Sands back-barrier from Blackpool Lake assists in draining flood-water into Start Bay. Any flooding of the Blackpool Sands back-barrier area is normally caused by a rise in groundwater levels following heavy rainfall.

Tidal Station / Site	Latitude N	Longitude W	Distance (m)	MHWST (m)	MLWST (m)	Tidal Range (m)	Chart Datum (m)	MSL (m)	MTL (m)
Start Point, Start Bay	50° 13'	3° 39'	10875	2.35	-2.05	4.4	-3.05*	3.20	0.15
Torquay (Standard Port)	50° 28'	3° 31'	16275	2.10	-2.10	4.2	-2.80	2.84	0.04
Blackpool Sands, Start Bay	50° 19'.09	3° 36'.41	-	2.25	-2.07	4.32	-2.95	3.06	0.11

Table 7.2. Tidal heights for Blackpool Sands interpolated from the nearest tidal stations. All heights are relative to the UK Ordnance Datum. Distance – approximate distance between Blackpool Sands and Tidal Station using northings (coast turns due north at Start Point). *Start Point is not given in Admiralty Tide Tables and is assumed to be the same as Salcombe Harbour entrance (see Table 6.2). Source: Hydrographic Office (2002).

7.2 Lithostratigraphy

Three core sites were extracted approximately 160 m north-west of the MHWST mark (Figure 7.2). Detailed results from cores BS-97-1, 2 and 3 are shown in Figures 7.4 to 7.6 and are correlated in Figure 7.7. Further stratigraphical reconstruction using geophysical survey is depicted in Figures 7.8 to 7.11.

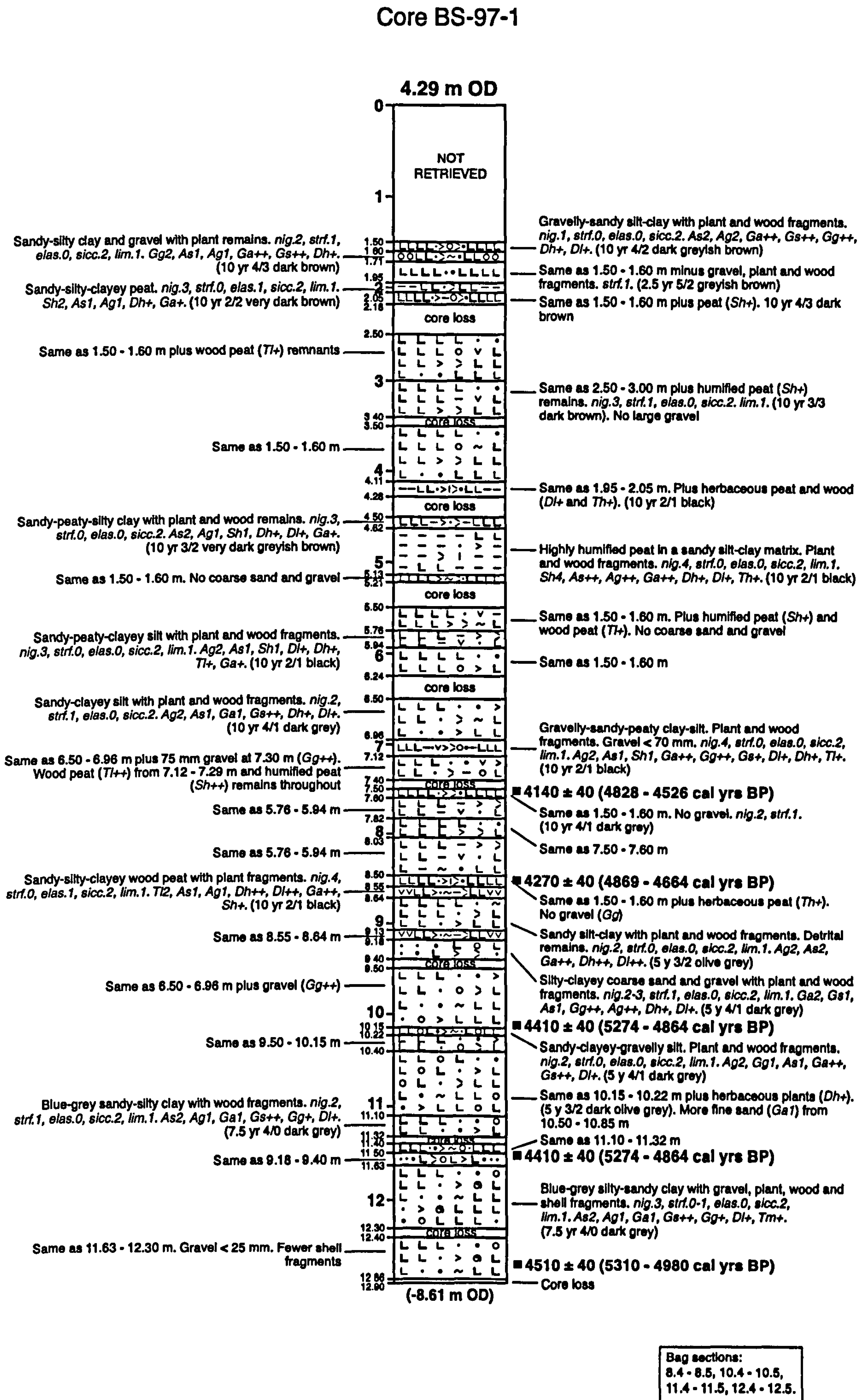


Figure 7.4. Lithostratigraphy of core BS-97-1.

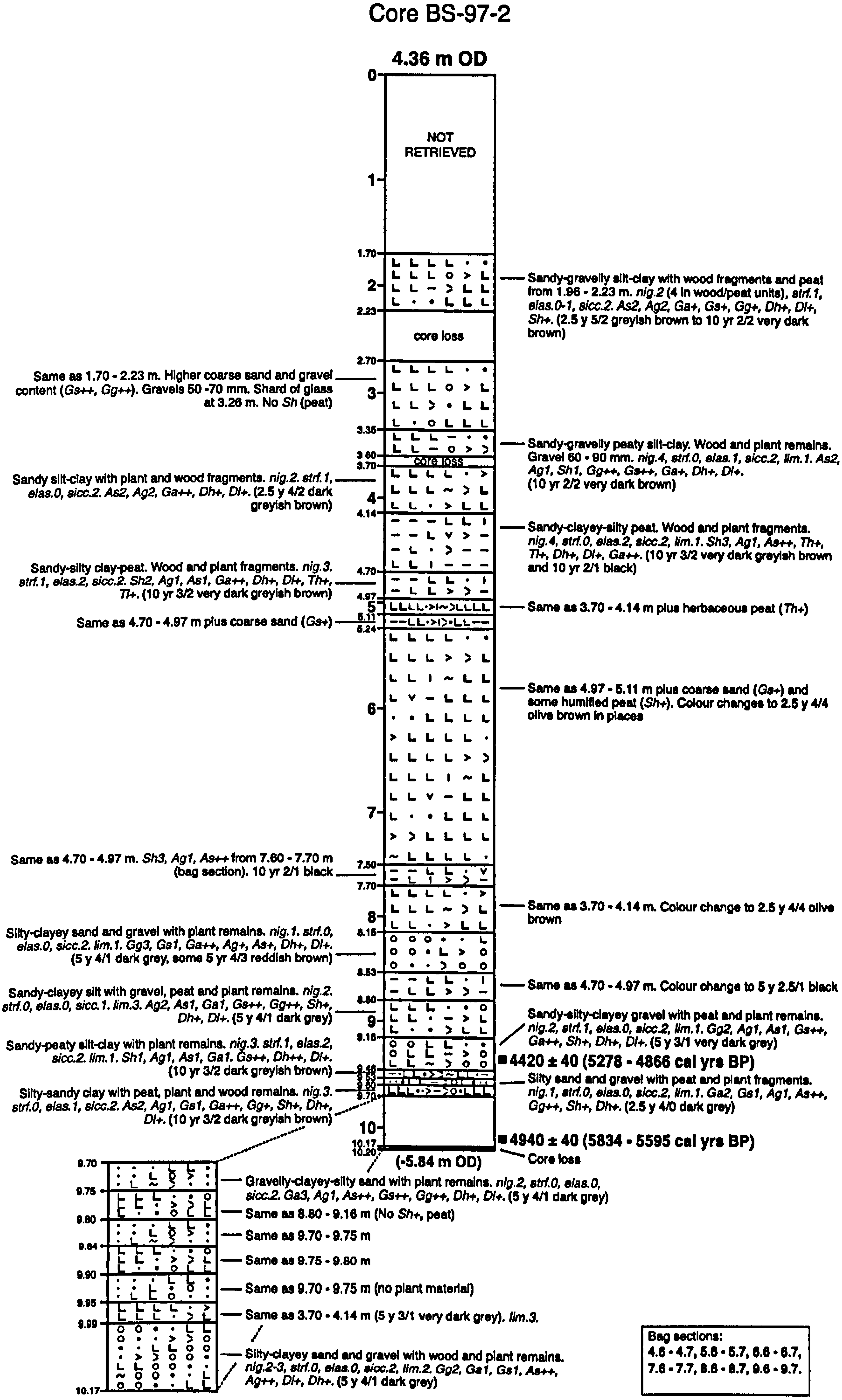


Figure 7.5. Lithostratigraphy of core BS-97-2.

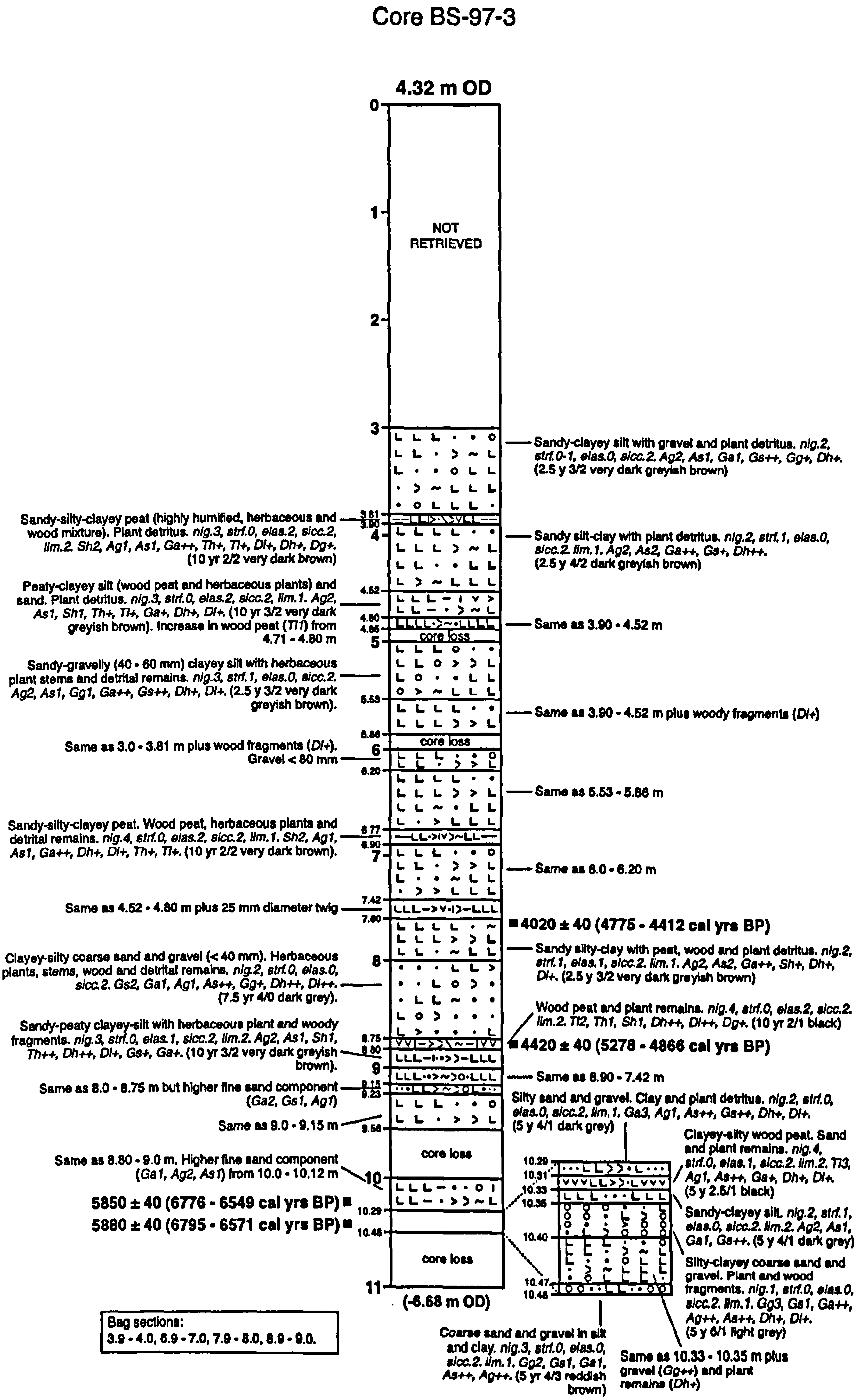


Figure 7.6. Lithostratigraphy of core BS-97-3.

7.2.1 Core BS-97-1

Core BS-97-1 (Figure 7.4) is located at SX 85292 47893 approximately 30 m east of Blackpool Lake (Figure 7.2) and is 12.90 m long. The uppermost 1.5 metres of fill and topsoil were not retrieved, 10 sections suffered core loss and 4 sections included bagged sediment. No fractured slate facies is found at the base of the borehole. The drill may not have reached the bedrock surface or it may have stopped directly above it. The basal section (12.86 to 11.63 m) contains blue-grey silty-sandy-clay with shell and wood fragments and is overlain by a dark olive-grey sandy-clayey-silt, gravel and wood facies (11.63 to 9.18 m). A dark greyish-brown sandy-clayey-silt (9.18 and 7.50 m) containing thin layers of black minerogenic-peat is covered by sandy-silty-clay, large (75 mm) slate and shale gravel and wood (7.50 to 6.96 m). Sandy-clayey-silt (6.96 to 5.13 m) is then overlain by minerogenic-peat (5.13 to 4.11 m) and sandy-silty-clay with wood fragments (4.11 to 2.50 m). The uppermost lithofacies is predominantly dark greyish-brown minerogenic-peat and gravel (2.50 to 1.50 m).

7.2.2 Core BS-97-2

Core BS-97-2 (Figure 7.5) is located at SX 85285 47886 approximately 9.8 m south-west of core BS-97-1 (Figure 7.2) and is 10.20 m long. The uppermost 1.70 metres of fill and topsoil were not retrieved, three sections suffered core loss and six sections included bagged sediment. The core base comprises fractured dark grey slates in silty-clayey-sand and wood (10.17 to 9.99 m). This is overlain by alternating layers of dark-grey sandy-clayey-silt with black humified wood-peat, and fine sandy-clayey-silty gravel (9.99 to 7.50 m). The overlying lithofacies is predominantly dark greyish-brown sandy-silty clay with plant and wood fragments (7.50 to 5.24 m) and a dark greyish-brown minerogenic-wood peat (5.24 to 4.14 m) is overlain by sandy-gravelly silty-clay and peat (4.14 to 1.70 m). Angular slate and shale gravel (90 mm) and coloured glass was identified as the infill section (3.70 to 2.70 m).

7.2.3 Core BS-97-3

Core BS-97-3 (Figure 7.6) is located at SX 85278 47878 approximately 11.8 m south-west of core BS-97-2 (Figure 7.2) and is 11 m long. The uppermost 3 metres were not retrieved, 4 sections suffered core loss and 4 sections included bagged sediment. The core base comprises fractured reddish-brown slate overlain by layers of sandy-clayey-silt and gravel, minerogenic-wood peat, sand lenses and sandy-clayey-silt and wood (10.48 to 10.00 m). Layers of dark greyish-brown silty-sandy-clay, humified wood-peat and sandy-gravel are found between 10.00 and 8.75 metres. Coarse grey gravel in sandy-silty-clay (8.75 to 8.00 m) is found overlying this, above which sandy-silty-clay and minerogenic wood-peat (8.00 to 6.20 m) is covered by a similar sequence containing large (80 mm) gravel particles (6.20 to 5.00 m). The uppermost lithofacies is very dark greyish-brown organic sandy-clayey-silt (5.00 to 3.00 m). The section not retrieved (3 metres to ground level) was identified on site as gravelly-sandy-clayey-silt, peat and topsoil.

7.2.4 Core correlation

The lithostratigraphic facies described in sections 7.2.1 to 7.2.3 are correlated in Figure 7.7. The basal lithostratigraphy of Blackpool Sands consists of fractured slate, sandy-clayey-silt and gravel below –6 m OD. This is overlain in most cores by the remnants of a basal minerogenic wood-peat. The minerogenic phase is dominated by sandy-clayey-silt and coarse gravel to roughly –3 m OD. Very thin layers of humified wood-peat are present in the minerogenic and gravel core facies. No pure organic phase is preserved in the Blackpool Sands cores, instead the uppermost sequence (from around OD) is predominantly infill (glass shard at +1.1 m OD in core BS-97-2) of sandy-silty-clay and gravel with a thin layer of peaty topsoil to above +4 m OD.

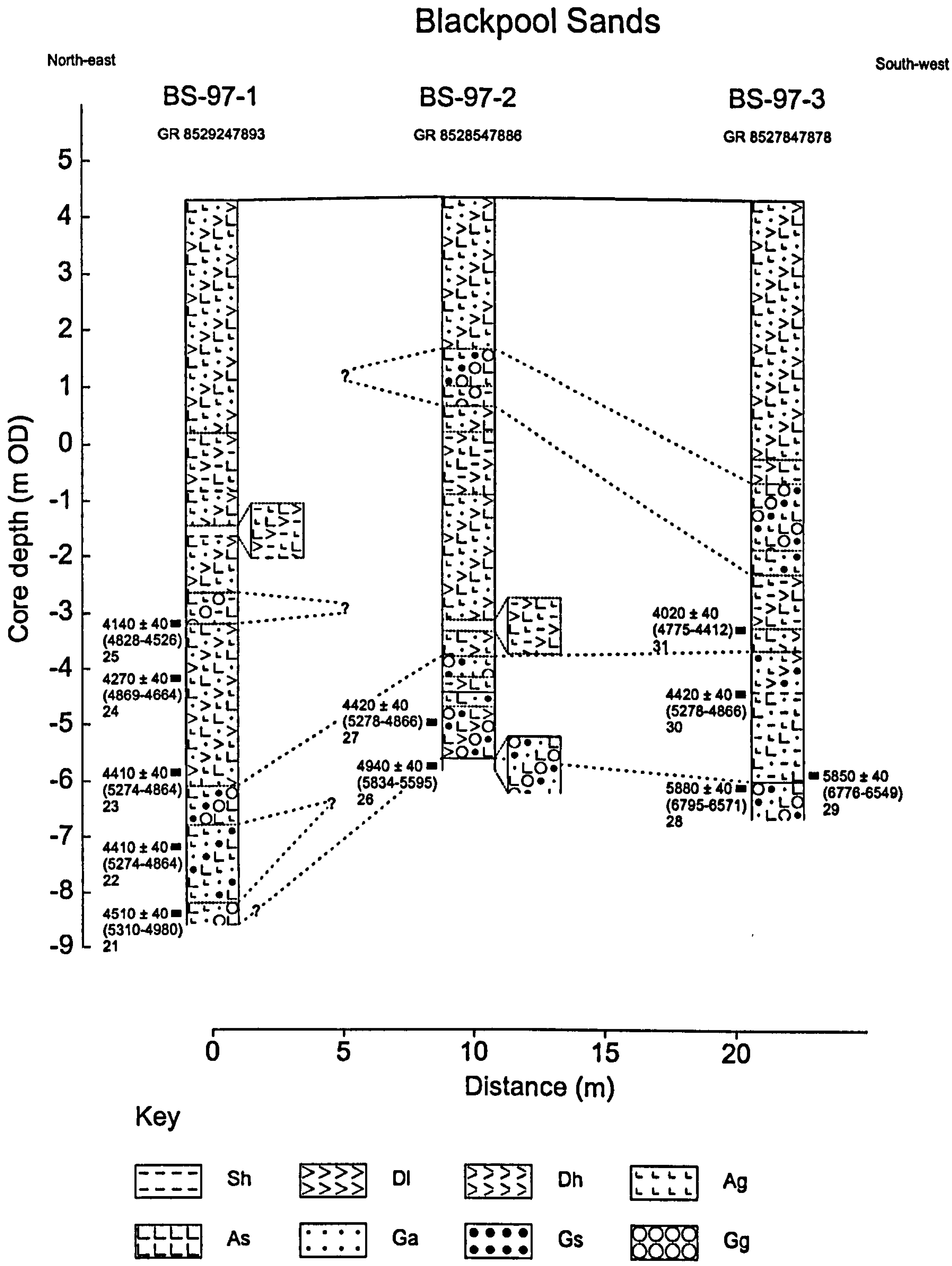


Figure 7.7. Blackpool Sands core correlation. SLIP numbers are shown below ages recorded in cal. years BP (see Table 7.4).

7.3 Electrical resistivity survey

Two shore-parallel (Figures 7.8 and 7.9), one shore-normal (Figure 7.10) and one shore-diagonal (Figure 7.11) electrical resistivity surveys were carried out and electrodes were spaced at 5 m intervals on all cables. Cores BS-97-1, 2 and 3 are used as ground-truths for the survey. The geographical locations of the survey lines are shown in Figure 7.2. A summary of the electrical resistivity of sediments at Blackpool Sands is given in Table 7.3.

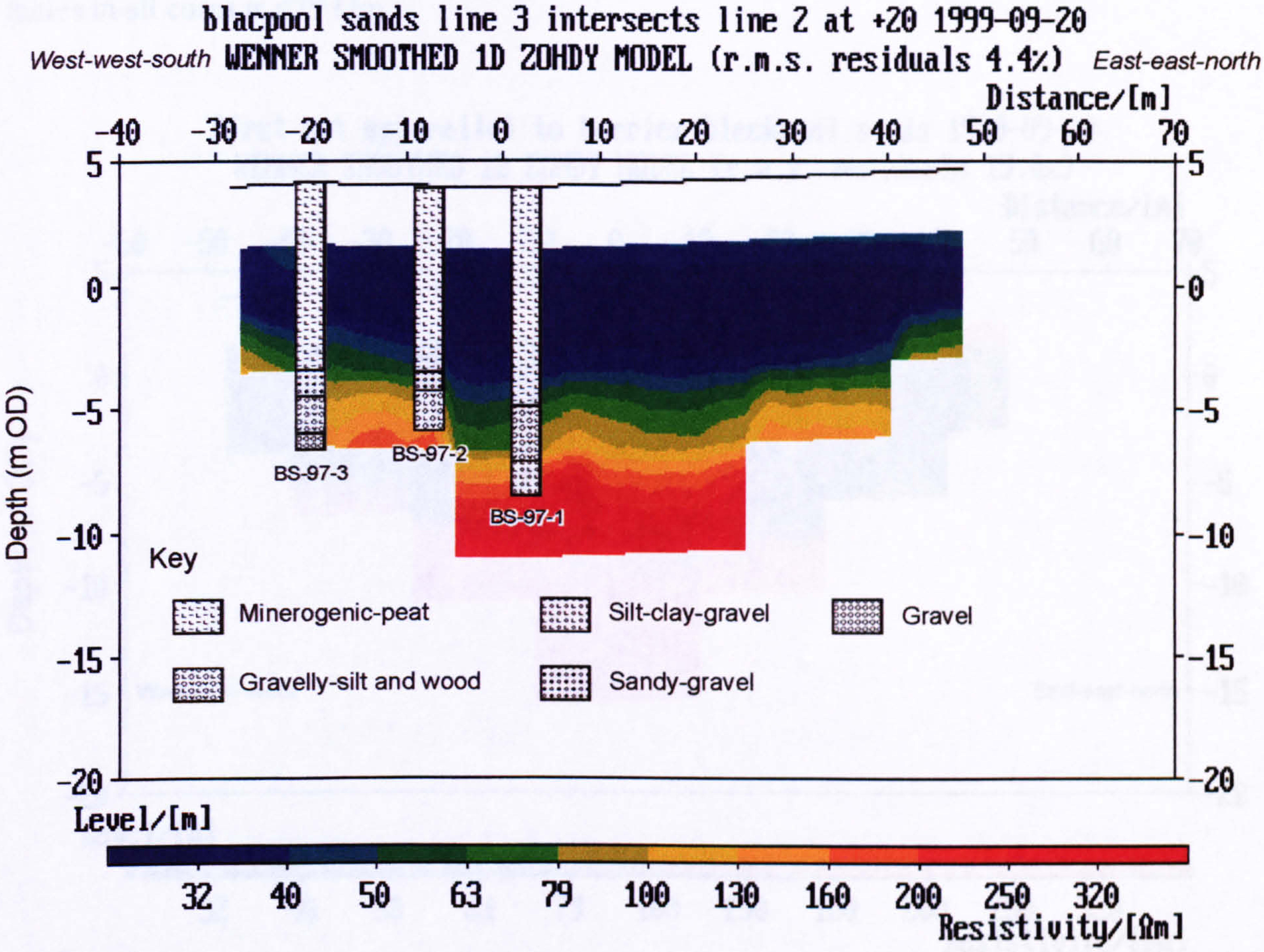


Figure 7.8. Shore-parallel resistivity profile (Bla 1).

7.3.1 Shore-parallel resistivity survey (Bla 1)

Bla 1 extends 105 m from Blackpool Lake footbridge at SX 85268 47859 (+3.86 m OD) to the Boat House at SX 85360 47896 (+5.06 m OD) and heights were recorded from 4 out of 22 electrodes. The borehole transect (Figure 7.7) is about 10 m south of the A379 (Figure 7.2) and does not allow adequate distance (100 m) for an electrical resistivity survey. The shore-parallel survey was therefore run out 25 m south-east of the boreholes and the cores positioned relative to this (Figure 7.8). The basal gravel of core BS-97-3 has resistivity values of between 130 and 160 Ωm at –7 m OD. Core BS-97-1 is located in the basin low and the basal silt-clay-gravel facies is between 100 and 250 Ωm at –9 m OD. A similar basal facies in core BS-97-2, although more organic than the other two cores, is between 79 and 160 Ωm at –6 m OD. The overlying gravel-dominated minerogenic facies in all cores has resistivity values between 50 and 160 Ωm below –4 m OD. Minerogenic-peat facies in all cores is ≤79 Ωm.

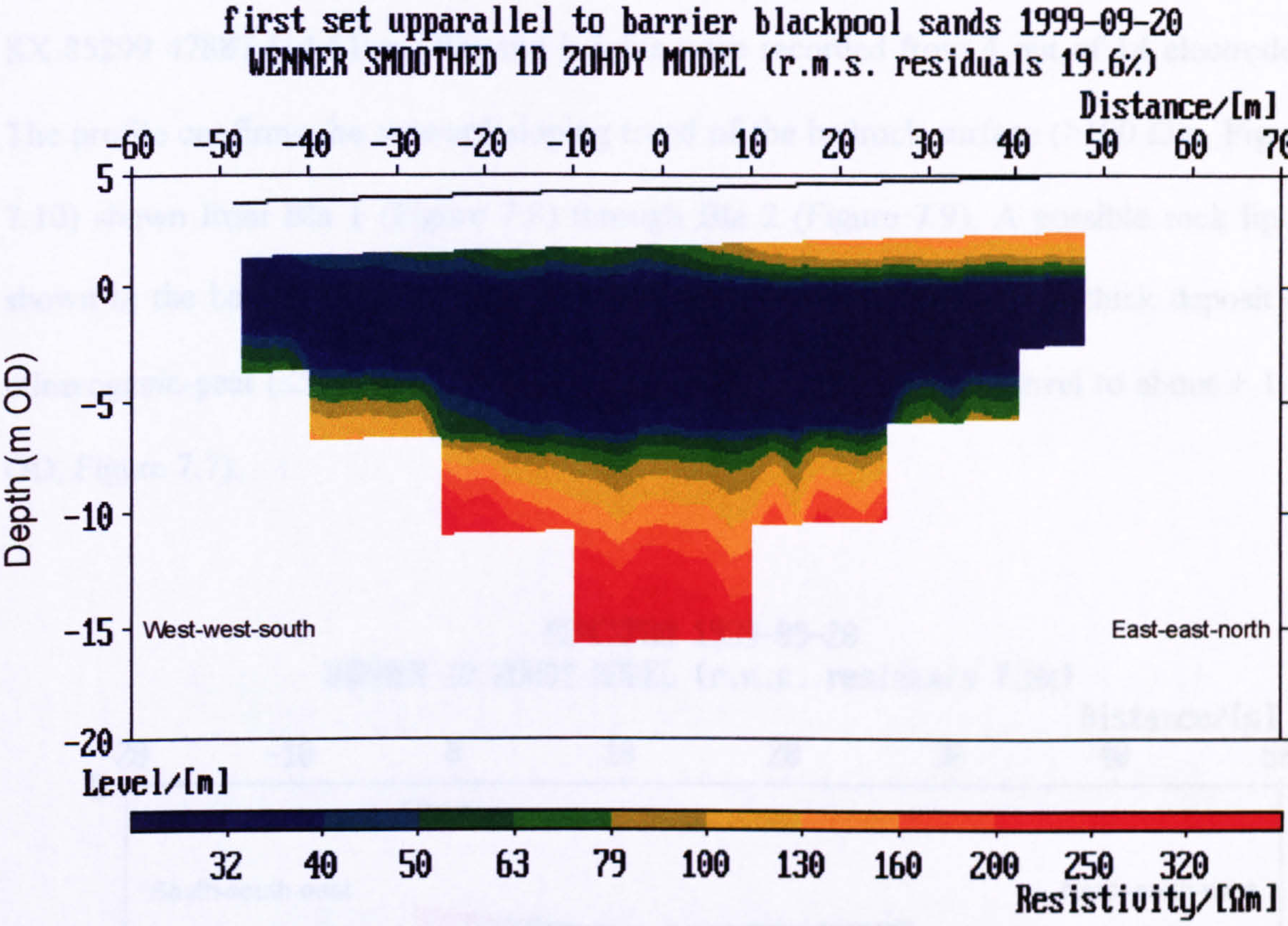


Figure 7.9. Shore-parallel resistivity profile (Bla 2).

7.3.2 Shore-parallel resistivity survey (Bla 2)

Bla 2 extends 125 m from Blackpool Lake at SX 85266 47840 (+3.78 m OD) to the low shrub at SX 85372 47895 (+5.35m OD) and heights were recorded from 3 out of 26 electrodes. Figure 7.9 reveals a basin-like profile similar to Bla 1 (Figure 7.8) delimited by the 160 Ωm isoline suggesting that the bedrock surface dips seaward to –11 m OD. This is probably overlain by a similar organo-minerogenic-gravel sequence around 50 to 160 Ωm below –7 m OD at –5 m distance. A thick layer of minerogenic-peat ≤79 Ωm, similar to Bla 1, is overlain by sediment of a higher resistance (<160 Ωm at ground level to OD) across most of Bla 2. This is probably the gravel infill extending from the car park entrance (SX 85430 47865, Figure 7.2) diagonally across the back-barrier surface.

7.3.3 Shore-normal resistivity survey (Bla 3)

Bla 3 extends 65 m from the track at SX 85332 47826 (+5.40 m OD) to core BS-97-1 at SX 85299 47887 (+4.51 m OD) and heights were recorded from 4 out of 14 electrodes. The profile confirms the seaward sloping trend of the bedrock surface (>100 Ωm, Figure 7.10) shown from Bla 1 (Figure 7.8) through Bla 2 (Figure 7.9). A possible rock lip is shown in the basal profile at zero point (a high of <55 Ωm) below the thick deposit of minerogenic-peat (≤55 Ωm) and gravel infill (<78 Ωm from ground level to about + 1 m OD, Figure 7.7).

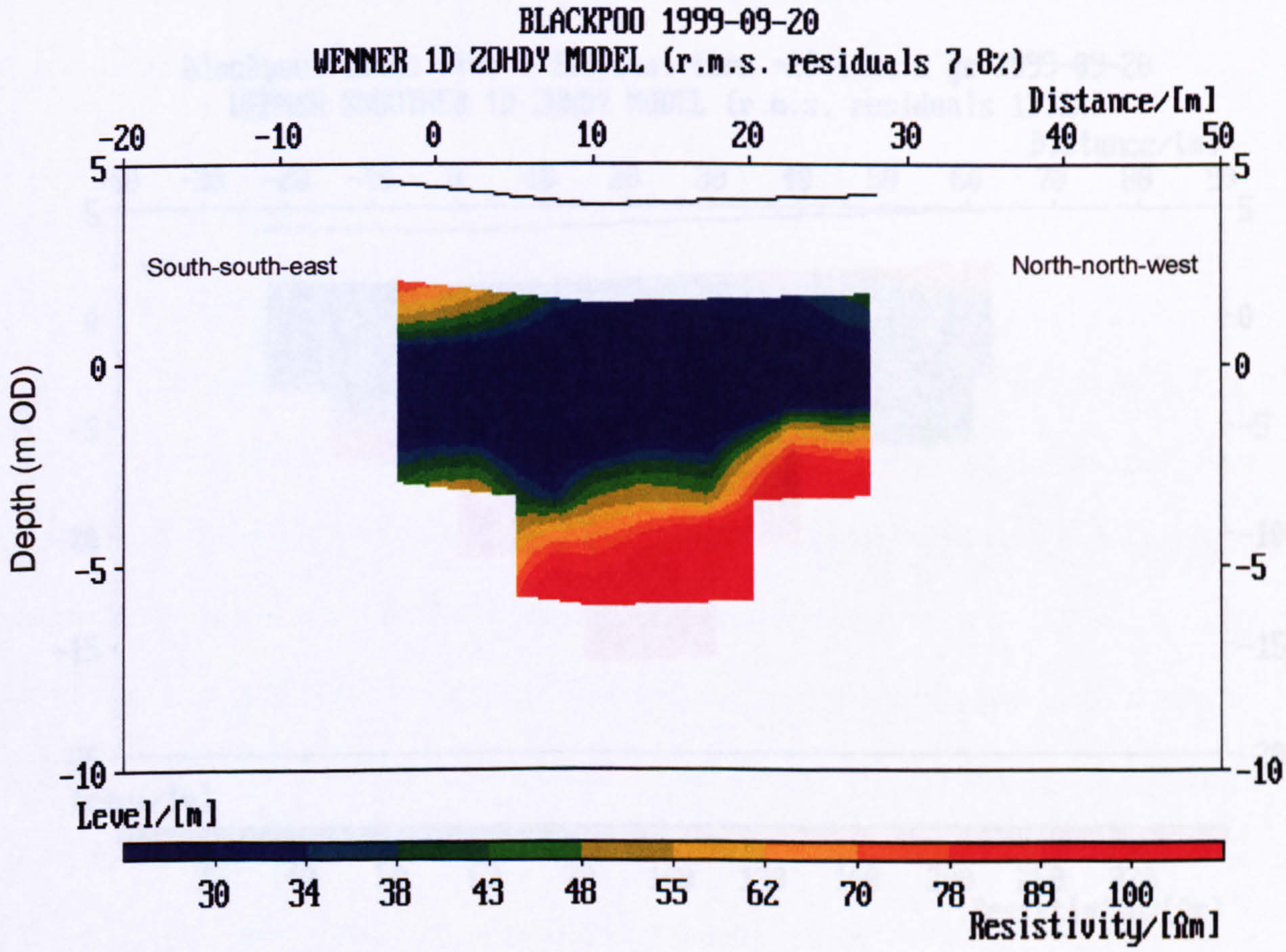


Figure 7.10. Shore-normal resistivity profile (Bla 3).

7.3.4 Shore-diagonal resistivity survey (Bla 4)

Bla 4 extends 125 m from Blackpool Lake gate at SX 85268 47869 (+3.90 m OD) to the car park gate at SX 85418 47860 (+5.45 m OD) and heights were recorded from 4 out of 26 electrodes. Figure 7.11 shows the basin floor sloping gently eastward from Blackpool Lake (<160 Ωm around –6 m OD at –20 m distance) to roughly +40 m distance. At this point, a rise in the basal profile (visible at <79 Ωm, no data exists beyond –5 m OD at +60 m distance) seems to confirm the rock lip below the shingle beach inferred on Bla 3 (Figure 7.10). The overlying gravelly facies (50 to 160 Ωm, Figure 7.8) below –2 m OD, near Blackpool Lake, and minerogenic-peat facies (≤79 Ωm) is covered by gravel infill (<160 Ωm) to OD at the eastern end of the profile. This is probably the aggregate laid down at the entrance to the overflow car park (Figure 7.2).

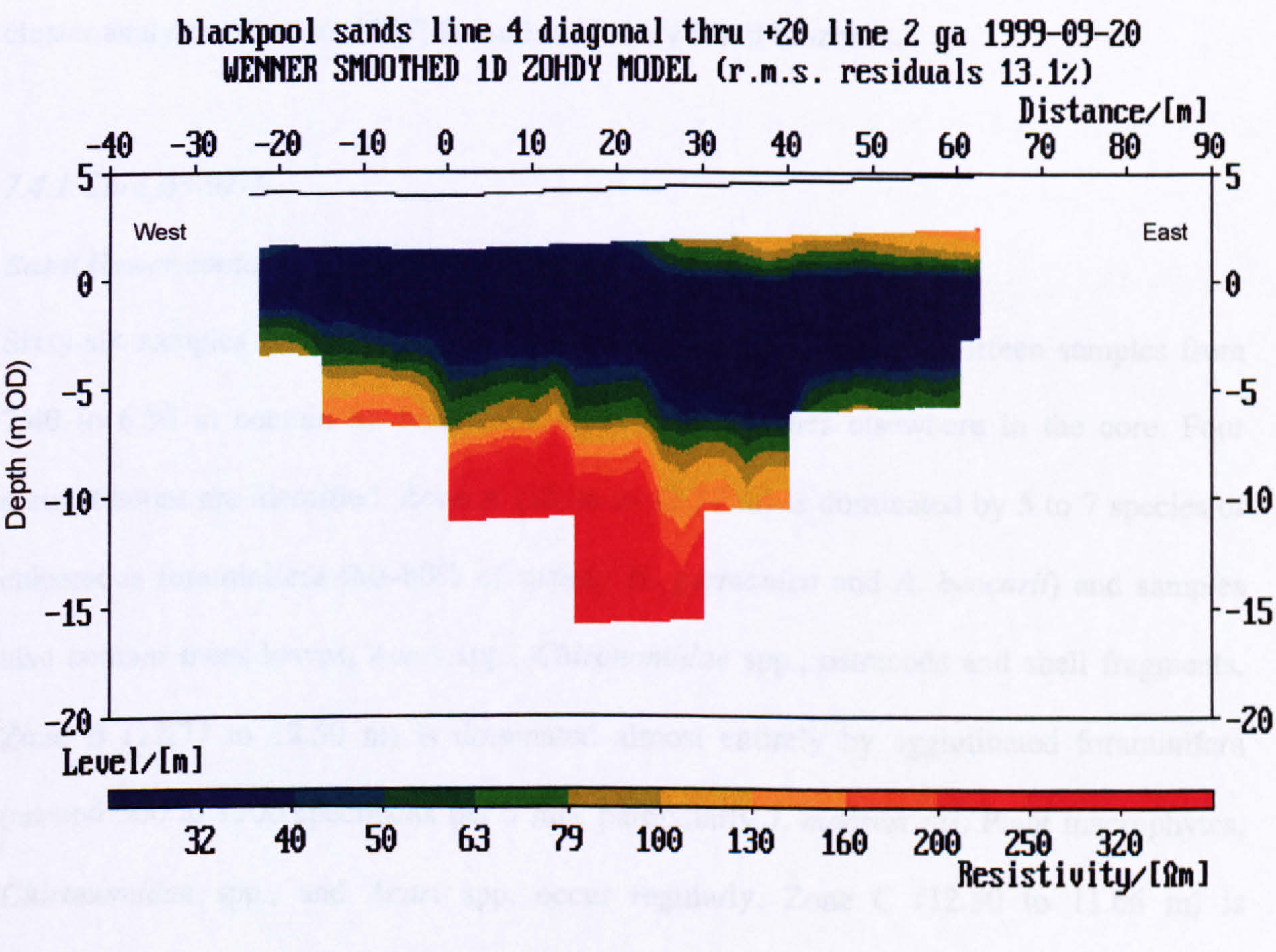


Figure 7.11. Shore-diagonal resistivity profile (Bla 4).

Survey	Basal gravel	Minerogenic fines	Gravel infill
Shore-parallel (Bla 1)	50 – 250	≤79	~ 50
Shore-parallel (Bla 2)*	50 – 160	≤79	< 160
Shore-normal (Bla 3)*	55 – 100	≤55	< 78
Shore-diagonal (Bla 4)*	50 – 160	≤79	< 160
Summary	50 – 250	≤79	~ 50 – 160

Table 7.3. Summary of the electrical resistivity of sediments at Blackpool Sands (units in Ωm). *Results for surveys Bla, 2, 3 and 4 are interpolated from borehole (depth) information. Basal gravel includes organic- and minerogenic-rich gravel. Minerogenic fines include sand, silt, clay and minerogenic-peat.

7.4 Biostratigraphy

The biostratigraphy of Blackpool Sands is reconstructed using foraminifera identified from cores BS-97-1, 2 and 3 (Figures 7.12, 7.13 and 7.14). The indicative meaning of each sample is calculated based on the transfer function described in chapter 3 and the CONISS cluster analyses (Grimm, 1987) is used to identify fossil biozones.

7.4.1 Core BS-97-1

Basal (lower contact) to in-core (upper contact) section (Figure 7.12)

Sixty-six samples from 12.86 to 7.50 m contain foraminifera and thirteen samples from 7.40 to 6.50 m contain no foraminifera plus five samples elsewhere in the core. Four distinct zones are identified. Zone A (12.86 to 12.77 m) is dominated by 5 to 7 species of calcareous foraminifera (60–80% of mainly *H. germanica* and *A. beccarii*) and samples also contain moss leaves, *Acari* spp., *Chironomidae* spp., ostracods and shell fragments. Zone B (12.77 to 12.50 m) is dominated almost entirely by agglutinated foraminifera (around 500 to 1500 specimens per 5 ml), particularly *J. macrescens*. Plant macrophytes, *Chironomidae* spp., and *Acari* spp. occur regularly. Zone C (12.30 to 11.66 m) is predominantly calcareous (around 90 to 100%). The remaining core record (11.66 to 7.50 m) is treated as one zone although CONISS shows a number of distinct clusters (interruptions are due mainly to bagged sections, Figure 7.12). Zone D contains only agglutinated species and is dominated by *J. macrescens*, *T. inflata* and *M. fusca*. Counts are very low (4 to 212 per 5 ml) and flora and fauna similar to zone B, including *Chara* oospores, are found in some samples. The uppermost samples (7.50 to 6.50 m) contain lower numbers of plant macrophytes.

7.4.2 Core BS-97-2

Basal (lower contact) to in-core (upper contact) section (Figure 7.13)

Thirty-eight samples between 10.17 and 9.35 m contain only agglutinated foraminifera except at 10.16, 10.14, 9.94, 9.90 and 9.80 where very low numbers (2 to 4 per 5ml) of *A. beccarii* are found. Sixteen samples between 9.35 and 8.20 m and one at 9.48 m contain no foraminifera and one zone is identified. Zone A (10.17 to 9.35 m) is dominated by agglutinated foraminifera (90–100%), in particular *J. macrescens* (around 80%). *T. inflata* and *M. fusca* also occur frequently. At the base (below 9.70 m) numbers of foraminifera are between 10 and 258 specimens per 5 ml. At 9.46 m core depth *T. inflata* reach 100% dominance but samples contain only 2 to 22 specimens per 5 ml between 9.60 and 9.35 m core depth. No shell fragments were found but moss leaves, *Acari* spp., *Chironomidae* spp., and other organic material occurred in all samples analysed.

Figure 7.12. Foraminifera stratigraphy (%) of core BS-97-1: basal (lower contact) to in-core (upper contact) section.

Figure 7.13. Foraminifera stratigraphy (%) of core BS-97-2: basal (lower contact) to in-core (upper contact) section.

Figure 7.14. Foraminifera stratigraphy (%) of core BS-97-3: basal (lower contact) to in-core (upper contact) section.

7.4.3 Core BS-97-3

Basal (lower contact) to in-core (upper contact) section (Figure 7.14)

Forty-one samples from 10.48 to 7.62 m contain foraminifera. Seven samples between 7.62 and 7.00 m and ten samples elsewhere throughout the core, most notably at the base (6 samples between 10.33 and 10.23 m), contain no foraminifera and two distinct zones are identified. Zone A (10.48 to 10.40 m) contains the most diverse assemblage, e.g., *J. macrescens*, *T. ochracea*, *T. inflata* and *M. fusca* occur with *H. germanica*, *A. beccarii*, *E. williamsoni* and *B. variabilis*, but numbers are very low (around 50–60 per 5 ml). Shell fragments are not present but moss leaves, *Acari* spp., *Chironomidae* spp., and diatoms occur in most samples. The remaining core record (10.40 to 7.62 m) is treated as one zone although CONISS shows a number of distinct clusters (interruptions are due mainly to bagged sections and core loss, Figure 7.14). Zone B is predominantly agglutinated (90–100%) consisting chiefly of *J. macrescens*, *T. inflata* and *M. fusca*. Numbers are around 100 to 200 foraminifera in samples above and below the core loss (10.00 to 9.56 m) but are frequently low (around 4 to 50 per 5 ml) above 8.85 metres. This increases to 500 per 5 ml around 7.85 m, where species diversity reaches 5, falling to four *J. macrescens* at 7.62 metres. No foraminifera were found above 7.62 m and the majority of samples contained moss leaves, *Acari* spp., and *Chironomidae* spp.

7.5 Geotechnical correction for sediment autocompaction

Sedimentary facies in cores BS-97-2 and 3 are corrected for autocompaction enabling in-core samples to be plotted nearer their original height of deposition. Sediment moisture, bulk density and liquid limits results from the core facies provide the input data for the Paul and Barras (1998) model (Figures 7.15 and 7.18). Void ratio, submerged unit weight, effective stress and the coefficient of consolidation (Figures 7.16 and 7.19) determine the total correction for each core (Figures 7.17 and 7.20).

7.5.1 Core BS-97-2

Input data (Figure 7.15)

Gravel facies contain the lowest moisture levels, e.g., 10% at 8.30 m increasing to 185% at 4.30 m in minerogenic-peat and about 160% at 7.60 m in peat. Fine minerogenic sediment moisture values are around 50%, similar to core NS-97-2 (Figure 5.21). Dry bulk density in basal gravel is about 1.5 g/cm^3, falling to 0.5 g/cm^3 in peat at 7.60 m. Minerogenic-peat and peat have higher liquid limits, e.g., 220% at 4.5 m, than minerogenic-fines and gravel, e.g., 40% at 10.16 m.

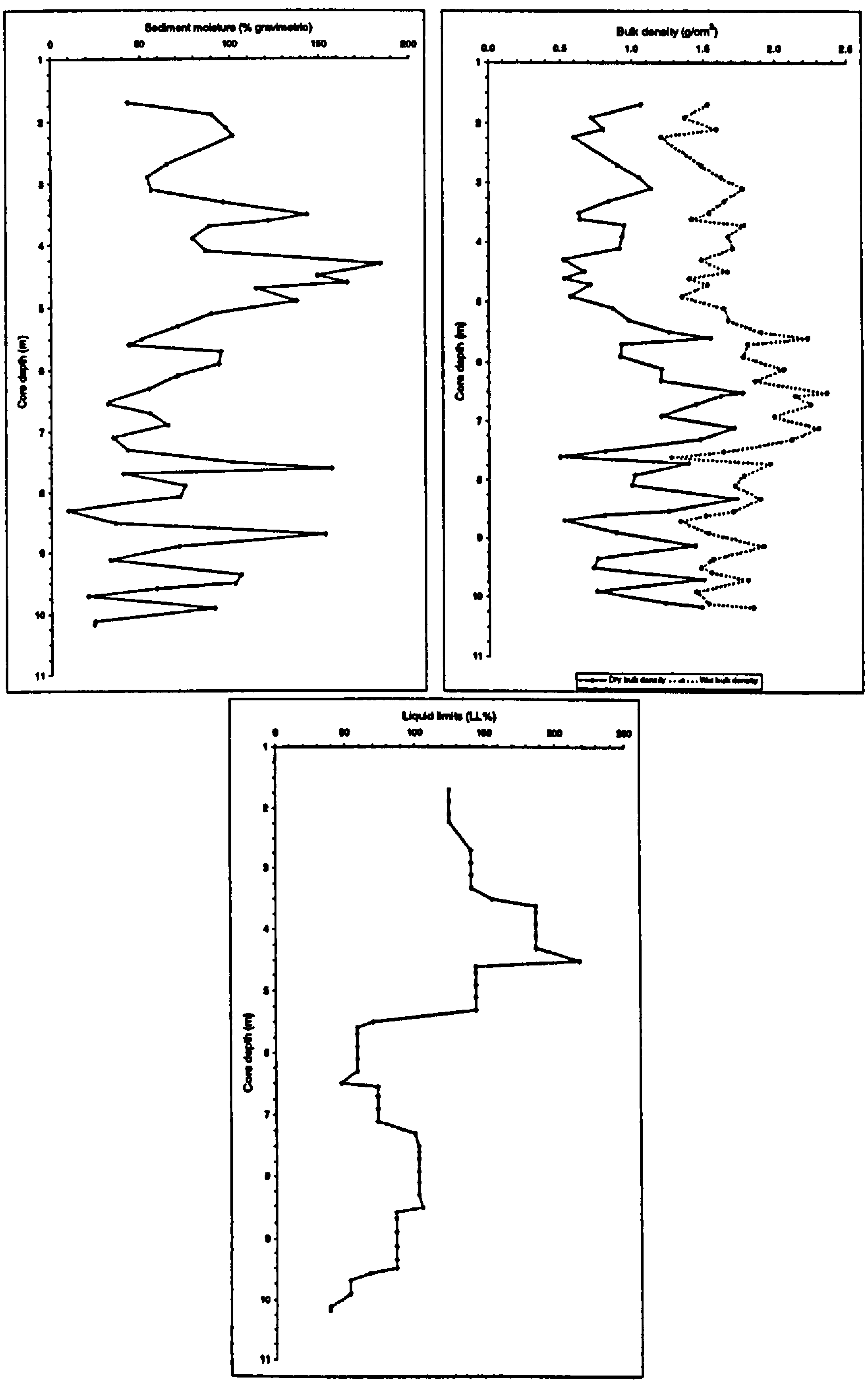


Figure 7.15. Input geotechnical data from core BS-97-2. 50 layers sampled between 0.05 and 0.23 m thick with the majority set at 0.20 m. See Figure 5.21 caption for explanation.

Additional data (Figure 7.16)

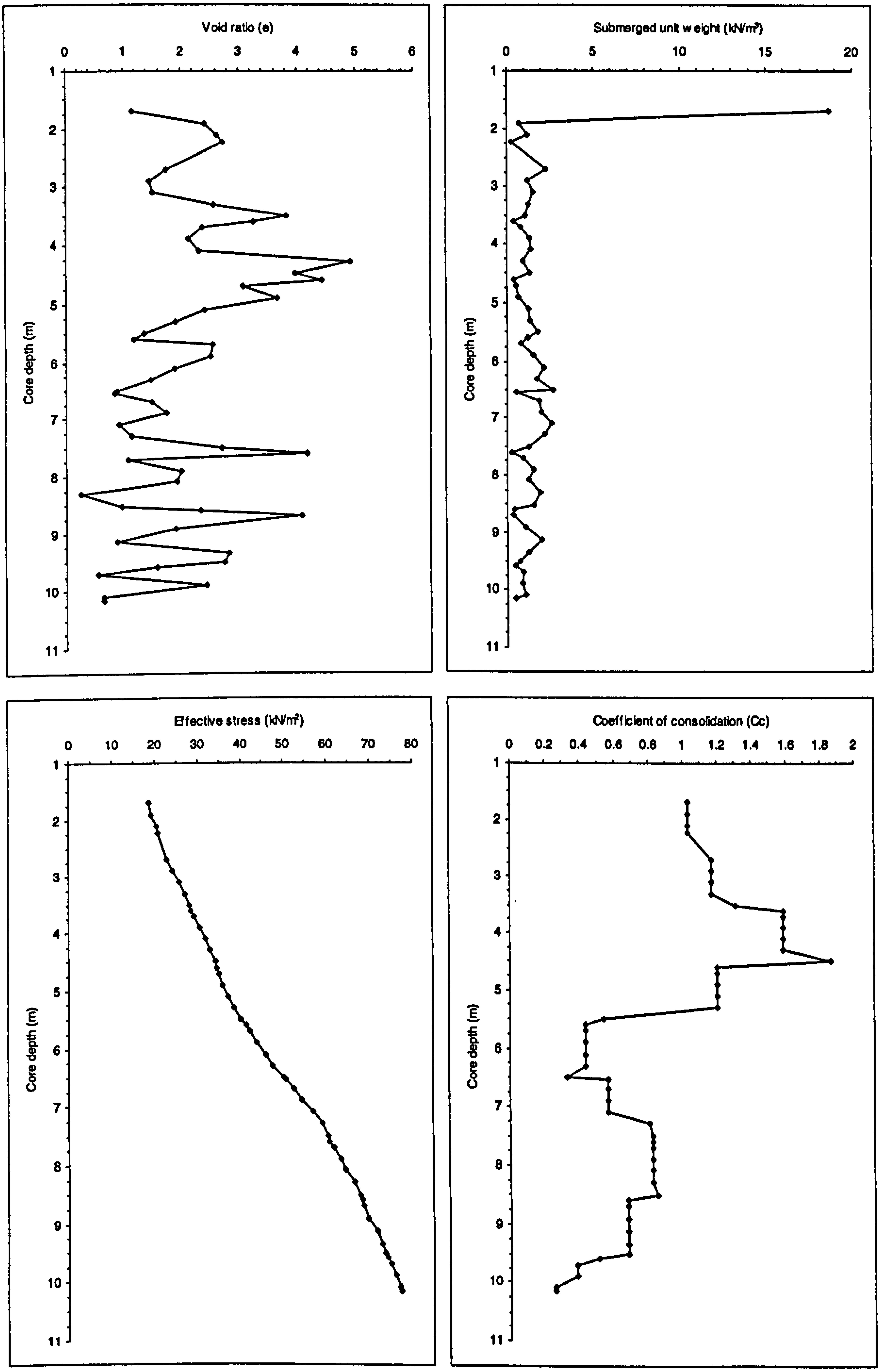


Figure 7.16. Additional geotechnical data from core BS-97-2. See Figure 5.22 caption for explanation.

Void ratio is greatest in minerogenic-peat and peat where moisture is highest, e.g., 5 at 4.30 m in minerogenic-peat and 0.3 at 8.30 m in gravel. Submerged unit weight is greater in minerogenic sediment (2.7 kN/m^3 at 6.5 m) than peat (0.3 kN/m^3 at 8.70 m). The value of 18.7 kN/m^3 at 1.70 m is the cumulative incremental submerged unit weight extrapolated for the uppermost 1.70 m. Effective stress increases with depth and the slope becomes shallower where the minerogenic content increases (5.10 kN/m^2 to 7.30 m). Post-depositional compression (Cc) is greater in peat (1.9 at 4.5 m) than minerogenic sediment (0.3 at 6.5 m).

Total geotechnical correction for core BS-97-2 (Figure 7.17)

Autocompaction is zero above the bedrock base at 10.16 m and quite small in the basal sediment, e.g., 2 cm at 10.10 m core depth. Autocompaction increases sharply to 67 cm at 8.08 m core depth, where a sequence of peat, minerogenic-peat and gravel is present (Figure 7.5). The post-depositional vertical displacement of sediment decreases between 7.10 and 5.60 m where minerogenic-fines increase (see effective stress, Figure 7.16). Autocompaction then increases in the uppermost organic-rich (peat) layers reaching almost 1 m at 2.10 m core depth (see section 7.5.3). Table 7.4 contains the total correction for SLIPs obtained from core BS-97-2.

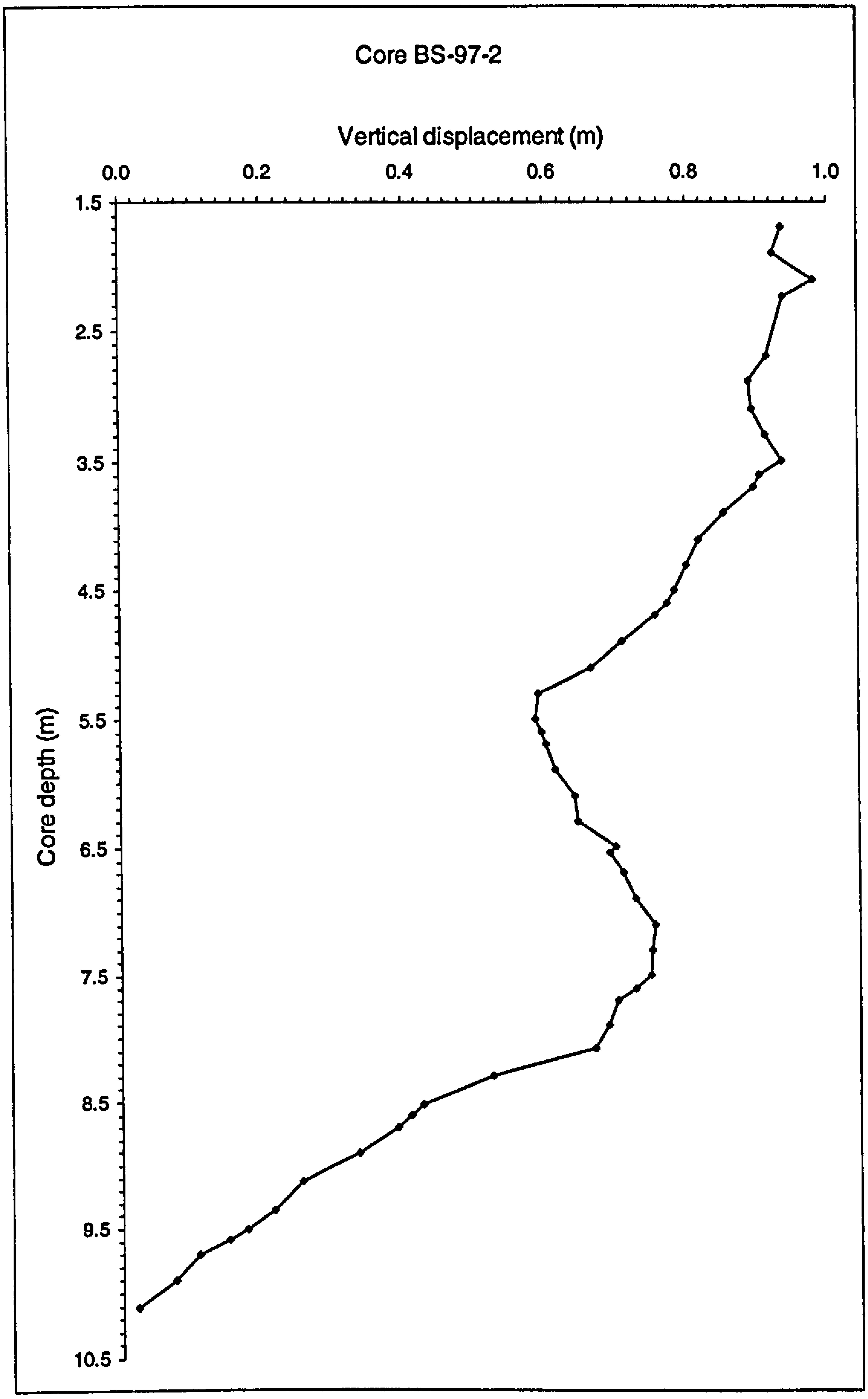


Figure 7.17. Total geotechnical correction for core BS-97-2 (ΔH in Figure 2.5). See Figure 5.23 caption for explanation.

7.5.2 Core BS-97-3

Input data (Figure 7.18)

Sediment moisture levels are lowest in sandy gravel, e.g., 15% at 9.2 m and 38% in the basal gravel, and highest in peat and minerogenic-peat (152% at 6.9 m). Dry bulk density is highest in minerogenic sediment (1.6 g/cm^3 at 6.6 m) and lowest in peat (0.5 g/cm^3 at 3.9 m). The liquid limits are lowest in gravel (32% at 8.6 m) and highest in minerogenic-peat and peat (114% at 4.6 m).

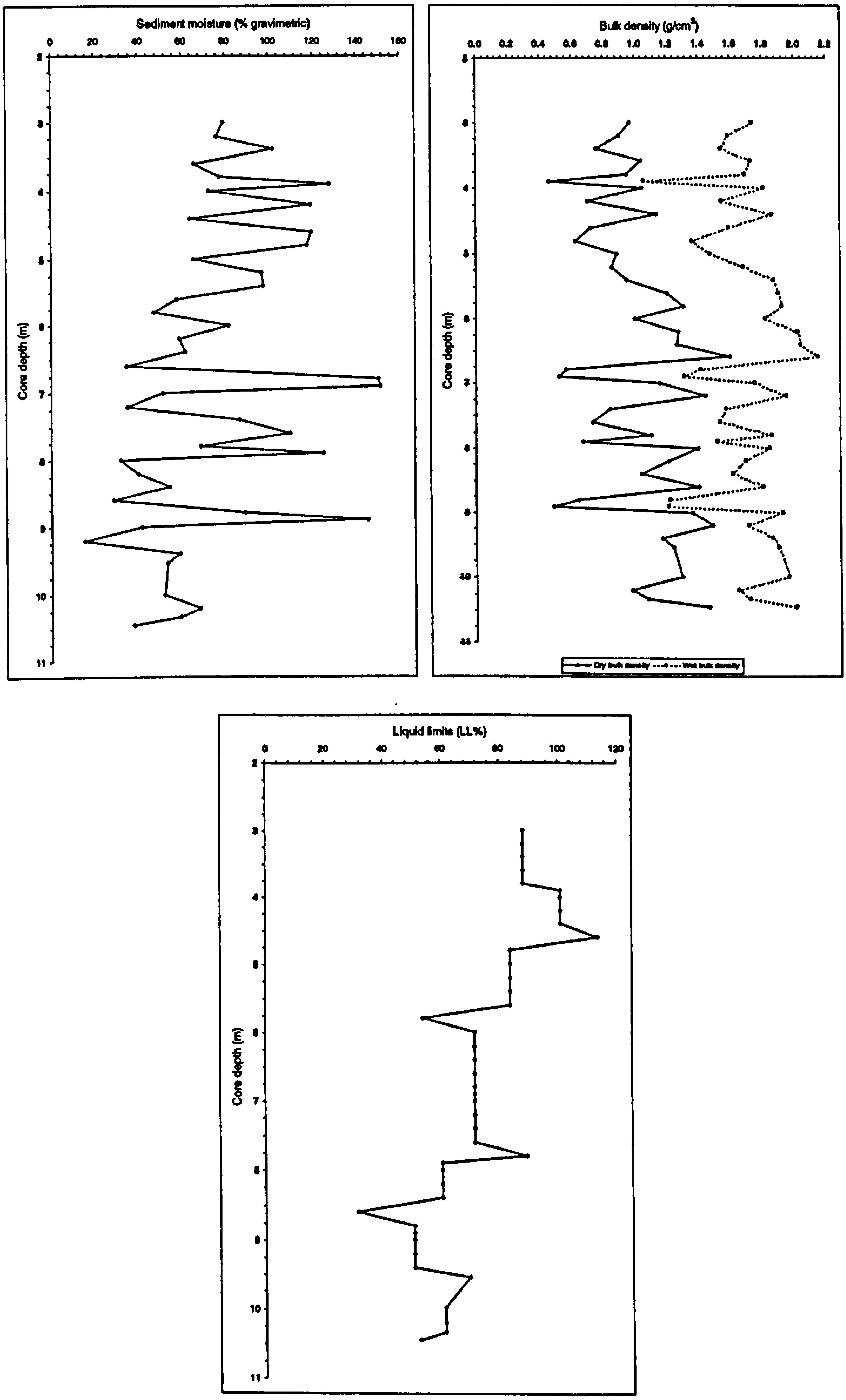


Figure 7.18. Input geotechnical data from core BS-97-3. 42 layers sampled between 0.1 and 0.2 m thick with the majority set at 0.20 m. See Figure 5.21 caption for explanation.

Additional data (Figure 7.19)

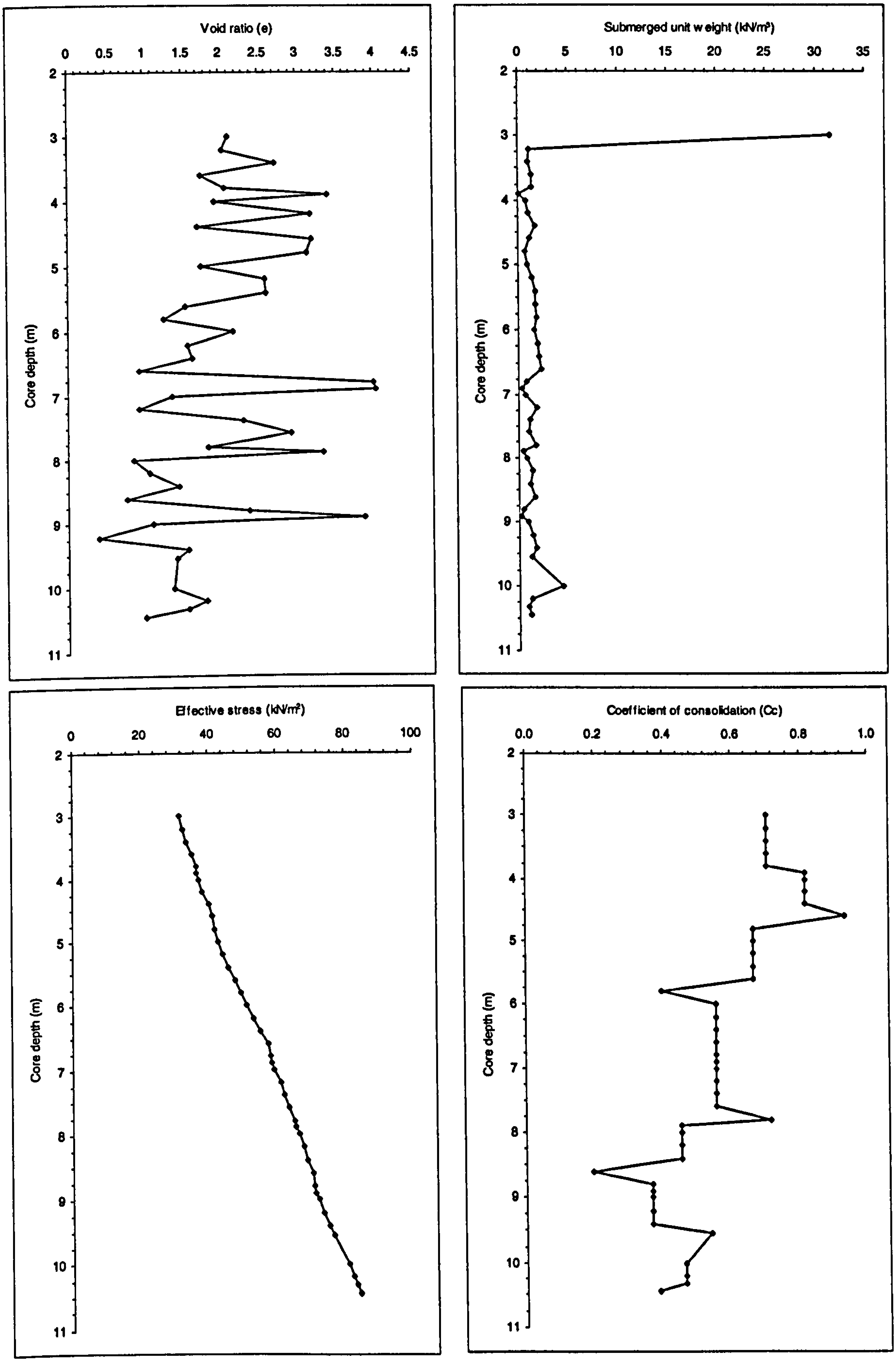


Figure 7.19. Additional geotechnical data from core BS-97-3. See Figure 5.22 caption for explanation.

Void ratio peaks in minerogenic-peat (4.1 at 6.9 m) and peat (3.4 at 3.9 m) where moisture is high, and is lower in sandy gravel (0.4 at 9.20 m). Submerged unit weight is greater in minerogenic (4.36 kN/m^3 at 10 m) than peat layers (0.06 kN/m^3 at 3.9 m) and the value at 3 m (31.55 kN/m^3) is the cumulative incremental submerged unit weight extrapolated for the uppermost 3 metres. Effective stress increases with depth and the inflection identified in core BS-97-2 (Figure 7.16) is not as pronounced here due to the lack of a thick facies of pure minerogenic-fines (Figure 7.6). Post-depositional compression (Cc) is greater in minerogenic-peat and peat (~0.9 at 4.60 m) as this sediment undergoes greater compression than fine minerogenic sediment and gravel (0.2 at 8.60 m).

Total geotechnical correction for core BS-97-3 (Figure 7.20)

The vertical displacement of basal sediment is quite small, e.g., 4 cm at 10.33 m core depth. This increases to 58 cm at 6.80 m core depth, where layers of peat and minerogenic-peat have compressed to varying degrees under the weight of the overlying minerogenic-fines and gravel infill (Figure 7.6). Values remain around 58 cm in the minerogenic-fines between 6.8 and 5.6 m, and increase in the organic-rich minerogenic sub-surface sediments reaching a maximum of 74 cm at 3.80 m core depth (see section 7.5.3). Table 7.4 contains the total correction for SLIPs obtained from core BS-97-3.

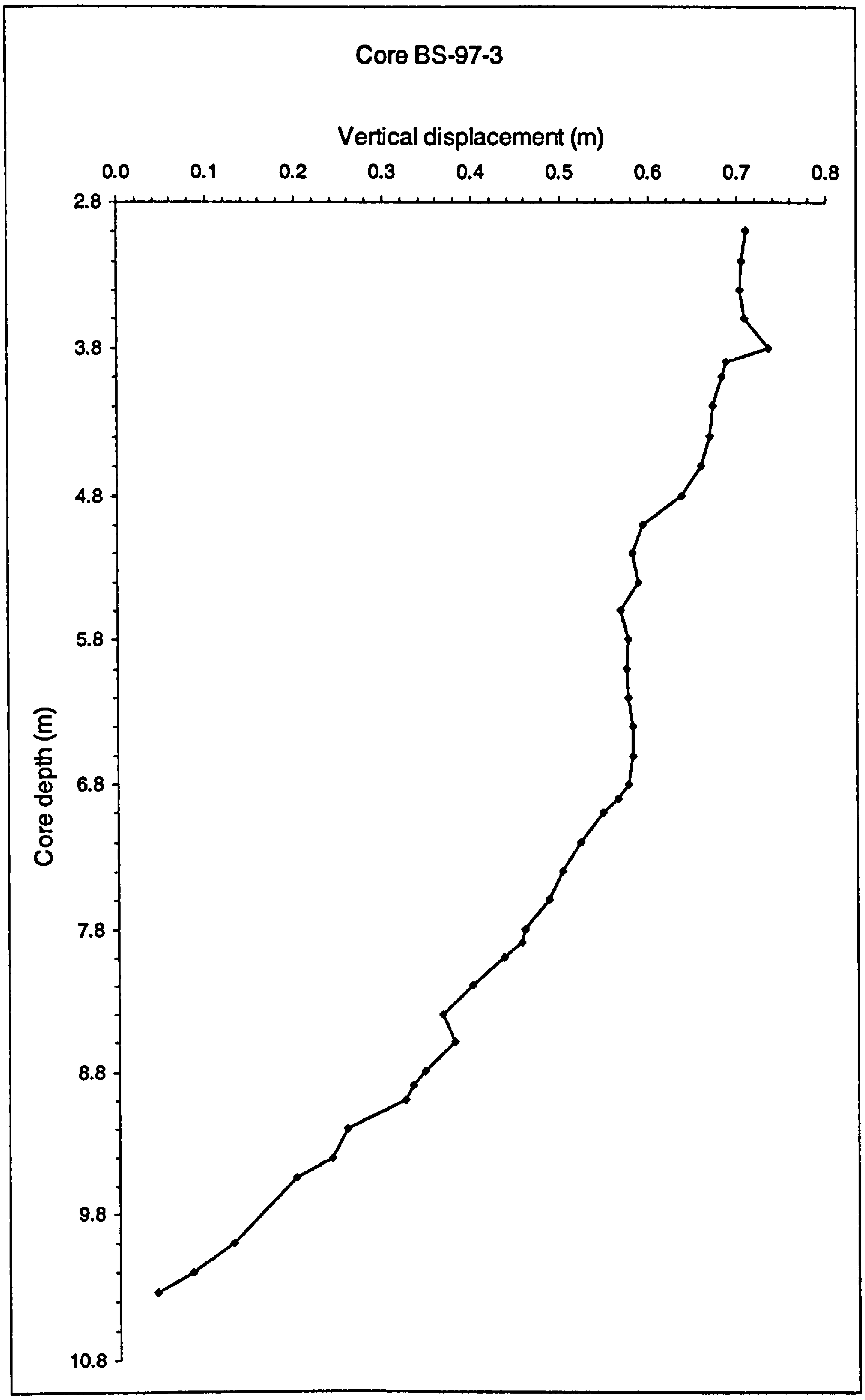


Figure 7.20. Total geotechnical correction for core BS-97-3 (ΔH in Figure 2.5). See Figure 5.23 caption for explanation.

7.5.3 Summary and limitations of geotechnical correction

Results from Blackpool Sands cores reinforce the argument that sea-level researchers routinely underestimate the effects of autocompaction on the altitude of non-basal SLIPs.

Post-depositional consolidation of sediment can be >0.7 m for minerogenic sequences (Figure 7.20) and 2 m for peat (core NS-97-2, Figure 5.23). Compaction of minerogenic layers is greater here than at North Sands. This is probably because Blackpool Sands sediments have been influenced by human modification, i.e., infill of the parking area. Although the majority of liquid limits values in Blackpool Sands cores are below 100%, high values do occur, e.g., >200% in core BS-97-2 (Figure 7.15) and are outside the usual range for the coefficient of consolidation index (Skempton, 1944). Core loss and surface sediment not retrieved results in a lack of input data and the model cannot always extrapolate values. Point estimates of groundwater level do not quantify trends over time and if the water table has often been deeper than current estimates effective stresses and compressions will be significantly greater. These observations add further weight to the initial argument (section 5.6.3) concerning the underestimation of autocompaction, as corrections are likely to be too low for specific layers. The results shown are therefore considered a minimum correction for Blackpool Sands cores.

7.6 Chronology

Blackpool Sands chronology is determined from eleven age measurements by AMS ^{14}C dating the basal and in-core sediment of cores BS-97-1, 2 and 3 (Table 7.4 and Figure 7.21). Basal SLIPs generated from cores BS-97-2 and 3 date the onset and cessation of marine conditions at Blackpool Sands *ca.* 6800 to 5600 and *ca.* 5300 to 4400 cal years BP respectively (Figures 7.13, 7.14 and 7.21). SLIPs acquired from core BS-97-1 date these events at *ca.* 5300 to 5000 and *ca.* 4800 to 4500 cal years BP respectively (Figures 7.12 and 7.21). These ages coincide closely with events from cores 2 and 3 but the altitude of SLIPs from the lowermost 3 metres of sediment appear anomalous (Figure 7.21). The basal SLIP is over 2.5 m below but 500 years younger than that of core BS-97-2. SLIP 22 is almost 1.4 m below SLIP 23 but the ages of the two wood fragments are identical (4410 ± 40 ^{14}C years BP, Table 7.4). There is no strong reason to suspect that the ages of SLIPs 21 to 25 are "contaminated" as they are all in stratigraphic order (Figure 7.22). An alternative explanation is rapid infill and unreliable indicative meanings from the foraminifera. It seems unlikely that all ages are systematically offset by the same amount of "young" carbon. However, with rapid infill of a channel more calcareous foraminifera, or test linings, would be expected in the facies. At the time of writing, a check analysis of SLIPs 22 and 23 (numbered SLIPs 32 and 33 respectively) was completed by NERC and results are reported in Table 7.4. SLIP 32 is the same age as originally reported (within 1σ) and SLIP 33 is older, although the results overlap at 2σ (Figures 7.21 and 7.22).

Index number	21	22	23	24
Radiocarbon laboratory number	CAMS-75520	CAMS-75521	CAMS-75522	CAMS-75523
Core	BS-97-1	BS-97-1	BS-97-1	BS-97-1
Material	Wood fragment (twig)	Wood fragment (twig)	Wood fragment (twig)	Wood fragment (twig)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	57.01 ± 0.26	57.75 ± 0.26	57.78 ± 0.26	58.80 ± 0.23
^{14}C age (years BP $\pm 1\sigma$)	4510 ± 40	4410 ± 40	4410 ± 40	4270 ± 40
Calibrated BP $\pm 1\sigma$ age ranges	(5298-5207) (5192-5048)	(5045-4872)	(5045-4872)	(4856-4829)
Calibrated BP $\pm 2\sigma$ age ranges	(5310-5031) (5015-4980)	(5274-5273) (5261-5184) (5121-5109) (5066-5058) (5053-4864)	(5274-5273) (5261-5184) (5121-5109) (5066-5058) (5053-4864)	(4869-4820) (4751-4732) (4716-4713) (4664-4664)
Median calibrated age (cal yrs BP)	5135	5020	5020	4837
Carbon content (% by weight)	52	50	54	54
$\delta^{13}C \pm 0.1$ (‰)	-29.90	-29.20	-26.6	-28.30
MTL sample height (m)	-8.53	-7.34	-5.95	-4.33
Miliammina fusca	28	4	4	0
Trochammina inflata	116	6	14	0
Jadammina macrescens	1308	152	8	4
Indicative meaning (m MTL)	2.25 ± 0.29	2.21 ± 0.29	2.52 ± 0.29	2.20 ± 0.29
Autocompaction (m)	0	0	0	0
SLIP (m MTL)	-10.78 ± 0.29	-9.55 ± 0.29	-8.47 ± 0.29	-6.53 ± 0.29

Table 7.4. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 7.21, 7.22 and 7.23. See Table 7.2 for MTL datum.

Index number	25	26	27	28
Radiocarbon laboratory number	CAMS-75524	CAMS-75525	CAMS-75526	CAMS-75527
Core	BS-97-1	BS-97-2	BS-97-2	BS-97-3
Material	Bulk sediment (minerogenic)	Wood fragment (twig)	Bulk sediment (minerogenic)	Wood fragment (twig)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	59.74 ± 0.27	54.06 ± 0.24	57.71 ± 0.23	48.08 ± 0.22
^{14}C age (years BP $\pm 1\sigma$)	4140 ± 40	4940 ± 40	4420 ± 40	5880 ± 40
Calibrated BP $\pm 1\sigma$ age ranges	(4816-4754) (4730-4717) (4712-4665) (4660-4605) (4604-4570) (4556-4551)	(5713-5667) (5663-5609)	(5204-5203) (5047-4960) (4932-4874)	(6746-6661)
Calibrated BP $\pm 2\sigma$ age ranges	(4828-4526)	(5834-5834) (5744-5595)	(5278-5170) (5127-5106) (5070-4866)	(6795-6769) (6759-6636) (6581-6571)
Median calibrated age (cal yrs BP)	4644	5655	5017	6700
Carbon content (% by weight)	6.8	52	2.8	46
$\delta^{13}C \pm 0.1$ (‰)	-28.80	-28.10	-28.60	-29.10
MTL sample height (m)	-3.34	-5.85	-5.11	-6.26
Miliammina fusca	0	24	0	2
Trochammina inflata	0	48	2	2
Trochammina ochracea	0	0	0	4
Jadammina macrescens	212	184	2	18
Polysaccammina ipohalina	0	2	0	0
Haynesina germanica	0	0	0	18
Ammonia beccarii	0	0	0	10
Elphidium williamsoni	0	0	0	2
Indicative meaning (m MTL)	2.20 ± 0.29	2.36 ± 0.29	2.46 ± 0.29	0.58 ± 0.29
Autocompaction (m)	0	0.02	0.21	0
SLIP (m MTL)	-5.54 ± 0.29	-8.19 ± 0.29	-7.36 ± 0.29	-6.84 ± 0.29

Table 7.4 continued. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 7.21, 7.22 and 7.23. See Table 7.2 for MTL datum.

Index number	29	30	31
Radiocarbon laboratory number	CAMS-75528	CAMS-75529	CAMS-75530
Core	BS-97-3	BS-97-3	BS-97-3
Material	Bulk sediment (minerogenic)	Wood fragment (twig)	Bulk sediment (minerogenic)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	48.25 ± 0.22	57.69 ± 0.22	60.65 ± 0.24
^{14}C age (years BP $\pm 1\sigma$)	5850 ± 40	4420 ± 40	4020 ± 40
Calibrated BP $\pm 1\sigma$ age ranges	(6723-6697) (6680-6638)	(5204-5203) (5047-4960) (4932-4874)	(4565-4563) (4527-4420)
Calibrated BP $\pm 2\sigma$ age ranges	(6776-6773) (6754-6549)	(5278-5170) (5127-5106) (5070-4866)	(4775-4773) (4607-4601) (4571-4555) (4553-4412)
Median calibrated age (cal yrs BP)	6712	5017	4471
Carbon content (% by weight)	6.2	55	4.1
$\delta^{13}C \pm 0.1$ (‰)	-26.80	-28.70	-29.00
MTL sample height (m)	-6.00	-4.55	-3.45
Miliammina fusca	0	2	0
Trochammina inflata	4	2	2
Haplophragmoides spp.	0	2	0
Jadammina macrescens	94	6	62
Indicative meaning (m MTL)	2.23 ± 0.29	2.16 ± 0.29	2.22 ± 0.29
Autocompaction (m)	0.08	0.35	0.49
SLIP (m MTL)	-8.15 ± 0.29	-6.36 ± 0.29	-5.18 ± 0.29

Table 7.4 continued. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 7.21, 7.22 and 7.23. See Table 7.2 for MTL datum.

Index number	32	33
Radiocarbon laboratory number	AA-54118	AA-54119
Core	BS-97-1	BS-97-1
Material	Wood fragment (twig)	Wood fragment (twig)
^{14}C Enrichment (% Modern $\pm 1\sigma$)	57.55 ± 0.30	56.72 ± 0.30
^{14}C age (years BP $\pm 1\sigma$)	4438 ± 42	4556 ± 43
Calibrated BP $\pm 1\sigma$ age ranges	(5212-5188) (5117-5112) (5051-4968)	(5312-5281) (5162-5136) (5105-5072)
Calibrated BP $\pm 2\sigma$ age ranges	(5286-5157) (5143-5099) (5089-4868)	(5441-5418) (5322-5207) (5193-5047)
Median calibrated age (cal yrs BP)	5008	5299
Carbon content (% by weight)	44	53
δ^{13}C $\pm$ 0.1 (‰)	-29.0	-26.6
MTL sample height (m)	-7.34	-5.95
Miliammina fusca	4	4
Trochammina inflata	6	14
Jadammina macrescens	152	8
Indicative meaning (m MTL)	2.21 ± 0.29	2.52 ± 0.29
Autocompaction (m)	0	0
SLIP (m MTL)	-9.55 ± 0.29	-8.47 ± 0.29

Table 7.4 continued. Radiocarbon dates and associated foraminiferal counts. Index numbers correspond with Figures 7.21, 7.22 and 7.23. See Table 7.2 for MTL datum.

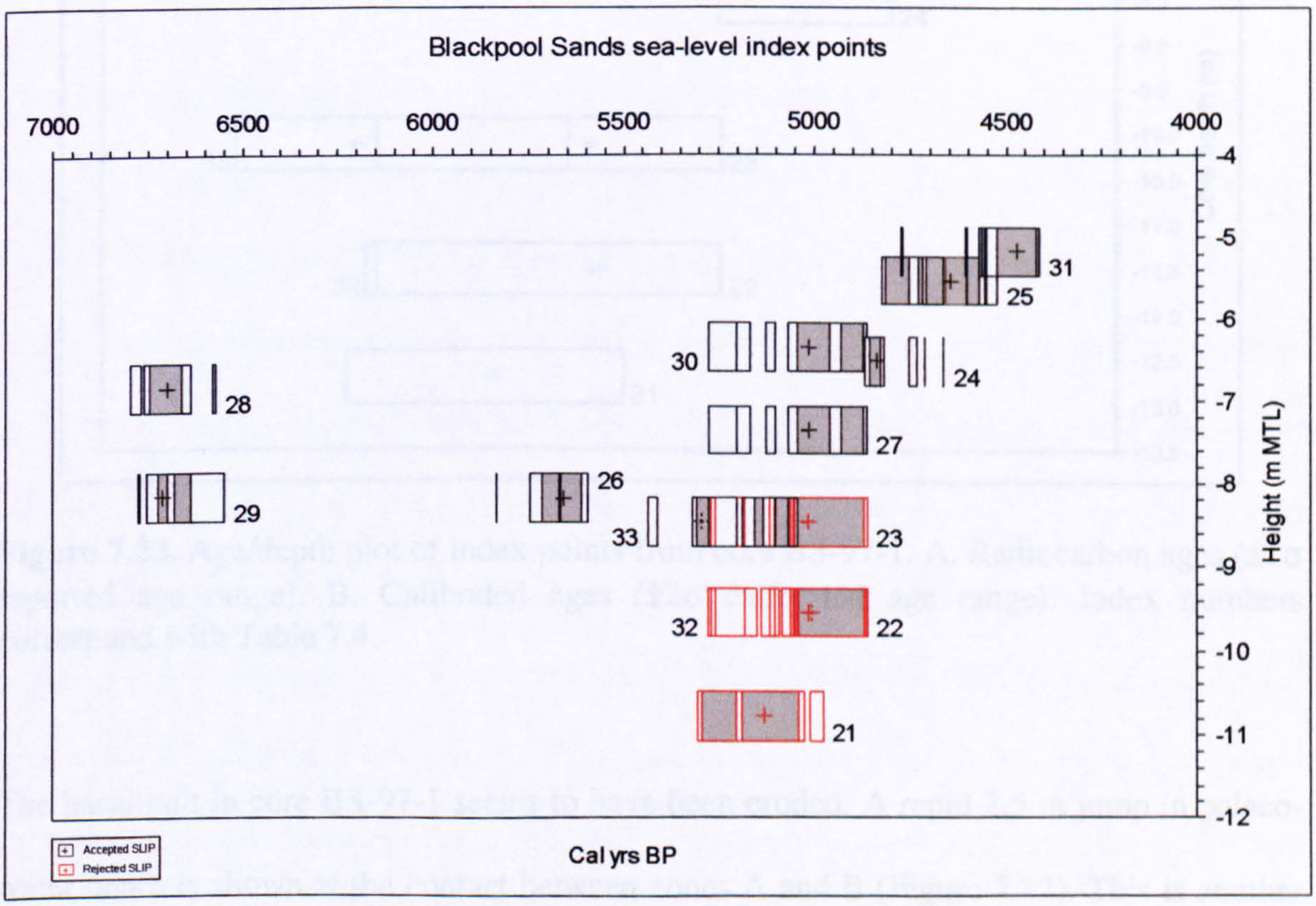


Figure 7.21. Blackpool Sands SLIPs. Index numbers correspond with Table 7.4. SLIPs: 21-25, 32 and 33 (core BS-97-1), 26 and 27 (core BS-97-2) and 28-31 (core BS-97-3). Median ages of SLIPs 32 and 33 are depicted by dashed lines.

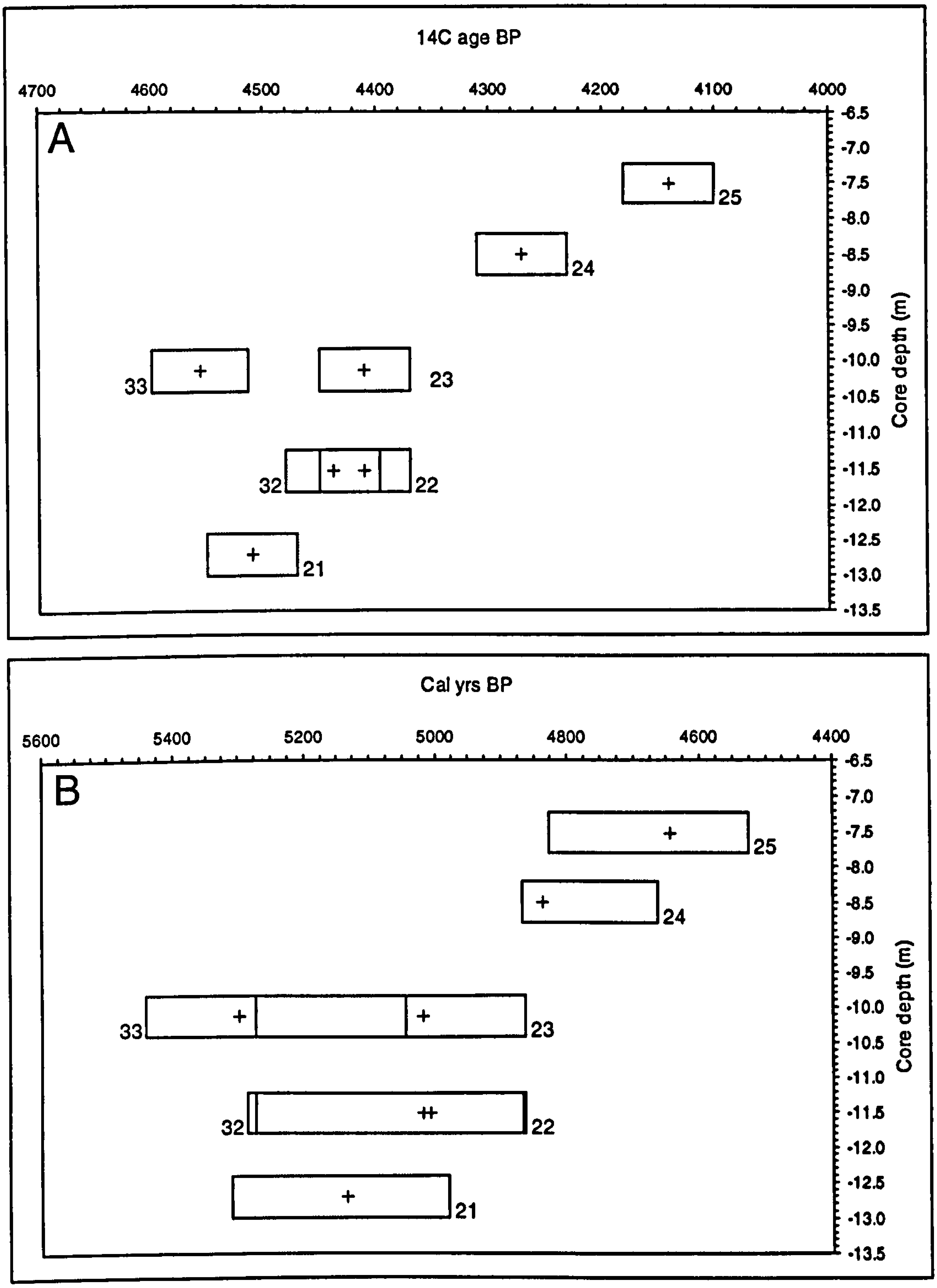


Figure 7.22. Age/depth plot of index points from core BS-97-1. A. Radiocarbon ages (±1σ reported age range). B. Calibrated ages (±2σ calibrated age range). Index numbers correspond with Table 7.4.

The basal unit in core BS-97-1 seems to have been eroded. A rapid 2.5 m jump in palaeo-water depth is shown at the contact between zones A and B (Figure 7.12). This is another argument to suggest that core BS-97-1 represents channel fill. Gravel appears throughout and there are calcareous foraminifera at the base. The peat above this unit, containing SLIP 21, is unconformable and probably a displaced "block". Another jump, of 3 m, in palaeo-water depth occurs at 11.65 m core depth in zone C and a similar unconformity may be present at the top of zone B (12.50 to 12.30 m core depth) but the contact is missing. The lithology is predominantly minerogenic sediment with relict peat, wood fragments and gravel (Figure 7.4). Foraminiferal analyses identify the lower section of core BS-97-1 as a salt-marsh environment, however, accretion rates seem too quick (5 m accumulation in 500 years (using index points 21 to 25) – interrupted by 0.5 m of mudflat sediments) for a salt marsh. The interpretation of the biostratigraphical record is problematic here, i.e., very low numbers of agglutinated foraminifera are preserved in the core between –7 and –3 m MTL (Figure 7.12 and Table 7.4). If the dissolution of calcareous foraminifera has occurred it should be considered a possibility in other cores at this site. It could hinder the palaeoenvironmental interpretation of core facies (Scott and Medioli, 1978; Green *et al.*, 1992, 1993).

Concerns regarding the low numbers of salt-marsh foraminifera in the uppermost facies of core BS-97-1 (0 to 212 per 5 ml of sediment from 10.20 to 7.50 m core depth, Appendix 2a) may be misplaced. No rapid jumps are present in the palaeo-water record and samples are in chronostratigraphic order (Figures 7.4 and 7.12). Furthermore, sediment accretion rates above the possible displaced block seem reasonable for a minerogenic salt marsh. The median calibrated age of SLIP 33 (*ca.* 5300 cal years BP) is almost 300 years older than the original age estimate (Table 7.4). Thus, at least 2.6 metres of minerogenic sediment and organic-rich sandy-clayey-silt was deposited in over 650 years (using median calibrated ages of SLIPs 33 and 25, Figure 7.22b), a rate of approximately 3.99 mm yr^{-1} (almost 4 m/ka). French and Spencer (1993) recorded an arithmetic mean accretion rate of 3.9 mm yr^{-1} in a contemporary study of a tide-dominated salt marsh on Scolt Head Island in Norfolk, UK. This is in excess of the local rate of RSL, estimated at 2 mm yr^{-1}, and similar to rates of mid-Holocene RSL rise at Blackpool Sands (Figure 7.21). Long *et al.* (1999) suggest that French and Spencer's (1993) estimated range of 1 to 8 mm yr^{-1} can be explained by differences in elevation and hence the period of tidal immersion. The proximity of the salt marsh to tidal creeks also causes an increase in local sediment accretion (Long *et al.*, 1999). This is particularly important when considering estimates of mid-Holocene accretion rates on the Blackpool Sands back-barrier minerogenic salt marsh. Core BS-97-1 was not corrected for autocompaction and index points 21, 22, 23 and 32 are rejected due to the concerns outlined here. SLIPs 24, 25 and 33 are considered borderline inclusions in the RSL history of Blackpool Sands (Figures 7.21 and 7.22).

Given the uncertainties surrounding the palaeo-water depth of the facies above the basal peat in all three cores, it would be most objective only to accept the basal SLIPs (26, 28, 29) from cores BS-97-2 and 3. However, it could be argued that SLIPs 27, 30 and 31 do not suffer the same problems as core BS-97-1 because the facies in cores 2 and 3 are quite different. Whilst there is evidence of disruptions in the basal 1.4 metres of core BS-97-1, the basal and overlying sequences in core BS-97-2 do not suffer from this problem (Figure 7.13). Furthermore, foraminifera in the basal peat of core BS-97-3 indicate a more gradual change in palaeo-water depth (Figure 7.14) and the contact is a conformable one. There is also no reason to suspect that dissolution has been a significant problem in Blackpool Sands cores. In core BS-97-1 no calcareous linings were found, core BS-97-2 contained only six linings and core BS-97-3 contained twenty linings of *A. beccarii*. However, this species is tolerant of a wide range of salinity (Murray, 1979) and was found more often than not intact in the cores alongside *H. germanica* and *E. williamsoni* (Figures 7.12 and 7.14). Core BS-97-1 also contains eight different calcareous species including some deeper water specimens, e.g., *A. mamilla*, *E. macellum* and *E. margaritaceum*, in the lower mudflat sequence (Figure 7.12). All SLIPs from cores BS-97-2 and 3 are accepted and were corrected for autocompaction (Figures 7.17, 7.20, 7.21 and Table 7.4).

The data show that mean sea level rose locally by ~1.1 – 3.6 m in 2400 to 1800 years respectively during the mid-Holocene, a rate of approximately 0.5 to 2 m/ka (cores BS-97-2 and 3, Figure 7.21). Sediment accumulation rates from cores BS-97-2 and BS-97-3 are estimated to be around 1 – 3 mm yr^{-1} and 1.5 – 2 mm yr^{-1} respectively and roughly 4 mm yr^{-1} in the uppermost minerogenic salt marsh in core BS-97-1. A comparison between SLIPs and model predictions indicates that sea-level rise since 6000 cal years BP is overestimated by the models, and all but Lambeck's (pers. comm. 1997) predictions agree closely with SLIPs *ca.* 6700 cal years BP (Figure 7.23).

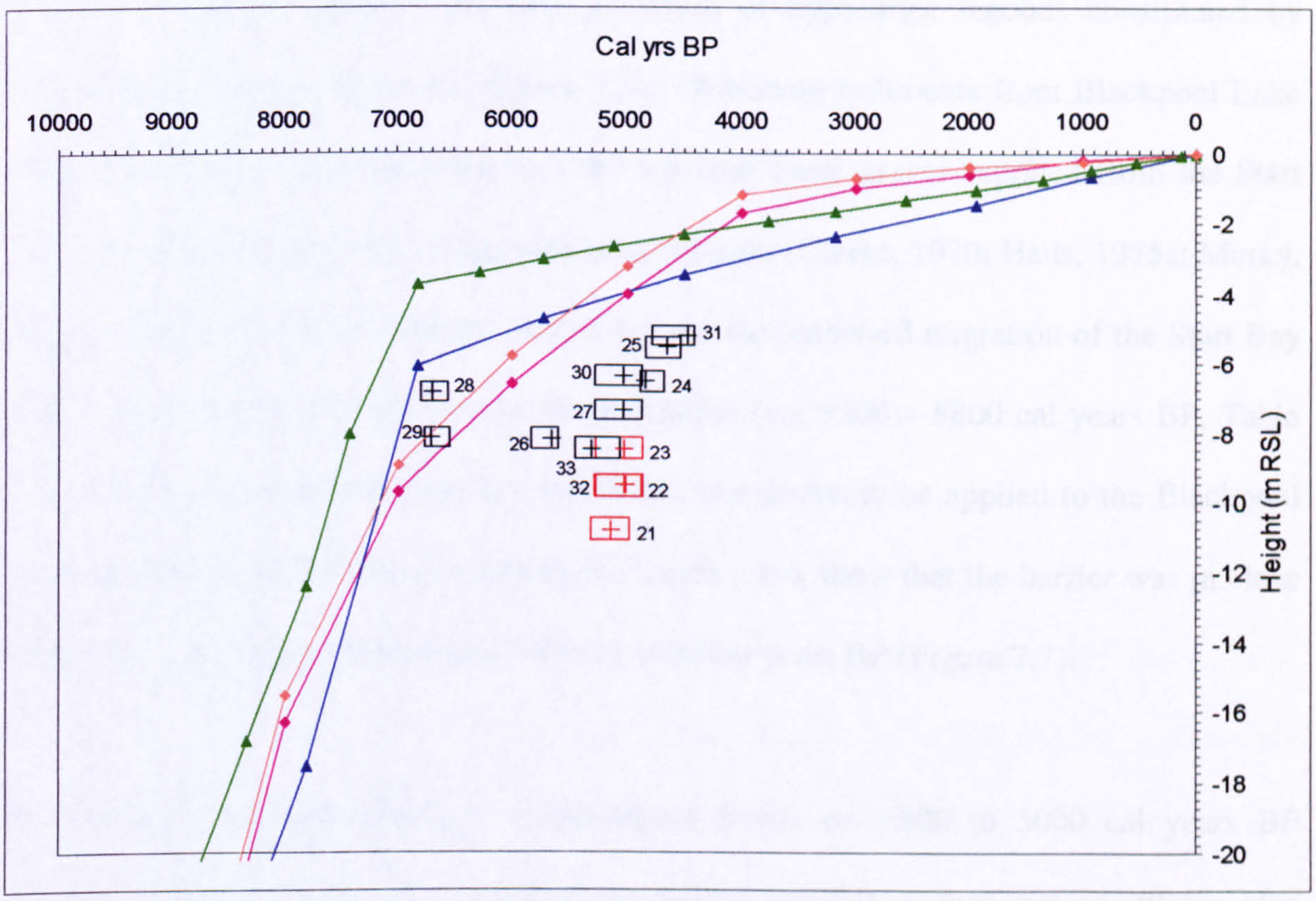


Figure 7.23. Blackpool Sands SLIPs and geophysical modelling results. Index numbers correspond with Table 7.4.

7.7 Palaeoenvironmental interpretation

In the early Holocene Blackpool Sands was probably a wooded valley bordering wetland drained by a river through igneous outcrops onto a sloping foreshore. This evolved on weathered Lower Devonian slates, hard grit and quartzite (Figure 7.3). The mouth of the river valley was basin shaped (Figure 7.8) and dipped seaward onto a rock lip running parallel to the shoreline (Figures 7.10 and 7.11). Alluvium mixed with locally derived gravel, and sand and plant material from marshland filled the basin to the lower valley slopes. The supratidal environment was then inundated by a barrier-lagoon complex migrating across Start Bay under rising seas (Figure 7.7, Hails, 1975a,b).

The earliest sign of a post-glacial marine transgression of Blackpool Sands is *ca.* 6800 cal years BP (Figure 7.21) but older marine deposits are likely to be found overlying and interbedded with basal gravel below the present day shingle barrier. Blackpool Sands early Holocene inter-tidal zone would have consisted of hyposaline lagoons constrained by barriers and fringing salt marsh (Figure 7.14). Waterlain sediments from Blackpool Lake would have mixed with lagoonal mud that covered basal gravel supplied from the Start Bay sub- and supratidal barrier, bay and bank deposits (Clarke, 1970; Hails, 1975a; Morey, 1983b). Morey's (1983b) estimate of 1 m yr^{-1} for the landward migration of the Start Bay barrier, based on Hail's (1975a) age determination (*ca.* 9300 – 8800 cal years BP, Table 6.1) of a peat-clay contact from Skerries Bank, can probably be applied to the Blackpool Sands coastline. Gravel facies in Blackpool Sands cores show that the barrier was in close proximity to its present position *ca.* 6800 to 4400 cal years BP (Figure 7.7).

Salt marsh conditions developed at Blackpool Sands *ca.* 6800 to 5000 cal years BP (Figures 7.12, 7.13 and 7.14). Shallow inter-tidal mudflat or lagoonal conditions also evolved on top of the weathered slates as local sea level rose at 1 to 2 m/ka (Figure 7.21). Sediment accretion kept pace with, and sometimes outpaced, sea level rise. The marine phase at Blackpool Sands was short-lived. Only two mudflat zones (0.08 m and 0.65 m of marine sediment *ca.* 5300 to 4800 cal years BP) are identified in core BS-97-1 (Figure 7.12) and one in core BS-97-3 (0.1 m of marine sediment *ca.* 6800 to 6500 cal years BP, Figure 7.14). However, problems identified with the position of index points 21, 22, 23 and

32 cast doubts on the history and geomorphological origins of core BS-97-1 (section 7.6). Changes in barrier morphology, creek infill and the rate and amount of offshore sediment material entering the basin *ca.* 6800 to 4400 cal years BP are possible causes for the sequence. Sediment accretion rates calculated from the basal 1.4 metres of core BS-97-1 are much faster (20 mm yr^{-1}) than cores BS-97-2 and 3 (1 to 3 mm yr^{-1}) and comparable salt-marsh facies discussed in the literature (Haslett *et al.*, 1998; Long *et al.*, 1999; Cahoon *et al.*, 2000). However, accretion rates estimated from the uppermost sediments of core BS-97-1 (~4 mm yr^{-1}) seem acceptable and indicate the likely proximity of the marsh facies to a creek network (French and Spencer, 1993). Furthermore, results are at the lower end of estimated ranges of marsh sediment accretion rates discussed elsewhere (Oenema and DeLaune, 1988; Reed, 1988). In addition to the hypotheses discussed in section 7.6 a tentative hypothesis proposed here is that the basal sequence of core BS-97-1 slumped down a creek wall as the fringing marsh fractured under external stresses due to changing barrier morphology. This would account for the RSL positions of the lowermost index points from core BS-97-1 (Figures 7.21 and 7.22). For this explanation it would not be necessary to consider the possible dissolution of calcareous foraminifera in the sequence.

Barrier closure occurred at Blackpool Sands *ca.* 4900 to 4400 cal years BP after at least 2400 years of inter-tidal conditions (Figures 7.12, 7.13 and 7.14). The timing of this event is similar to North Sands (Table 5.3) and Beesands (Morey, 1983b, Table 6.1), but earlier than closure at a much larger system on the Slapton Sands coastline (*ca.* 3200 to 2850 cal years BP, Morey, 1976, 1983b, Table 6.1). Trees, shrubs and peat covered the site down to the shoreline and the gravel barrier inundated this environment *ca.* 2800 to 2350 cal years BP as its volume increased under slowly rising seas and local storm events (Pengelly, 1869; Morey, 1974, 1976). The submerged forest at Blackpool Sands has been revealed on a number of occasions during the last 200 years (Pengelly, 1869) and recent storm events have continued to breach the barrier (Blackpool Sands Visitor Site, 2002).

7.8 Summary

Prior to this study no other Holocene SLIPs have been generated specifically to date the onset of post-glacial sea-level rise at Blackpool Sands. The submerged forest and peat bed is dated *ca.* 2800 to 2350 cal years BP (Morey, 1974, Table 7.1) but this information is not included in the Holocene SLIP database (2003) for south-west Britain. The lithostratigraphy suggests that inter-tidal barriers migrated shoreward on top of weathered Lower Devonian slate and alluvium during the early Holocene marine transgression. Foraminiferal analysis shows that basal salt-marsh peat developed *ca.* 6800 cal years BP and very thin facies of mudflat sediment were deposited until *ca.* 4800 cal years BP. The rate of local sea-level rise, *ca.* 6800 to 5000 BP, is around 2.5 m/ka. Geophysical models, particularly Peltier (pers. comm. 2000, 2002) and Lambeck (pers. comm. 2000), agree very closely with SLIPs 28 and 29. The remaining index points are overestimated by about 2 to 5 metres. Around 5000 cal years BP Blackpool Sands valley became choked with fine minerogenic-peat and coarse gravel as sea level rise slowed to ~1.5 m/ka until barrier closure occurred *ca.* 4400 cal years BP (SLIP 31). This initiated the start of a back-barrier fresh-water minerogenic-peat marsh system. The present-day back-barrier environment has only the beach to protect itself from further sea-level rise and storm surges.

Chapter 8

Discussion

8.1 Holocene relative sea-level change along the south coast of Devon

The primary aim of this thesis is to establish a record of early to mid-Holocene relative sea-level change for the south Devon coast. Thirty-three Holocene sea-level index points have been generated of which twenty-four are accepted and nine are rejected (Figure 8.1). Eleven of the accepted SLIPs are borderline inclusions. These categories are explained in sections 4.5, 5.7, 6.5 and 7.6. The accepted SLIPs enhance the current Holocene SLIP database for the south-west peninsula and southern Britain (Figure 1.4 and Table 1.1) by 23% and increase south Devon's database almost fivefold (Figure 8.2). The Holocene sea-level history recovered from cores in this project shows a RSL rise of around 13 metres from *ca.* 9000 to 4400 cal years BP (Figure 8.1). Early Holocene basal SLIPs from Bantham Sands, North Sands and Slapton Sands contribute to filling the gap in the data below –14.5 m RSL *ca.* 9500 to 8500 cal years BP (Figure 8.2). Mid-Holocene SLIPs from Blackpool Sands show that the vertical spread of data from south Devon *ca.* 7000 to 4000 cal years BP is ~3.9 m, around half that of the database for south-west Britain (see section 1.4.1). No SLIPs younger than 4000 cal years BP were generated from cores in this project but Bantham Sands and Slapton Sands (see Morey, 1976, 1983b) cores may offer the potential for younger Holocene SLIPs.

8.1.1 The early Holocene RSL history of south Devon

The early Holocene marine transgression of south Devon occurred between 9000 and 7000 cal years BP when sea level was rising at ~5.4 ± 2.1 m/ka (Table 8.1). Rates of RSL rise around south-west Britain based on data available prior to this study are 6 to 7 m/ka *ca.* 10,000 to 7000 cal years BP (Table 1.2) and agree closely with those calculated from the new data. Similar timings for the early Holocene marine transgression are shown by sea-level curves from the south-east coast of Devon (Figure 1.6d, Morey, 1976) and the English Channel (Figure 1.7b, Heyworth and Kidson, 1982). However, both curves are based on only two samples dated between 8600 and 8100 ^{14}C years BP that are rejected by the Sea-Level Research Unit at Durham (Environmental Research Centre, 2003).

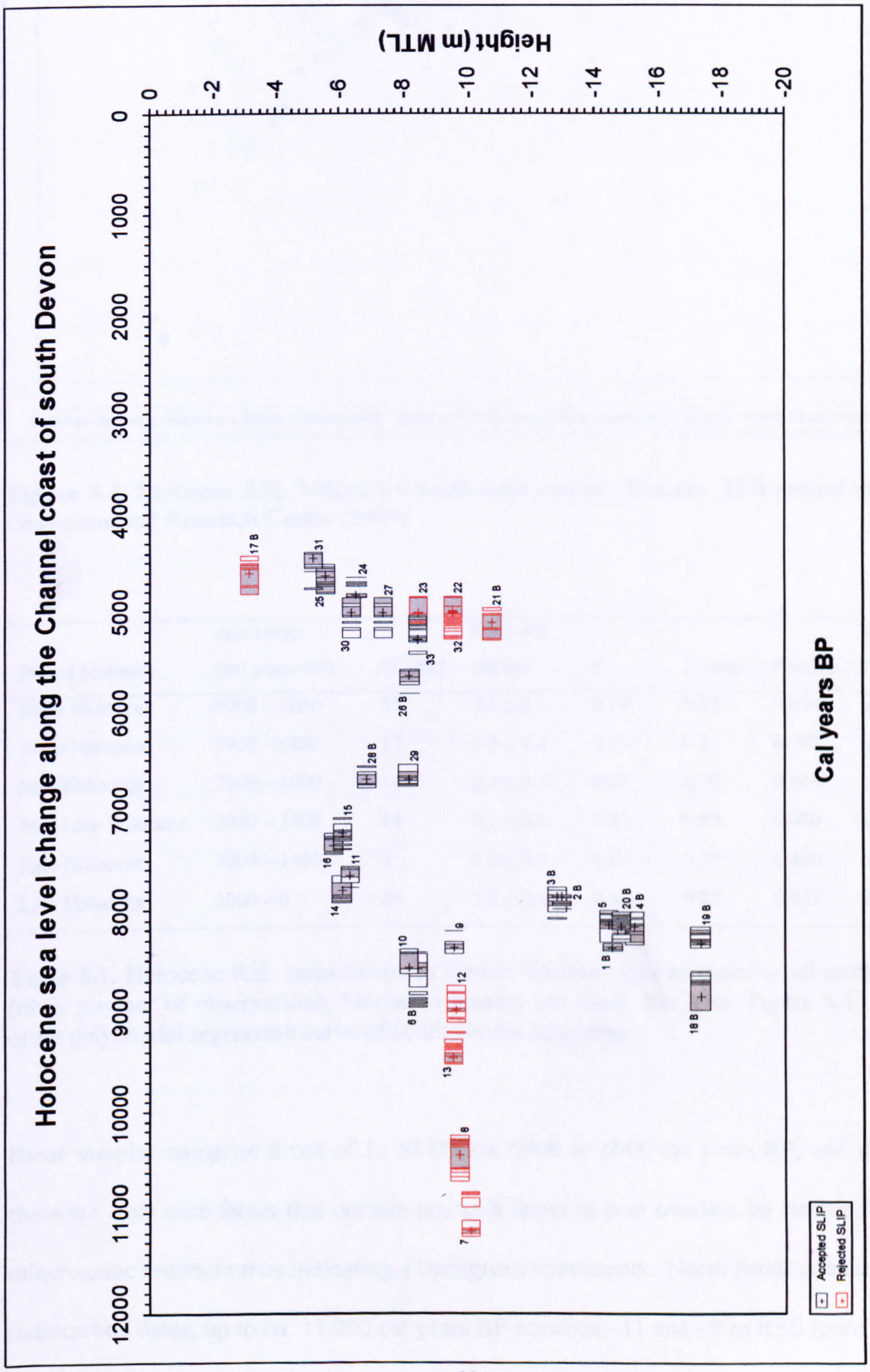


Figure 8.1. Holocene RSL history for south Devon. Corrected heights are shown for SLIPs 6 to 11 (see Table 5.5). Index numbers correspond with Tables 4.4, 5.3, 6.4 and 7.4. B = Basal SLIP.

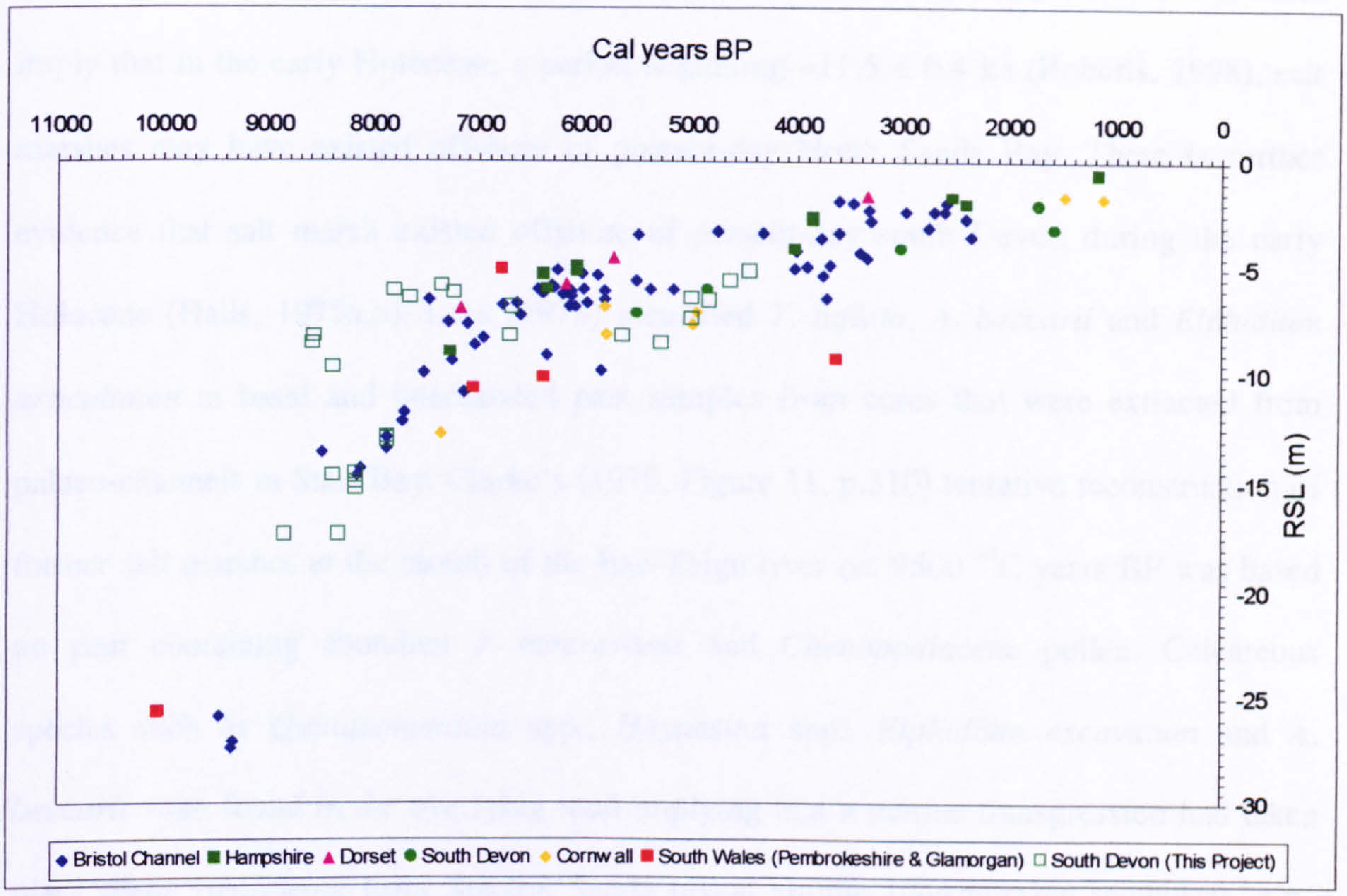


Figure 8.2. Holocene RSL history for south-west Britain. Sources: This project and The Environmental Research Centre (2003).

Period (dataset)	Age range (cal years BP)	N (obs)	RSL rate (m/ka)	r^2	r^2 (adj)	P-value	Sig? (95%)
Early Holocene	9000 – 7000	15	5.4 ± 2.1	0.34	0.28	0.024	yes
Early Holocene	9000 – 6000	17	4.0 ± 1.4	0.36	0.31	0.012	yes
Mid-Holocene	7000 – 4000	11	0.9 ± 0.4	0.37	0.30	0.046	yes
Mid-Late Holocene	7000 – 1500	14	1.1 ± 0.2	0.82	0.80	0.000	yes
Late Holocene	3000 – 1500	3	1.0 ± 0.7	0.62	0.25	0.420	no
Late Holocene	3000 – 0	4*	1.3 ± 0.3	0.90	0.85	0.052	no

Table 8.1. Holocene RSL rates for south Devon. Median ages are used in all analyses. N (obs): number of observations; *Includes present sea level. See also: Figure 8.3b for 3rd order polynomial regression curve of south Devon RSL data.

Basal samples comprise 8 out of 10 SLIPs *ca.* 9000 to 6000 cal years BP, and many of these are from core facies that contain brackish fauna in peat overlain by marine fauna in minerogenic sediment thus indicating a transgressive sequence. North Sands contains older radiocarbon dates, up to *ca.* 11,200 cal years BP between –11 and –9 m RSL (core NS-97-

3, index points 7, 8, 12 and 13). These are rejected as SLIPs as being too old. The dates imply that in the early Holocene, a period beginning ≈11.5 ± 0.4 ka (Roberts, 1998), salt marshes may have existed offshore of present-day North Sands Bay. There is further evidence that salt marsh existed offshore of present-day south Devon during the early Holocene (Hails, 1975a,b). Lees (1975) identified *T. inflata*, *A. beccarii* and *Elphidium articulatum* in basal and intercalated peat samples from cores that were extracted from palaeo-channels in Start Bay. Clarke's (1970, Figure 11, p.310) tentative reconstruction of former salt marshes at the mouth of the Exe-Teign river *ca.* 9500 ^{14}C years BP was based on peat containing abundant *J. macrescens* and *Chenopodiaceae* pollen. Calcareous species such as *Quinqueloculina* spp., *Haynesina* spp., *Elphidium excavatum* and *A. beccarii* were found in the overlying mud implying that a marine transgression had taken place there. Sediments from Slapton Sands reveal similar transgressive sequences, e.g., salt-marsh foraminifera dominate the basal facies of core SS-99-1 (Figure 6.12) and are overlain by lagoonal species of the type found by Clarke (1970) and Lees (1975).

Bantham Sands, North Sands and Slapton Sands contain thick (6 – 12 m) sequences of lagoonal sediments and SLIPs from these cores indicate that the pace of the early Holocene marine transgression abated within 2000 years (*ca.* 9000 – 7000 cal years BP, Figure 8.1). The lithostratigraphy from Blackpool Sands contains evidence that marine conditions existed at the site *ca.* 6700 cal years BP around –7 m RSL, after which high marsh prevailed (core BS-97-3, Table 7.4) as the rate of RSL rise slowed significantly (Table 8.1). SLIPs 11, 14, 15 and 16 from core NS-97-3 (see section 5.7) position this event much earlier (*ca.* 7900 – 7100 cal years BP, Figure 8.1) and plot the maximum height (around –6 m RSL) proposed for early Holocene RSL in the region. It is interesting to note that the slowing down of early Holocene sea-level rise *ca.* 6800 cal years BP (Figure 8.1) is reflected in data from elsewhere around south-west Britain (Hawkins, 1971, 1979; Kidson and Heyworth, 1973, 1976; Morey, 1976; Heyworth and Kidson, 1982; Healy, 1993, 1995, 1996, Figure 8.2). However, records of Holocene RSL histories prior to *ca.* 6800 cal years BP are very limited in south Devon and Cornwall (Waller and Long, 2003) although similar barrier-dominated sedimentary sequences have been identified in response to rising RSL (e.g., Hails, 1975a,b; Healy, 1993, 1995). Sea-level data from the uppermost inter-tidal facies of core NS-97-3 may therefore provide new evidence for sedimentary responses to RSL rise along south Devon's coastline *ca.* 7000 cal years ago (Figure 8.3).

Borderline SLIPs 6, 9, 10, 11, 14, 15 and 16 from core NS-97-3 shape significantly the profile of the local early Holocene RSL history for south Devon (Figure 8.3a,b). SLIPs 6, 9 and 10 are *ca.* 1000 years older than SLIPs from the Bristol Channel and Hampshire, southern England (Figure 8.2). This comparison supports further the argument that minerogenic sediments from the entire lowermost facies in core NS-97-3 may contain older carbon from earlier offshore salt-marsh deposits (see section 5.7.2). SLIPs 11, 14, 15 and 16 plot at a similar height to SLIP No. 86 (Q672) from Westward Ho!, North Devon (Churchill, 1965a, Table 1.1) and appear to confirm the end of the early Holocene phase of rapid sea level rise and minerogenic sedimentation mooted by Hawkins (1971, 1979), Hails (1975a,b) and others in the region. This phase is followed by a decelerating rate of RSL and a long period of organic and organogenic sedimentation (Figures 4.6 and 5.8) of the type identified by Healy (1993, 1995) in sequences from the south coast of Cornwall.

Rates of early Holocene RSL rise at North Sands are as high as 17 mm yr^{-1} (section 5.7.3). Hawkins (1971) calculated RSL rates ~20 mm yr^{-1} during the Boreal transgression of south-west England. However, some late Weichselian and Boreal peat in the Bristol Channel do not conform to the proposed sea level pattern (Hawkins, 1971, Figure 3, p.122). Godwin and Willis (1964) state that peat was growing on old drift freshwater organic mud long before the sea intervened. Detritus mud in the Gordano Valley formed behind sand bars and was independent of sea level at that time (Jefferies *et al.*, 1968).

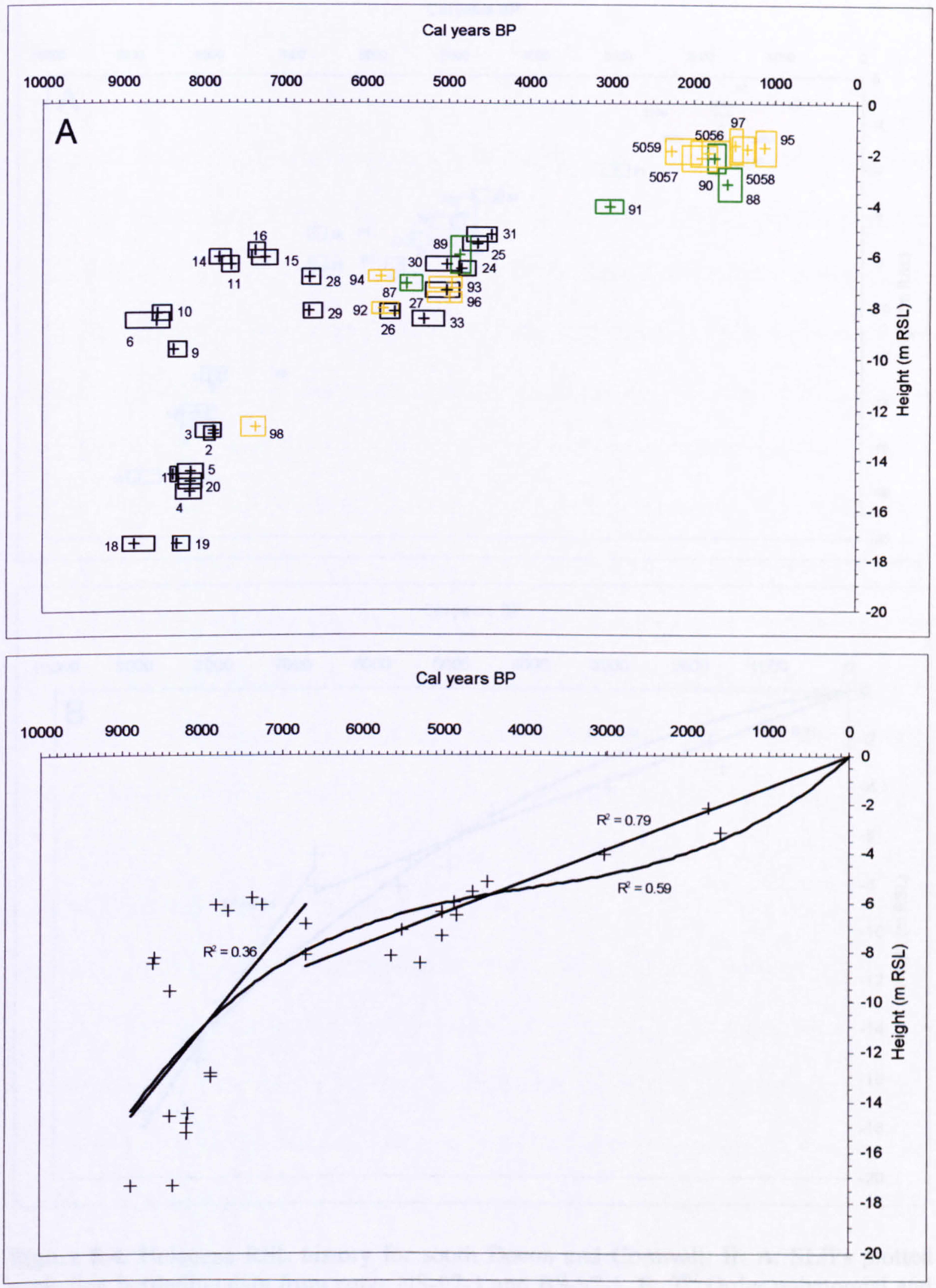


Figure 8.3. Holocene RSL history for south Devon and Cornwall: I. A. SLIPs. Colour key (see Figure 2.6): Black: This study (non-basal samples not compaction corrected include SLIPs 24, 25 and 33), Gold: Cornwall and Scilly Isles, Bright green: south Devon. Scilly Isles index numbers correspond with Table 8.3. B. 3rd Order polynomial regression performed on all data from south Devon. Linear regressions are on early-mid and mid-late Holocene RSL data. Sources: This project; Ratcliffe and Straker (1996); Environmental Research Centre (2003).

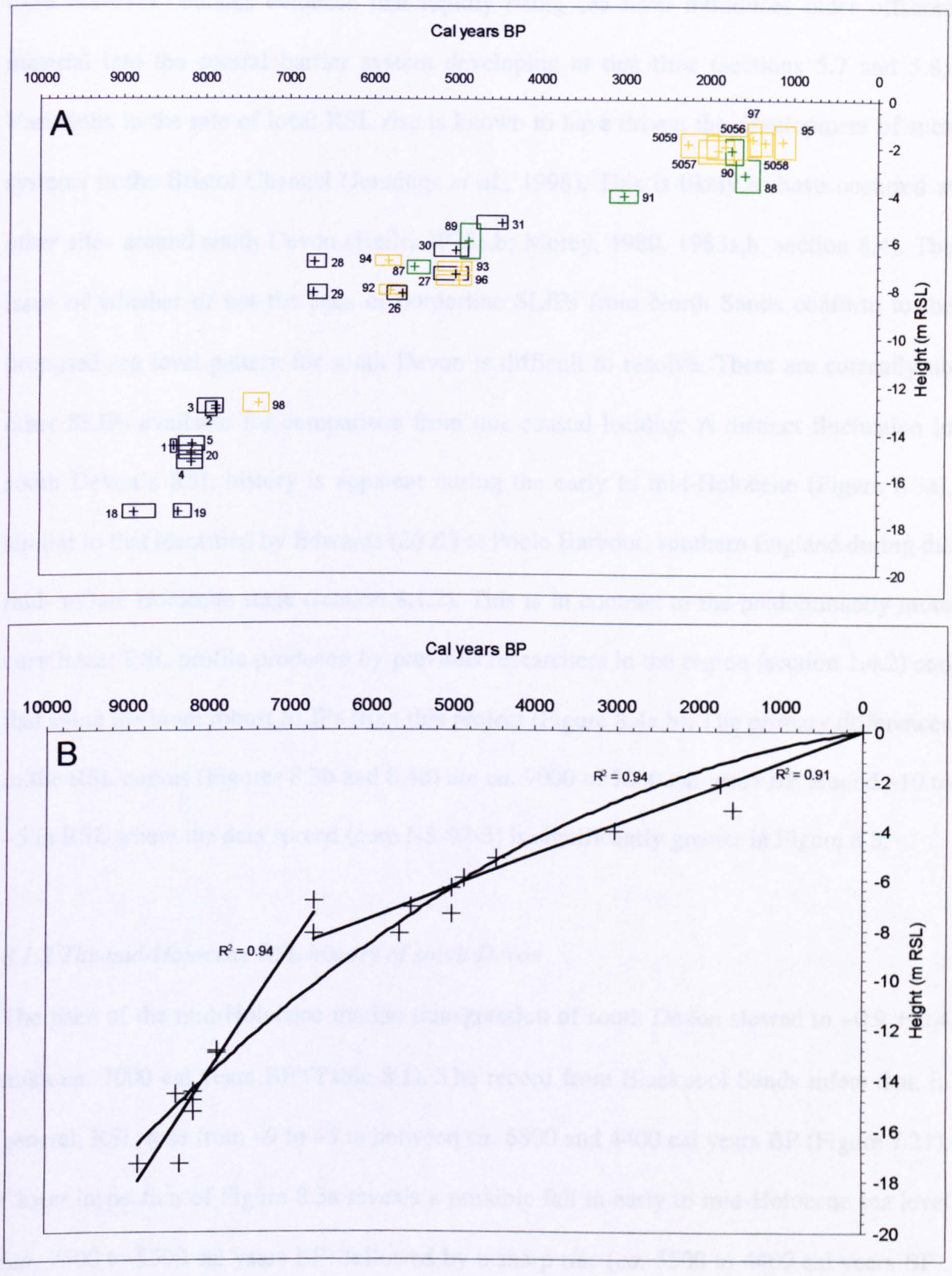


Figure 8.4. Holocene RSL history for south Devon and Cornwall: II. A. SLIPs plotted excluding borderline data from cores NS-97-3 and BS-97-1. B. 2^{nd} Order polynomial and linear regressions performed on all data from south Devon. See Figure 8.3 caption for explanation. RSL rates are around 4.6 ± 1.2 m/ka (r^2 (adj) = 0.66, p-value = 0.008) *ca.* 9000 to 7000 cal years BP and 4.8 ± 0.4 m/ka (r^2 (adj) = 0.93, p-value = 0.000) *ca.* 9000 to 6000 cal years BP. See Table 8.1 for comparisons with Figure 8.3b.

Core NS-97-3 contains evidence that rapidly rising sea level introduces older offshore material into the coastal barrier system developing at that time (sections 5.7 and 5.8). Variations in the rate of local RSL rise is known to have driven the development of such systems in the Bristol Channel (Jennings *et al.*, 1998). This is likely to have occurred at other sites around south Devon (Hails, 1975a,b; Morey, 1980, 1983a,b, section 8.4). The issue of whether or not the ages of borderline SLIPs from North Sands conform to the proposed sea level pattern for south Devon is difficult to resolve. There are currently no other SLIPs available for comparison from this coastal locality. A distinct fluctuation in south Devon's RSL history is apparent during the early to mid-Holocene (Figure 8.3a), similar to that identified by Edwards (2001) at Poole Harbour, southern England during the mid- to late Holocene stage (section 8.1.2). This is in contrast to the predominantly more curvilinear RSL profile produced by previous researchers in the region (section 1.4.2) and that using the more robust SLIPs from this project (Figure 8.4a,b). The primary differences in the RSL curves (Figures 8.3b and 8.4b) are *ca.* 9000 to 7000 cal years BP around –10 to –5 m RSL where the data spread (core NS-97-3) is significantly greater in Figure 8.3.

8.1.2 The mid-Holocene RSL history of south Devon

The pace of the mid-Holocene marine transgression of south Devon slowed to ~0.9 ± 0.4 m/ka *ca.* 7000 cal years BP (Table 8.1). The record from Blackpool Sands infers that, in general, RSL rose from –9 to –5 m between *ca.* 6800 and 4400 cal years BP (Figure 7.21). Closer inspection of Figure 8.3a reveals a possible fall in early to mid-Holocene sea level (*ca.* 7500 to 5500 cal years BP) followed by a sharp rise (*ca.* 5500 to 4400 cal years BP). However, the vertical spread of mid-Holocene data from authors such as Hawkins (1971, 1979), Kidson and Heyworth (1976) and Healy (1995) seem to indicate a smooth RSL envelope, comparable to that shown in Figure 8.4a,b, with no undulations. On the other hand, Heyworth and Kidson (1982) constructed a Holocene sea-level curve for the English Channel (curve B in Figure 1.7) similar to the RSL curve in Figure 8.3b. They suggest that any oscillations in their RSL histories are smaller than other uncertainties, e.g., vertical error estimates and the spread of SLIPs from the locality. Therefore, Figure 8.3a,b is likely to signify a decelerating rate of RSL *ca.* 7000 cal years BP followed by an increase during the late Holocene. Jennings *et al.* (1998) also predict a decelerating rate of RSL *ca.* 7500 to 5000 years BP followed by a late Holocene rise at Porlock, Somerset. They propose possible RSL falls during the Boreal era, when organic beds formed offshore in the Bristol Channel and Swansea Bay. Fluctuations in RSL would have provided optimum conditions for gravel barrier growth at a time when sediment supply to the system was plentiful.

Mid- to late Holocene RSL rates from this project (1.1 ± 0.2 m/ka, Table 8.1) compare favourably with published data from south-west Britain (1 m/ka, Table 1.2) enforcing the argument that index points 21, 22, 23 and 32 (core BS-97-1) are unreliable as SLIPs (section 7.6). The most likely explanations for the anomalous chronology of the lowermost salt-marsh facies in core BS-97-1 appears to be rapid infill of the creek or that the sedimentary sequence may have slumped into a former marine creek (sections 7.6 and 7.7). Mid-Holocene SLIPs from the literature (Hails, 1975b; Healy, 1995, Table 8.2) plot at similar RSL heights to those from cores BS-97-2 and 3 (Figures 8.3a and 8.4a) adding weight to the argument that SLIPs 24 to 31 and 33 are reliable contributors to mid-Holocene RSL histories in south Devon.

Heyworth and Kidson (1982) and Hawkins (1971, 1979) present regional sea-level curves constructed from SLIPs questionably linked to a reference tide level (Shennan, 1983; Haslett *et al.*, 1997). Samples from submerged forests, coastal trackways and ancient peat beds provide histories not synonymous with sea level but at some height above it (Haslett *et al.*, 1998). SLIPs from this project are carefully evaluated for possible errors associated with indicative meaning and autocompaction. Mid-Holocene sea-level data from the Environmental Research Centre (2003) have been similarly screened (Tables 1.1 and 8.2).

Protocols for the publication of sea-level data have improved in recent years but it is notable that palaeoenvironmental information, chronological and vertical error margins are sometimes not reported or published incompletely in the literature. Vertical error estimates applied to SLIPs from Cornwall and Devon include ±0.20 m for indicative meaning, ±0.05 m for coring compaction, ±0.05 m tidal correction, ±0.05 – 0.20 m survey (levelling) error and ±0.01 m vertical error per 1 cm sample thickness (Horton, pers. comm. 2003). No vertical correction is applied for autocompaction of non-basal sediments. Horizontal errors include the use of median ages and maximum and minimum (2σ) age ranges calibrated using the probability method B (95%) in CALIB 4.2 (Stuiver *et al.*, 1998a). Centres of crosses or boxes are plotted using the median age and height of the SLIP.

SLIP No.	Lab. No.	Location	Facies contact (sample location)	Basal or non-basal	Sea-level indicator	Vertical error	Reference
87	SRR165	Beesands barrier	Freshwater *Phragmites* peat below shingle	non-basal (upper contact)	Pollen	± 0.25	Hails (1975a,b)
88	SRR317	N. Hallsands barrier	Tree stump below washover gravel	non-basal (upper contact)	Pollen	± 0.65	Hails (1975a,b)
89	SRR164	Beesands barrier	Freshwater *Phragmites* peat below shingle	non-basal (upper contact)	Pollen	± 0.72	Hails (1975a,b)
90	SRR492	Slapton Ley (lake bed)	Fen peat below washover sand	non-basal (upper contact)	Pollen	± 0.55	Morey (1976)
91	SRR493	Slapton Ley (lake bed)	Mud/peat below fen (*Carex*) peat	non-basal (upper contact)	Pollen	± 0.23	Morey (1976)
92	Q2777	Marazion marsh	Wood peat below sand and shells	0.7 m above base	Diatoms and pollen	± 0.21	Healy (1995)
93	Q2780	Marazion marsh	Wood peat below sand and shells	2.2 m above base	Diatoms and pollen	± 0.21	Healy (1995)
94	Q2776	Marazion marsh	Wood peat below sand and shells	2.4 m above base	Diatoms and pollen	± 0.21	Healy (1995)
95	Q2775	Marazion marsh	Wood peat below fine sand	7.3 m above base	Diatoms and pollen	± 0.70	Healy (1995)
96	Q2774	Marazion marsh	Wood peat below sand and shells	1.3 m above base	Diatoms and pollen	± 0.21	Healy (1995)
97	Q2778	Marazion marsh	Wood peat below fine sand	5 m above base	Diatoms and pollen	± 0.70	Healy (1995)
98	Q2781	Trewornan, N. Cornwall	Minerogenic-peat below fine sand, silt, clay and shells	At least 0.5 m above base	Diatoms and pollen	± 0.40	Healy (1993)

Table 8.2. Palaeoenvironmental information associated with SLIPs from Cornwall and Devon. Other data (latitude/longitude, chronology, heights and associated error margins) and corresponding SLIP numbers can be found in Table 1.1. Source: Environmental Research Centre (2003).

Healy's (1995) mid-Holocene data from south Cornwall (SLIPs 92, 93, 94 and 96, Figures 8.3a and 8.4a) are predominantly from intercalated peat units and the south Devon (SLIPs 87 and 89, Hails, 1975b) data are from similar non-basal organic facies (Table 8.2). Vertical errors applied to these SLIPs do not include an estimate of autocompaction (Horton, pers. comm. 2003) and they may plot higher than their current positions. The magnitude of any geotechnical correction is difficult to predict without laboratory analyses but work carried out on minerogenic-peat sequences in this project resulted in 0.8 to 2.2 m of vertical displacement. However, without a geotechnical analysis and / or basal dates of similar age from the same cores it is only possible to speculate.

The vertical spread of data during the mid-Holocene in this project is half that of the national database for south-west Britain (Figure 8.2). However, this may be reduced further given the problems of chronology and stratigraphy outlined in a number of cores in this project, e.g., at North Sands and Blackpool Sands. Heyworth (1978) related stratigraphical changes to normal locally evolving coastal processes (Kidson, 1977) and in so doing proposed a smooth sea-level curve (Shennan, 1983). Research in the Somerset Levels by Beckett and Hibbert (1979) suggest that this approach may be oversimplifying south-west Britain's sea-level histories. The argument centres on the controversy between "wigglers" (Greensmith and Tucker, 1973; Devoy, 1979) and "smoothers" (Heyworth and Kidson, 1982) and has been debated for decades (Gehrels, 1999). Although the resolution of mid-Holocene SLIPs from south Devon has been significantly improved in this project, it is difficult to interpret any potential fluctuations in sea-level histories because of the altitudinal spread of the data. This is a problem around many stretches of the UK coastline, e.g., the vertical spread of mid-Holocene SLIPs from the Thames Estuary is as much as 5 metres (Wilkinson *et al.*, 2000), and is particularly related to research carried out on sediments from complex barrier-lagoon systems (Healy, 1993).

Closer inspection of the RSL trend depicted in Figure 8.1 reinforces earlier doubts about index point 17 from core NS-97-4 (see section 5.7.1). It seems to be an "outlier" positioned 2 m above SLIP 31 from Blackpool Sands. This serves to highlight further the problems associated with determining reliably the altitudinal reference of mid-Holocene SLIPs, in particular the indicative meaning of back-barrier sediments. However, the age of the sample is not in doubt and places the cessation of marine influence at North Sands around the same time as Blackpool Sands (*ca.* 5300 to 4400 cal years BP). Furthermore, the entire core section above the dated horizon consists of freshwater peat and indicates that the environment was probably supratidal (Figures 5.7 and 5.19). There appears to be no reason to question the rejection of index point 17 and with a more reliable indicative meaning should plot at a similar height to the cluster of SLIPs below it (Figures 8.3a and 8.4a). Foraminifera found in the sample were probably re-deposited in a freshwater marsh setting at the limit of the inter-tidal marsh environment (section 5.7.1).

Healy (1993) discussed the effect of storm surges on the coastal forests of Cornwall, suggesting that such events are likely to have been the cause of inundation of forests during the mid- to late Holocene. Heyworth and Kidson (1982) suggest that this occurs every 1600 years at times of maximum perigee spring tides, when the sun is in perihelion and both sun and moon have zero declination. One such combination is recorded in south-west England *ca.* 4800 ^{14}C years ago BP. They propose also that the effect of exceptional tides on any sea-level reconstruction can produce a signal that may be at least 1 m higher than MSL (Heyworth, 1978; Kidson and Heyworth, 1979). Such events may have occurred at sites in this project during the mid- to late Holocene and any estimation of the indicative meaning of a SLIP from such facies would require careful scrutiny. The effect on the interpretation of mid-Holocene RSL histories is the construction of a sea-level envelope allowing maximum and minimum rates of RSL rise to be calculated (Gehrels *et al.*, 1996). Results here seem to indicate decelerating rates in the mid-Holocene (Figures 8.3 and 8.4).

8.1.3 The late Holocene RSL history of south Devon

Late Holocene sea-level change in south Devon is constructed from Hails' (1975a,b) and Morey's (1976) data (SLIPs 88, 90 and 91, Table 8.2). Healy's (1993, 1995) SLIPs (95 and 97) from west Cornwall assist in extending the record from south-west England to *ca.* 1000 cal years BP (Figures 8.3a and 8.4a). The south Devon data show that RSL may have risen as much as 2.7 m between *ca.* 2900 and 1800 cal years BP (Figure 1.1). Regression analyses carried out on the three SLIPs yields a rate of 1.0 ± 0.7 m/ka and 1.3 ± 0.3 m/ka including the origin (Table 8.1). The combined south Devon and west Cornwall dataset show that regional RSL was rising at a rate of 0.8 – 1.6 m/ka *ca.* 3200 – 1000 cal years BP (Figure 8.3a).

The data suggest a slight rise in the rate of RSL during the late Holocene (Table 8.1 and Figure 8.3a) but the identification of a clear local and regional trend is very difficult given the limited number of SLIPs available for the area (Waller and Long, 2003). Morey (1983b) identifies a period of barrier stability at Slapton Ley with supratidal conditions dominating the lagoon since *ca.* 3000 cal years BP (section 6.6). Local storm surges during the late Holocene (Heyworth and Kidson, 1982) have interrupted the evolution of freshwater conditions on at least two occasions (Morey, 1976) and washover fans identified in the Ley are evidence of this (section 8.4). MSL has risen at least 5 m since *ca.* 4500 cal years BP at Blackpool Sands and possibly North Sands (the indicative meaning of index point 17 is probably supratidal, see section 8.1.2). However, the evolving freshwater marsh environment at these sites is likely to have been protected by the growth in height of the gravel barrier during the late Holocene (Morey, 1983b). A more recent record of RSL change can probably be obtained from Thurlestone Sands (Figure 1.12) and possibly Bantham Sands (Figures 4.11 and 4.13) back-barrier marshes. The uppermost facies at these sites contain sediments from which an indicative meaning can be derived that are positioned around OD and should produce ages younger than 4000 cal years BP.

Waller and Long (2003) plot index points from Crabs Ledge on Tresco in the Isles of Scilly (Ratcliffe and Straker, 1996) that reconstruct a similar late Holocene (*ca.* 2350 – 1250 cal years BP, Table 8.3) RSL history to Healy's (1993, 1995, 1996) west Cornwall data. The radiocarbon dates were obtained from peat described as inter-tidal salt marsh based on the pollen, diatoms and macrofossil analyses. It would seem that the gradual late Holocene submergence history of the Isles of Scilly could, in addition to the data from Healy (1993), provide informative comparisons with the RSL history from south Devon (Figure 8.3a). They also provide additional data for a time period that is poorly covered by SLIPs from south Devon.

Radiocarbon laboratory number	^{14}C age (years BP)	Max (Median) Min Cal age (BP) ($\pm 2\sigma$ range)	Sample Height (m OD)	Altitude (m MTL)	Facies contact / material dated	Sea-level indicator
GU-5056	1880 ± 100	2041 (1822) 1560	1.22	-1.97 ± 0.5	Top 1 cm of exposed peat	Pollen, diatoms and seeds
GU-5057	1980 ± 80	2145 (1907) 1723	1.03	-2.16 ± 0.5	Basal 1 cm of exposed peat	Pollen, diatoms and seeds
GU-5058	1480 ± 80	1534 (1350) 1265	1.39	-1.80 ± 0.5	Top 1 cm of exposed peat	Pollen, diatoms and seeds
GU-5059	2180 ± 100	2354 (2273) 1903	1.30	-1.89 ± 0.5	Basal 1 cm of exposed peat	Pollen, diatoms and seeds

Table 8.3. Sea-level data obtained from Tresco, Isles of Scilly. Samples extracted from an archaeological trench on Crabs Ledge, south Tresco (SV897138). Calibrations carried out by the author. Vertical errors are estimated. MSL at St. Mary's = 3.19 m OD. Source: Ratcliffe and Straker (1996).

8.2 Testing the geophysical model predictions for south Devon

The secondary aim of this thesis is to test current geophysical models of ice retreat and Earth rheology against the south Devon Holocene RSL history. Results generated from models developed by Lambeck (1993a,b, 1995), Peltier (1998) and Peltier *et al.* (2002) are shown in Figure 1.2 and the developmental stages and variables upon which the models are based are explained in section 1.5. Peltier (pers. comm. 2002) provided post-glacial RSL responses to a range of additional components using the ICE-4G (VM2) global model with a lithospheric thickness of 90 km (Figure 8.5). This followed advances on the global geophysical theory of the glacial isostatic adjustment (GIA) process by Peltier *et al.* (2002). Each of the five sets of results are a total prediction of RSL from that theory and has input regarding ice heights and Earth rheology properties, including relaxation spectra for viscosity models VM1, 2 and 3 (Peltier, 1998, p.622), at each time-step.

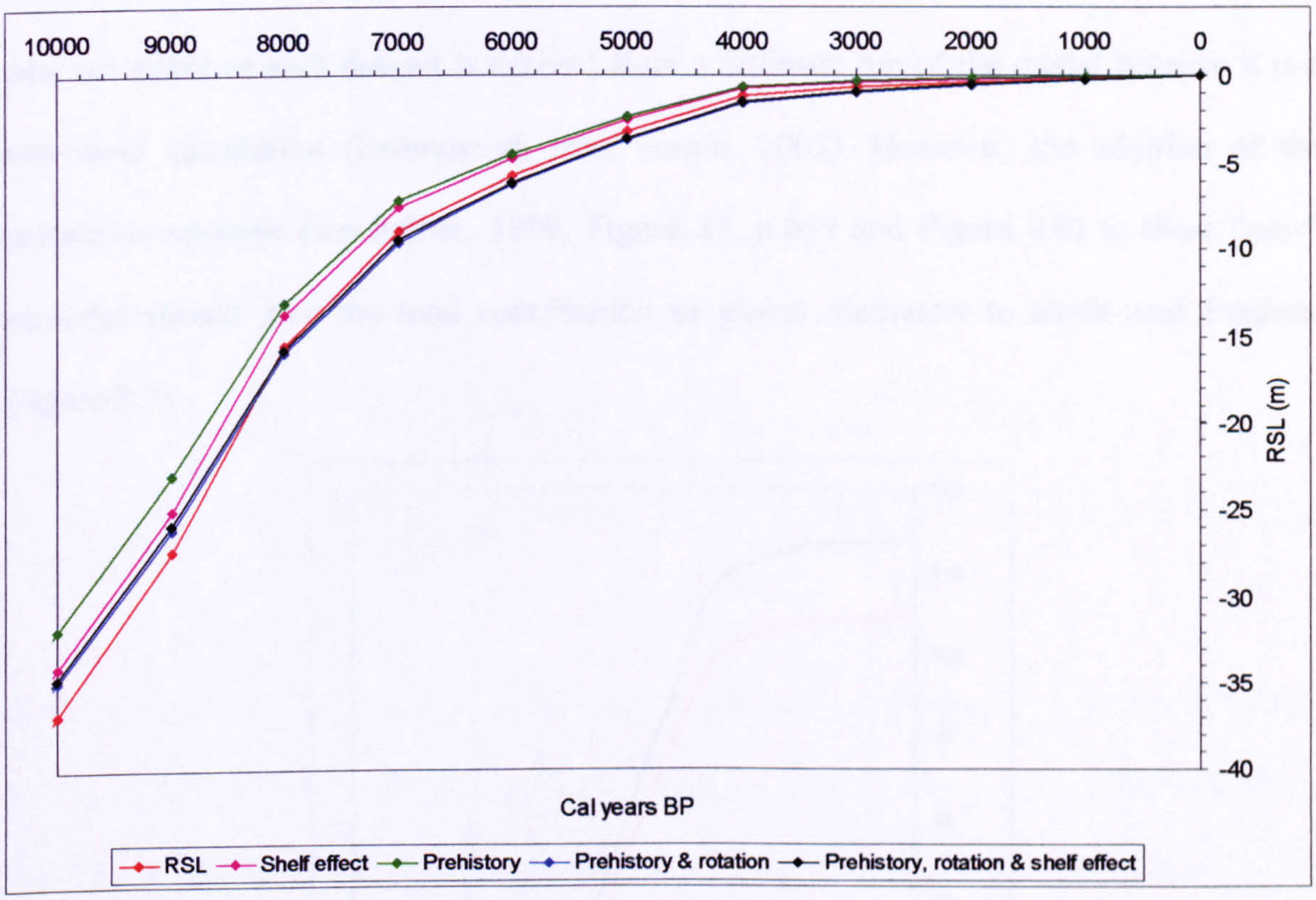


Figure 8.5. Summary of Peltier's (1998) RSL model predictions for south Devon. Predictions computed from the modified ice-model (ICE-4G) and earth-model (VM2) + 90 km lithosphere.

Shelf effect attempts to account for the loading occasioned when dry areas of the Earth's surface become oceans (Peltier and Drummond, 2002; Mitrovica, 2003). Each time-step includes an additional thickness of water in the oceans and the hydro-isostatic response of the Earth's crust to that loading. Prehistory is a crude ice history of the Earth before the Last Glacial Maximum (LGM). The SPECMAP $\delta^{18}O$ history of Imbrie *et al.* (1984) is used to reconstruct results from 780 KBP. The ICE-4G model derived by Peltier (1994) used this data and was employed by Peltier (1998) to reconstruct the history of deglaciation from LGM to present. The effect on planetary rotation by global post-glacial rebound processes is also a significant component of the sea level equation and its iterative solution (Peltier, 1998, p.618). For example, as GIA proceeds following deglaciation the planet becomes less oblate, rotation accelerates which in turn affects Earth surface and ocean tidal loading (Farrell, 1972; Peltier, 1974, 1976; Farrell and Clark, 1976; Peltier and Andrews, 1976; Clark *et al.*, 1978; Peltier *et al.*, 1978). These effects cannot be added to achieve a total net effect as each dataset is derived from a different run of the model because it is a non-linear calculation (Drummond, pers. comm. 2003). However, the addition of the eustatic component (see Peltier, 1998, Figure 37, p.659 and Figure 8.6) to these "new" variables should give the total contribution of global meltwater to south-west England (Figure 8.7).

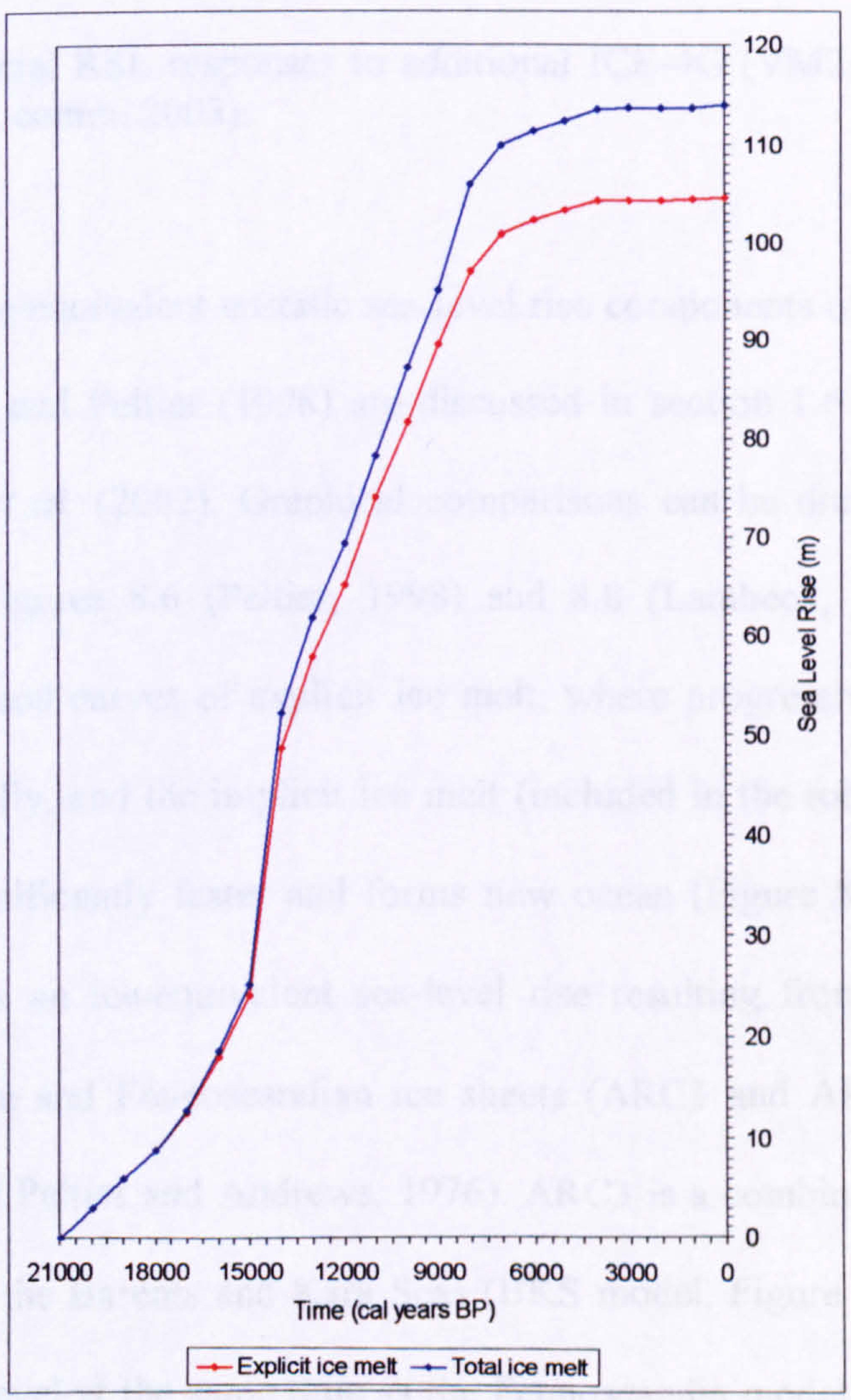


Figure 8.6. Eustatic sea-level rise (ICE-4G model history). Source: Peltier (pers. comm. 2003).

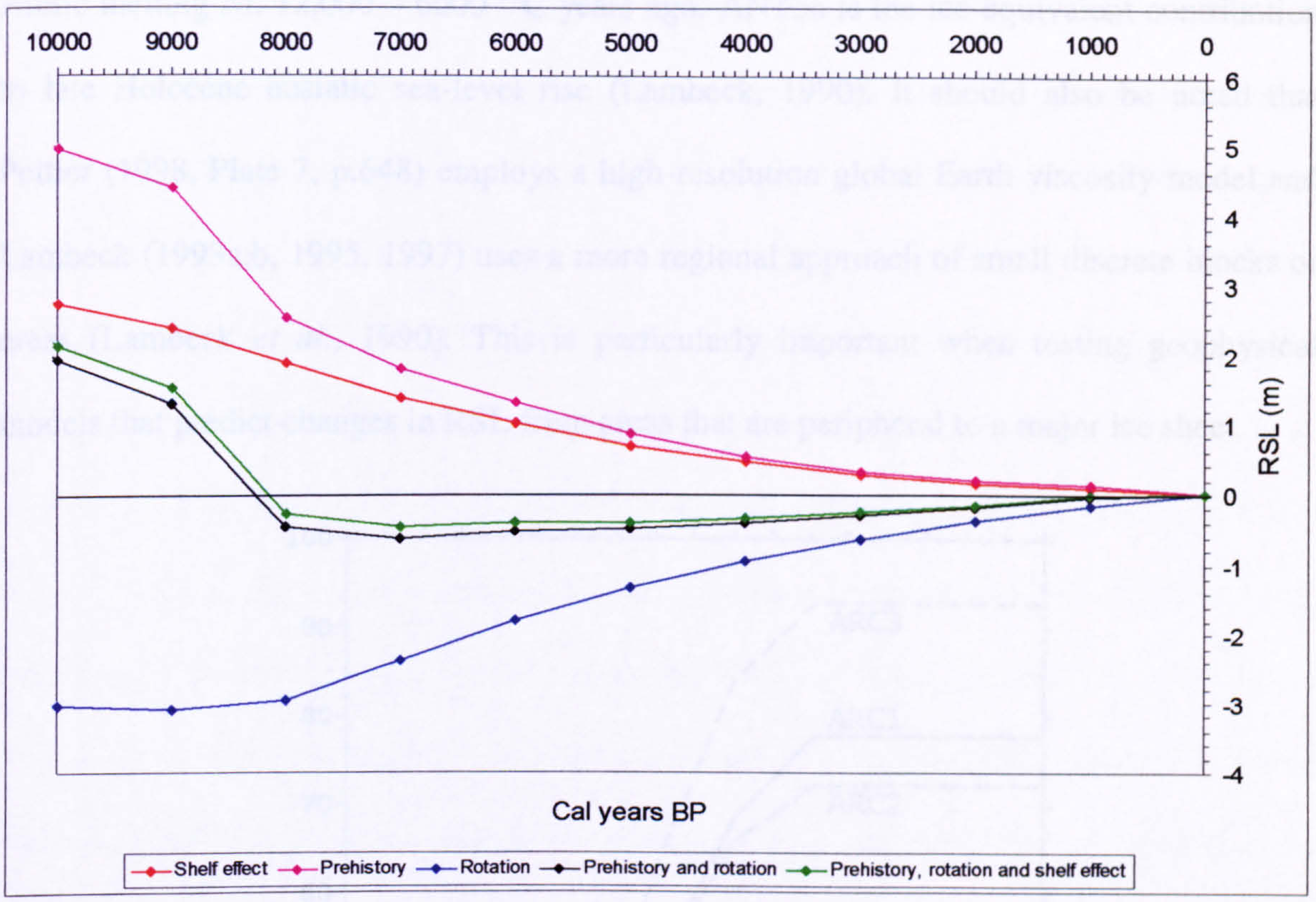


Figure 8.7. Post-glacial RSL responses to additional ICE-4G (VM2+90km) components. Source: Peltier (pers. comm. 2003).

Differences in the ice-equivalent eustatic sea-level rise components of geophysical models by Lambeck (1995) and Peltier (1998) are discussed in section 1.6.1 and by Plag *et al.* (1998) and Peltier *et al.* (2002). Graphical comparisons can be drawn between the two components from Figures 8.6 (Peltier, 1998) and 8.8 (Lambeck, 1990). Peltier (pers. comm. 2003) produced curves of explicit ice melt, where progressive ice melts increase ocean depths gradually, and the implicit ice melt (included in the total ice melt scenario), where ice melts significantly faster and forms new ocean (Figure 8.6). Lambeck (1990, Figure 8.8) employs an ice-equivalent sea-level rise resulting from the melting of the combined Laurentide and Fennoscandian ice sheets (ARC1 and ARC2, from ICE1 and ICE2 respectively in Peltier and Andrews, 1976). ARC3 is a combination of ARC2 and a schematic model of the Barents and Kara Seas (BKS model, Figure 8.8) adding a further 15 m to global sea-level at the same time as the Fennoscandia model (ARC1). ANT1 is in phase with Arctic melting, ANT2 occurs before Arctic melting and ANT3 lags behind

Arctic melting *ca.* 12,000 – 6000 ^{14}C years ago. ANT3a is the ice-equivalent contribution to late Holocene eustatic sea-level rise (Lambeck, 1990). It should also be noted that Peltier (1998, Plate 7, p.648) employs a high-resolution global Earth viscosity model and Lambeck (1993a,b, 1995, 1997) uses a more regional approach of small discrete blocks or areas (Lambeck *et al.*, 1990). This is particularly important when testing geophysical models that predict changes in RSL from areas that are peripheral to a major ice sheet.

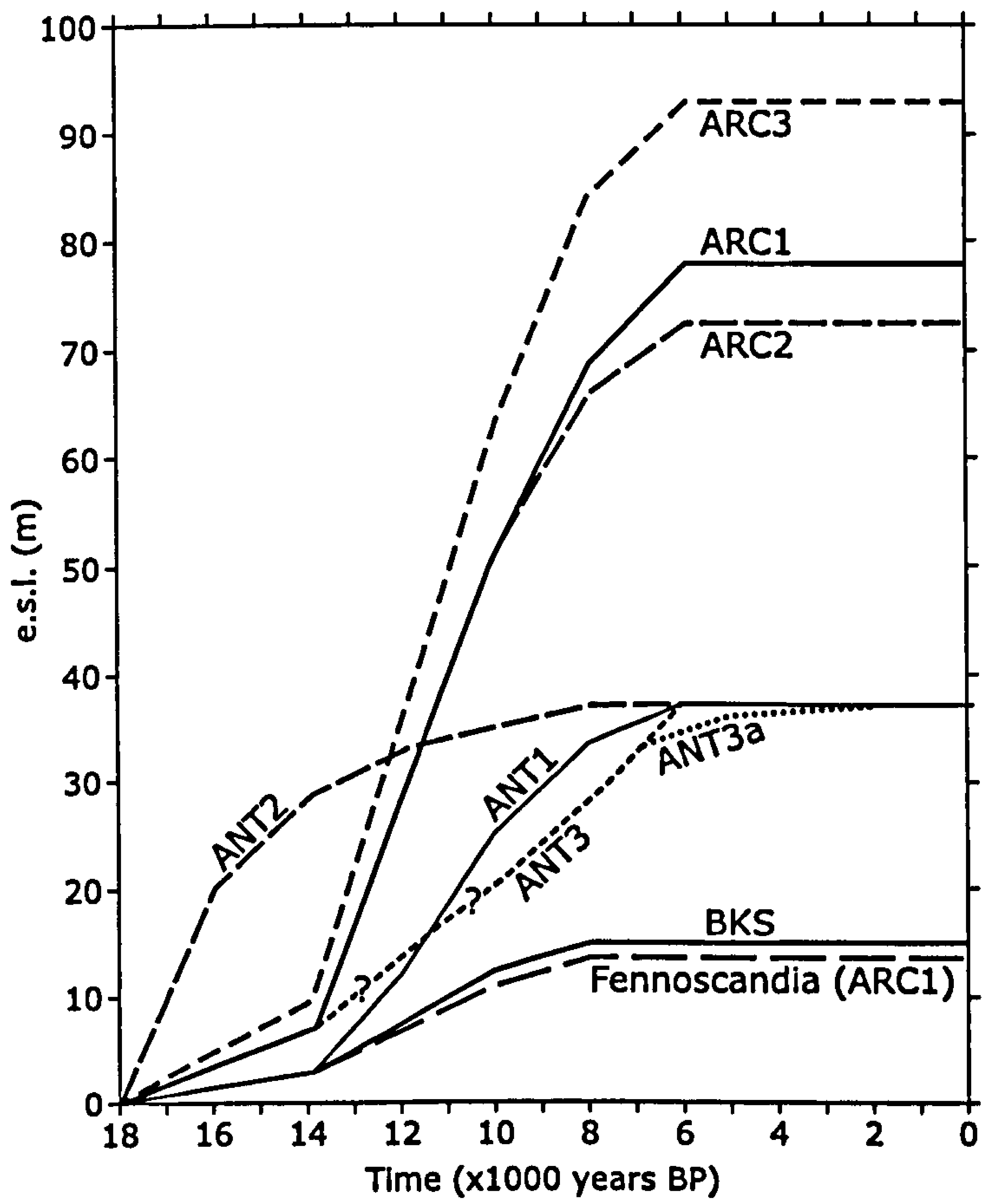


Figure 8.8. Ice-equivalent eustatic sea-level (e.s.l.) rise scenarios employed by Lambeck (1990). See text for explanation.

Index points from this project and other south Devon data (Tables 1.1 and 8.2) are used to test the models. Empirical data from other regions in south-west England are not included. This is because RSL histories computed from different regions can shift the predictions down by >1 m thus diminishing the sense that there is a significant misfit to theory (Peltier, pers. comm. 2003). The RSL model appears to give too rapid a rate of sea-level rise post-7000 cal years BP (Peltier, pers. comm. 2003, Figure 8.5) when compared with data from south Devon. However, this is diminished by the addition of rotational feedback into the calculation. The model runs of prehistory and shelf effect produce the greatest RSL height differential throughout the entire Holocene whilst rotation produces lower RSL heights. The RSL response to the individual components has been isolated by subtracting the results from the unmodified RSL predictions (Figure 8.7). A significant inflection in the RSL response to the individual components occurs at 8000 cal years BP where all contributions except rotation produce a fall in RSL. At 7000 cal years BP rotation produces a rise in RSL when in combination with other processes. In the ICE-4G model the eustatic variation of sea level is assumed to cease after 4000 cal years BP (Peltier, 1998, Figure 8.6). Resolving the discrepancies between model predictions and SLIPs could be achieved by decreasing the rate of ice melting from 7000 cal years BP onwards and continuing the melting for times subsequent to 4000 cal years BP (Peltier, pers. comm. 2003). This is especially noteworthy given the comparisons between mid-Holocene empirical data and model results (Figure 8.9a).

The greatest disparity in RSL height between the models occurs around 7000 to 6000 cal years BP and is more pronounced between Lambeck's (1997) and Peltier's (2000) south Devon data (Figure 8.9a). A number of significant observations can be made when comparing the four RSL histories of Lambeck (1997, 2000) and Peltier (2000, 2002) with each other. For example, Lambeck (2000) plots RSL 2.5 m lower *ca.* 6800 cal BP than Lambeck (1997) and Peltier (2002) plots RSL 0.8 m higher *ca.* 7000 cal BP than Peltier (2000). The late Holocene (*ca.* 4000 – 0 cal years BP) data shows that Peltier predicts much higher sea-level positions than Lambeck. In general, the rates of early Holocene (*ca.* 9000 – 7000 cal years BP) RSL rise are very similar, around 8 – 10 m/ka, and the mid-Holocene (*ca.* 7000 – 4000 cal years BP) rates of Lambeck's data are approximately 1.5 – 2.0 m/ka slower than those predicted by Peltier's (Table 8.4). The late Holocene rates are similar although Lambeck (2000) predicts a faster RSL rise to the present day. It is interesting to note that the entire Holocene sea-level history produced by Lambeck (2000) predicts faster rates of sea-level rise than Lambeck (1997). The differences between the RSL predictions from the models are due primarily to the parameters outlined here and in section 1.6.1 (see also Plag *et al.*, 1998 and Peltier *et al.*, 2002).

Period	Model	Age range (Cal years BP)	Number of time-steps	RSL rate (m/ka)
Early Holocene	Lambeck (1997)	8900 – 7800	2	8.4
	Lambeck (1997)	8900 – 7400	3	9.2
	Lambeck (2000)	8900 – 6800	2	10.1
	Peltier (2000)	9000 – 7000	2	9.4
	Peltier (2002)	9000 – 7000	2	9.2
Mid-Holocene	Lambeck (1997)	6800 – 3700	5	0.6
	Lambeck (2000)	6800 – 3100	3	1.1
	Peltier (2000)	7000 – 4000	3	2.7
	Peltier (2002)	7000 – 4000	3	2.6
Late Holocene	Lambeck (1997)	3700 – 1300	4	0.5
	Lambeck (2000)	4400 – 900	3	0.8
	Peltier (2000)	4000 – 1000	3	0.5
	Peltier (2002)	4000 – 1000	3	0.3

Table 8.4. Modelled Holocene RSL rates for south Devon. Rates calculated using the "connecting-the-dots" method.

Testing the models with the empirical data from south Devon produces some significant results (Figure 8.9a). For this purpose the chosen sea-level envelope is that described in section 8.1 and SLIPs 87-91 listed in Table 1.1. Lambeck (1997) is in close agreement with basal SLIPs from Bantham Sands, North Sands and Slapton Sands between –18 and –12 m RSL (*ca.* 8400 to 7800 cal years BP) and non-basal SLIPs 11, 14, 15 and 16 from core NS-97-3 (*ca.* 7500 cal years BP). However, RSL is plotted around –10 to –8 m at least 1000 years later than SLIPs 6, 9 and 10 (Figure 8.9a). Mid- to late Holocene RSL height is significantly overestimated by Lambeck (1997), particularly *ca.* 6800 to 5000 cal years BP (around 3 to 5.5 m). In contrast, Lambeck (2000) underestimates early Holocene RSL by 3 to 6 metres and overestimates mid- to late Holocene sea levels by only 2 to 4 metres. It is interesting to note that Lambeck (2000) predicts the same height (around –6 m RSL) and similar timing (*ca.* 7000 cal years BP) for the deceleration in early to mid-Holocene RSL rates from SLIPs in cores NS-97-3 and BS-97-3. The misfits are also apparent in a ^{14}C versus height plot (Figure 8.9b) and show that calibrations carried out in this project do not produce misfits that are an artefact of time non-linearities, as opposed to ice and viscosity parameters. In summary, Lambeck's modifications to his 1993a,b and 1995 model parameters appear to have incorrectly lowered RSL during the early Holocene but significantly improved the fit during the mid- to late Holocene. However, the timing for the point of deceleration of RSL rise (*ca.* 7000 cal years BP at –6 m RSL) appears to be more accurately predicted by Lambeck's (2000) modification.

Peltier (2000 and 2002) are in close agreement with the early Holocene basal data, underestimating RSL height by around 2 to 3 metres (Figure 8.9a). Both models underestimate significantly all accepted SLIPs from core NS-97-3. The best predictor of mid- to late Holocene (*ca.* 5700 to 4500 cal years BP) RSL position in comparison with accepted SLIPs (24 to 31 and 33) from Blackpool Sands is Peltier (2000). Basal SLIP 28 is underestimated by around 1 to 2 metres *ca.* 6700 cal years BP but is deemed a close fit with the models. Peltier's (2000) prediction fits basal index point 17 from North Sands core NS-97-4 exactly (Figure 5.31) but the sample is rejected on the grounds of its anomalous indicative meaning (see section 5.7). The sample is most likely from a high marsh or supratidal palaeo-setting and would probably plot lower than Peltier (2000) by a similar magnitude to SLIP 31. In summary, Peltier's (2002) modifications to ICE-4G (VM2) is the better predictor of early Holocene RSL change and his earlier (2000) modifications the best predictor of mid- to late Holocene RSL history. Additionally, the curvilinear form of the empirical sea-level envelope depicted in Figure 8.4 is mirrored closely by the RSL histories produced by Peltier (2000, 2002). This is in contrast, generally, to the similarities between Lambeck's (1997, 2000) predictions and Figure 8.3.

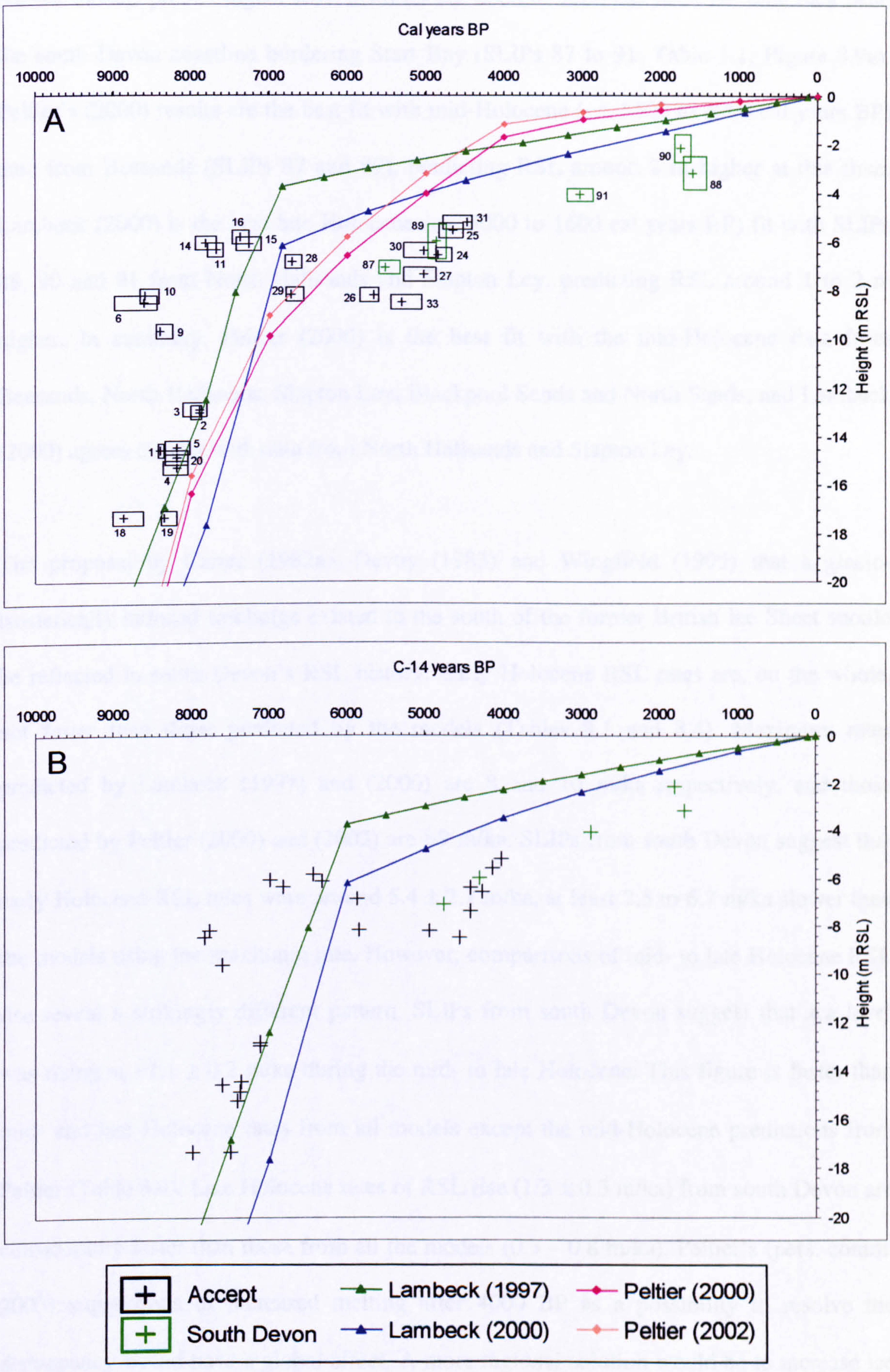


Figure 8.9. South Devon's Holocene RSL history versus geophysical modelling results (Models: Lambeck, 1993a,b, 1995; Peltier, 1998; Peltier *et al.*, 2002). A. Accepted and south Devon's published SLIPs. B. South Devon's Holocene RSL history versus Lambeck's models: C-14 time scale. Crosses on SLIPs are used to test model predictions.

All the models predict higher RSL histories *ca.* 6000 to 1000 cal years BP than data from the south Devon coastline bordering Start Bay (SLIPs 87 to 91, Table 1.1, Figure 8.9a). Peltier's (2000) results are the best fit with mid-Holocene (*ca.* 5500 to 4900 cal years BP) data from Beesands (SLIPs 87 and 89), predicting RSL around 2 m higher at this time. Lambeck (2000) is the best late Holocene (*ca.* 3000 to 1600 cal years BP) fit with SLIPs 88, 90 and 91 from North Hallsands and Slapton Ley, predicting RSL around 1 to 2 m higher. In summary, Peltier (2000) is the best fit with the mid-Holocene data from Beesands, North Hallsands, Slapton Ley, Blackpool Sands and North Sands, and Lambeck (2000) agrees closely with data from North Hallsands and Slapton Ley.

The proposal by Carter (1982a), Devoy (1983) and Wingfield (1995) that a glacio-isostatically induced forebulge existed to the south of the former British Ice Sheet should be reflected in south Devon's RSL history. Early Holocene RSL rates are, on the whole, not faster than those predicted by the models (Tables 8.1 and 8.4). Maximum rates predicted by Lambeck (1997) and (2000) are 8 and 10 m/ka respectively, and those predicted by Peltier (2000) and (2002) are >9 m/ka. SLIPs from south Devon suggest that early Holocene RSL rates were around 5.4 ± 2.1 m/ka, at least 2.5 to 6.7 m/ka slower than the models using the maximum rate. However, comparisons of mid- to late Holocene RSL rise reveal a strikingly different pattern. SLIPs from south Devon suggest that sea level was rising at $\sim 1.1 \pm 0.2$ m/ka during the mid- to late Holocene. This figure is faster than mid- and late Holocene rates from all models except the mid-Holocene predictions from Peltier (Table 8.4). Late Holocene rates of RSL rise (1.3 ± 0.3 m/ka) from south Devon are considerably faster than those from all the models (0.3 – 0.8 m/ka). Peltier's (pers. comm. 2003) suggestions of increased melting after 4000 BP as a possibility to resolve the discrepancy would have a global effect. A more regional solution would be to increase ice loads over the British Isles to amplify the "forebulge effect".

The ICE-4G (VM2) Earth model is especially important in this region because RSL histories in southern England are controlled by the same processes observed on the eastern coast of the U.S. where it was originally tested (Peltier, 1998, p.638). However, the RSL record in southern England is governed by forebulge collapse from both the Fennoscandian ice sheet to the east and the Scottish ice sheet to the north. Hence, results from the Fenlands and Bristol Channel in Peltier (1998, Figure 8.10) that show a similar mid- to late Holocene misfit (the model overestimates the data) with the curve produced by VM2, to the data in this project may indicate some predictive problems with the Earth rheology model. Mid-Holocene misfits (generally overestimates) are also apparent in Lambeck's (1993a,b; 1995) Earth model (GB-3) comparisons of sea-level observations with predictions for the Fenlands, but less so for the Bridgwater, Bristol Bay data (Lambeck *et al.*, 1996, Figure 8.11).

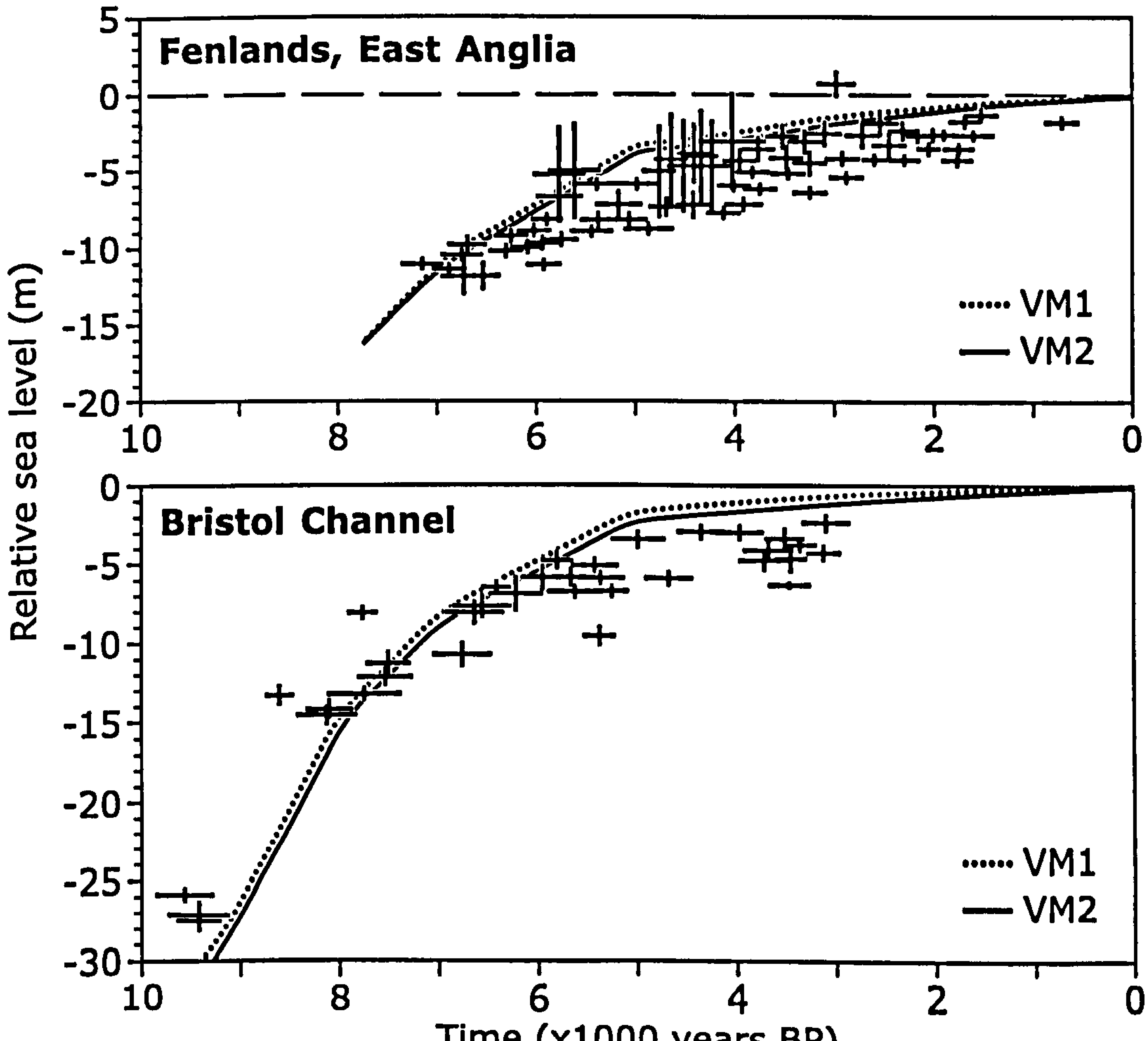


Figure 8.10. ICE-4G (VM1 and 2): Comparisons of predicted and observed sea-level data from the Fenlands, East Anglia and the Bristol Channel (Peltier, 1998).

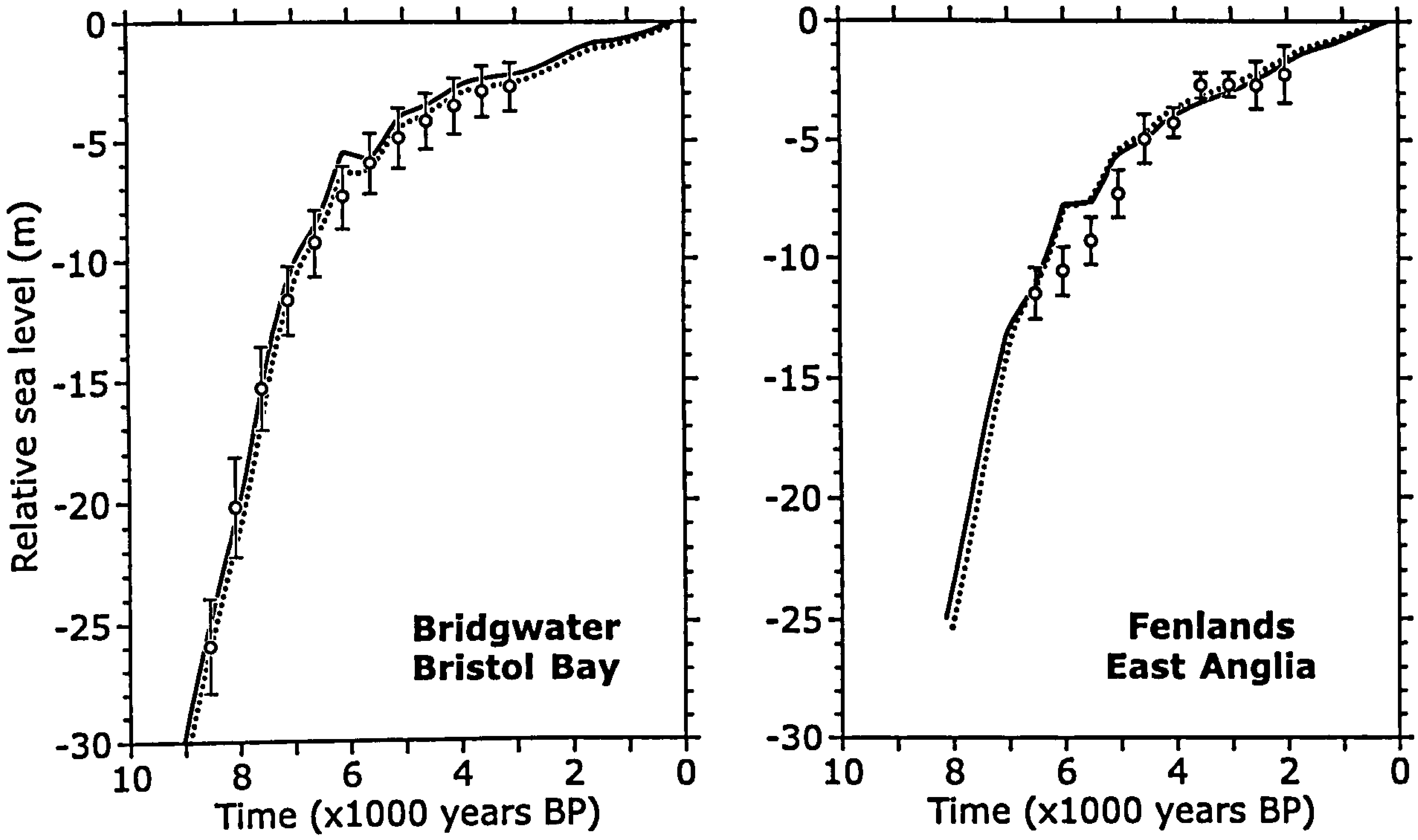


Figure 8.11. GB-3: Comparisons of predicted and observed sea-level data from the Fenlands, East Anglia and Bridgwater, Bristol Bay (Lambeck *et al.*, 1996). Solid lines: 5-layer Earth model, dashed lines: 3-layer Earth model.

Geophysical models are based on mathematical formulae derived from estimations of former ice loads, ice retreat reconstruction, lithospheric thickness and mantle viscosity. They are dependent on comparisons with local and regional RSL histories but are rarely calculated at time slices smaller than 500 or 1000 years. Shennan *et al.* (2000) state that "at present there is no unique solution of Earth and ice model parameters that will explain all the sea-level observations" and this statement is undoubtedly true in relation to observations made in this project. For example, some of the issues recently addressed by the modellers have been revisions to Fennoscandian ice sheet models (Lambeck *et al.*, 1996), the continuation of far-field deglaciation throughout the early to mid-Holocene (Fleming *et al.*, 1998), and the provision of physical evidence to resolve imperfections with the British ice sheet model (Ballantyne *et al.*, 1998).

A comparison of Lambeck's (1997) and (2000) RSL histories for south Devon (Figure 8.9a,b) show that frequent changes to theory, from testing of the models with empirical evidence, often bring remarkable results. The maximum ice thickness used by the modellers is one such issue that is constantly revised with respect to field evidence, e.g., the height of trim-lines on Scottish peaks (Watts, 1977; Ballantyne, 1984; Sutherland, 1991). The lowering of Lambeck's RSL history for south Devon (1997 to 2000) is primarily a result of changes to GB-3 with respect to a number of these parameters. The variables include thicker ice over northern Scotland, a new Scandinavian ice model, far-field deglaciation continuing post-7000 years BP (Shennan *et al.*, 2000) and changes to hydrodynamic models used by Lambeck (e.g., Wingfield, 1995). Furthermore, alterations to ice sheet thickness, ice volume and the deglaciation history over northern Britain is likely to have an effect on the evaluation of rates of crustal movement in the south. This is especially important given that current opinion suggests considerable rates of ongoing late Holocene crustal subsidence in south-west England (Shennan and Horton, 2002).

8.3 Crustal movement along the south Devon coastline

The tertiary aim of this thesis is to calculate rates of crustal motion along the south Devon coastline. Shennan's (1989) map of post-glacial crustal movements around Great Britain showed between –0.1 and –1.4 mm yr^{-1} subsidence in south-west England (Figure 1.13a). Crustal residuals were based on over 400 SLIPs available from a databank of 904 around the UK, and Mörner's (1984) regional 'eustatic' curve. Shennan and Horton (2002) improved the quantity and quality of the dataset, but it is notable that south-west Britain remains an area where data points are still relatively sparse (Figure 1.3). In particular, the crustal residual for Devon is based on a total of only 8 Holocene SLIPs, including 5 from south Devon, 1 from north Devon (index nos. 86-91, Table 1.1) and 2 limiting dates (lab. no. SRR237, Hails, 1975a, Table 6.1 and lab. no. NPL86, Clarke, 1970, Table 7.1). The limiting dates used by Shennan and Horton (2002) are not included in the Holocene SLIP database because of problems identified with the basal peat facies from which the radiocarbon samples were extracted (Horton, pers. comm. 2003). Furthermore, all 8 (pre-

1980) SLIPs have had vertical errors evaluated by the Sea-Level Research Unit at the University of Durham (Environmental Research Centre, 2003, Table 1.1) except for any possible autocompaction of the 6 intercalated samples (see section 8.1). This is likely to have some effect on the best estimate of crustal subsidence since *ca.* 4000 years ago, but how much is speculative.

Shennan and Horton (2002) state that "sediment consolidation, arising from autocompaction as the sediment accumulates and from land drainage (anthropogenic) increases the subsidence in areas with thick sequences of Holocene sediments..." adding an extra 0.2 mm yr^{-1} (0.5 – 1.1 mm yr^{-1} in south-east England) to the best estimate. This is a serious consideration if crustal residuals are to be calculated from index points uncorrected for autocompaction. Geotechnical corrections to Holocene sequences in this project have produced as much as 2.2 m vertical adjustment for marine clays overlain by thick peat facies and 1.2 m for minerogenic sequences (sections 5.6 and 7.5). It is therefore crucial that basal or intercalated SLIPs corrected for autocompaction are used to calculate best estimates of crustal subsidence here.

Other factors influencing the post-glacial isostatic adjustment of a coastline cannot be ignored. The English Channel connected with the North Sea *ca.* 8300 cal years BP (Waller and Long, 2003) and this event would have added to the effect of the post-glacial isostatic adjustment of south-west England. The subsequent sediment loading of the English Channel may have increased crustal subsidence on the south coast although Lambeck (1997) suggests that this effect was minimal. Similarly, glacio-hydro-isostatic rates are likely to have been significantly greater than any local crustal tectonic movement (Lambeck, 1997). It is interesting to note that previous authors (Hawkins, 1971; Emery and Aubrey, 1985; Woodworth, 1987; Allen and Rae, 1988; Allen 1991) have commented on crustal instability in the region following research into topics such as changes in tidal range and sediment transport. However, data that may explain these additional factors acting on crustal movement in south-west England is currently limited.

Dataset	No. of SLIPs	No. of limiting dates	Total SLIPs	Rate since *ca.* 4.5 kyr BP (m/ka)	S.E. (m/ka)	Best estimate (m/ka)
This project	24	0	24	–0.89	±0.25	–1.16
Shennan and Horton (2002)	6$	2*	8	–1.43	±0.18	–1.23

Table 8.5. Estimates of crustal subsidence along the south Devon coastline. Rate (from four SLIPs younger than 4500 cal years BP) and best estimate (from youngest SLIP 31, see Figure 8.12) calculated according to Shennan and Horton (2002). Negative sign (rate and best estimate) = subsidence. *Rejected from south-west Britain Holocene SLIP database. $Includes north Devon SLIP No. 86 (Table 1.1). S.E. = Standard Error of the rate since *ca.* 4500 cal years BP.

This project increases the number of south Devon SLIPs from 5 to 29 index points (Table 8.5). These are plotted against the rate of subsidence published by Shennan and Horton (2002) and the most recent model predictions (Lambeck, pers. comm. 2000; Peltier *et al.*, 2002) of south Devon's RSL history from this project (Figure 8.12). Shennan (1989) determined subsidence rates in south-west England by subtracting Mörner's (1984) eustatic sea-level heights from each SLIP then calculating a best-fit linear trend. The limited number, scatter and spread of the SLIPs severely restricted the calculation. The eustatic model adopted by Shennan (1989) was approximately zero since *ca.* 4000 years ago so net rates and best estimates were calculated from SLIPs around this time period. Shennan and Horton (2002) assume that global sea levels have remained stable during the past 4000 to 5000 years and local RSL rise has been linear and entirely due to land subsidence. It should be noted that subsidence rates would be lower if melting did occur after *ca.* 4000 cal years BP (see section 8.2). The best-fit line was placed from the median age and mid-point of MTL for SLIP 31 (the youngest SLIP in this dataset at *ca.* 4471 cal years BP) through the origin in accordance with Shennan and Horton (2002). Linear regression produces a best estimate of crustal subsidence of –1.16 m/ka (Table 8.5 and Figure 8.12) similar to Shennan and Horton (2002).

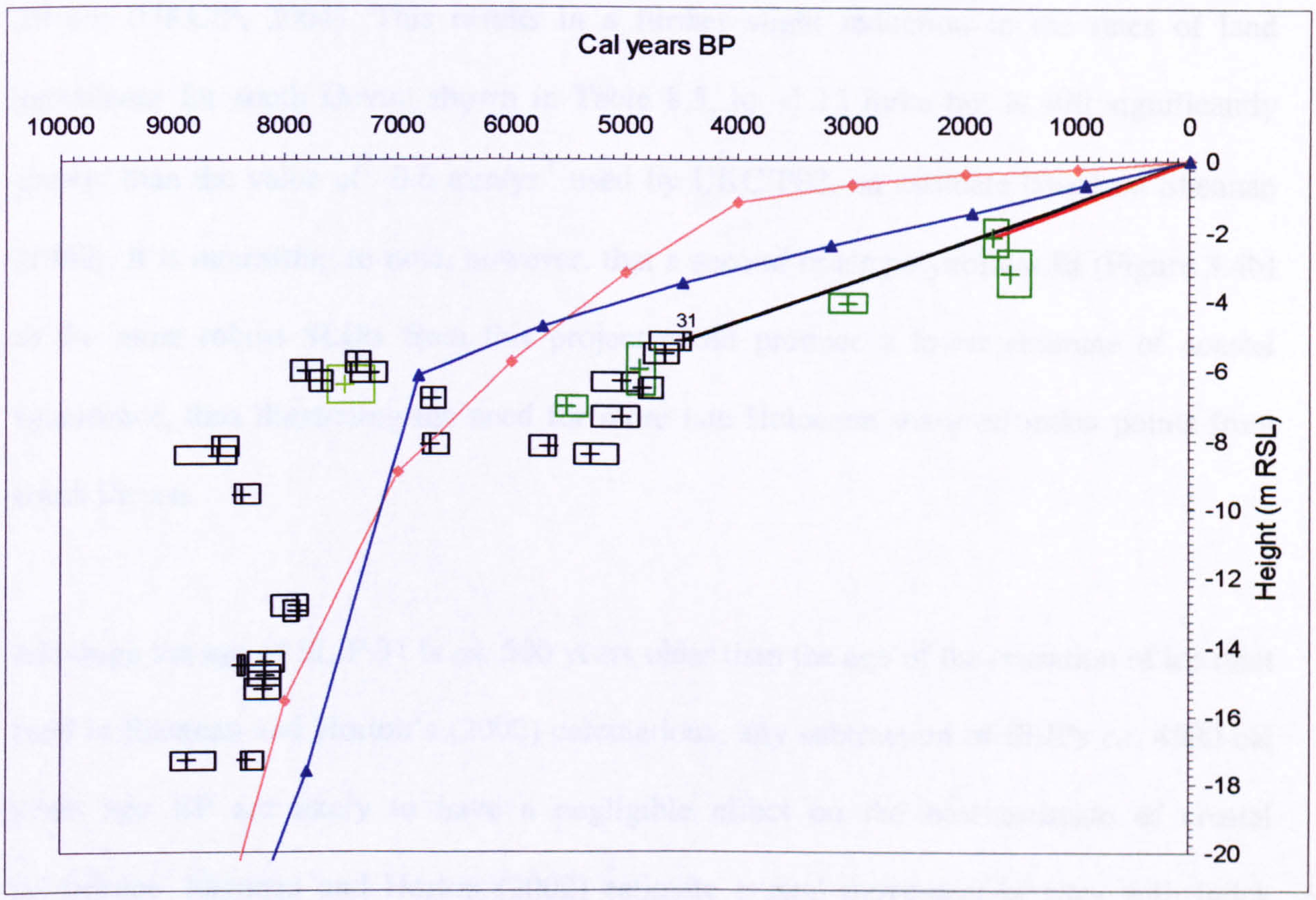


Figure 8.12. The late Holocene estimate of crustal subsidence for south Devon. Solid black line: best estimate for south Devon (Table 8.5), solid red line: estimate from Shennan and Horton (2002), solid coral line: Peltier *et al.* (2002) model RSL predictions, solid blue line: Lambeck's (2000) model RSL predictions. SLIPs: black: this project, bright green: south Devon, lime: north Devon (see Figure 2.6). Rejected limiting dates not shown.

An important consideration here is the effect on subsidence rates in south-west England if melting did occur after *ca.* 4000 cal years BP (Peltier, pers. comm. 2003). At LGM, Great Britain was covered by at least 1200 m of ice over Scotland (Lambeck, 1995) resulting in a (glacio-isostatically-induced) forebulge south of Britain (Shennan *et al.*, 2002). The south-west of England remained ice-free throughout the last glaciation, post-glacial RSL rise being the result of global sea-level rise and land subsidence as GIA took place in the region (Peltier, 1998). Land levels across the whole of southern Britain began to fall as the forebulge collapsed and shrank in size following deglaciation (Carter, 1982a; Devoy, 1983; Wingfield, 1995), a process that is still going on today (Shennan and Horton, 2002). However, the melting of glaciers and ice caps during the last century and thermal expansion of the oceans has contributed to recent global sea-level rise in the order of 10 to 20 cm (UKCIP, 2004). This results in a further slight reduction in the rates of land subsidence for south Devon shown in Table 8.5, to –1.12 m/ka but is still significantly greater than the value of –0.6 mm/yr^{-1} used by UKCIP02, an estimate based on Shennan (1989). It is interesting to note, however, that a second order polynomial fit (Figure 8.4b) to the more robust SLIPs from this project could produce a lower estimate of coastal subsidence, thus illustrating the need for more late Holocene sea-level index points from south Devon.

Although the age of SLIP 31 is *ca.* 500 years older than the age of the cessation of ice melt used in Shennan and Horton's (2002) calculations, any subtraction of SLIPs *ca.* 4500 cal years ago BP are likely to have a negligible effect on the best estimate of crustal subsidence. Shennan and Horton (2002) estimate crustal movement at sites with index points older than 4000 cal years BP, e.g., North Wales, Tay Valley, Dornoch Firth and Moray Firth, citing the dissimilarities in the literature regarding the timing of cessation of global ice melting (Horton, pers. comm. 2003). For example, Mitrovica and Milne (2002) and Peltier (1998) assume that the global deglaciation event ceased *ca.* 5000 BP, Lambeck (1991, 1995) models the event at *ca.* 7000 – 6000 BP and Devoy (1995) suggests that GIA reasserted itself in Britain *ca.* 6000 BP. However, these estimates are frequently revised with respect to the growing body of empirical evidence used to test theory (Peltier, 1994). The best estimate for crustal subsidence in south Devon from this project is only 0.07 m/ka lower than that determined by Shennan and Horton (2002, Table 8.5, Figure 8.13). This is not significant but may yet be lowered further given that the geotechnical correction for SLIP 31 is likely to be greater than that already predicted (Table 7.4 and section 7.5.3).

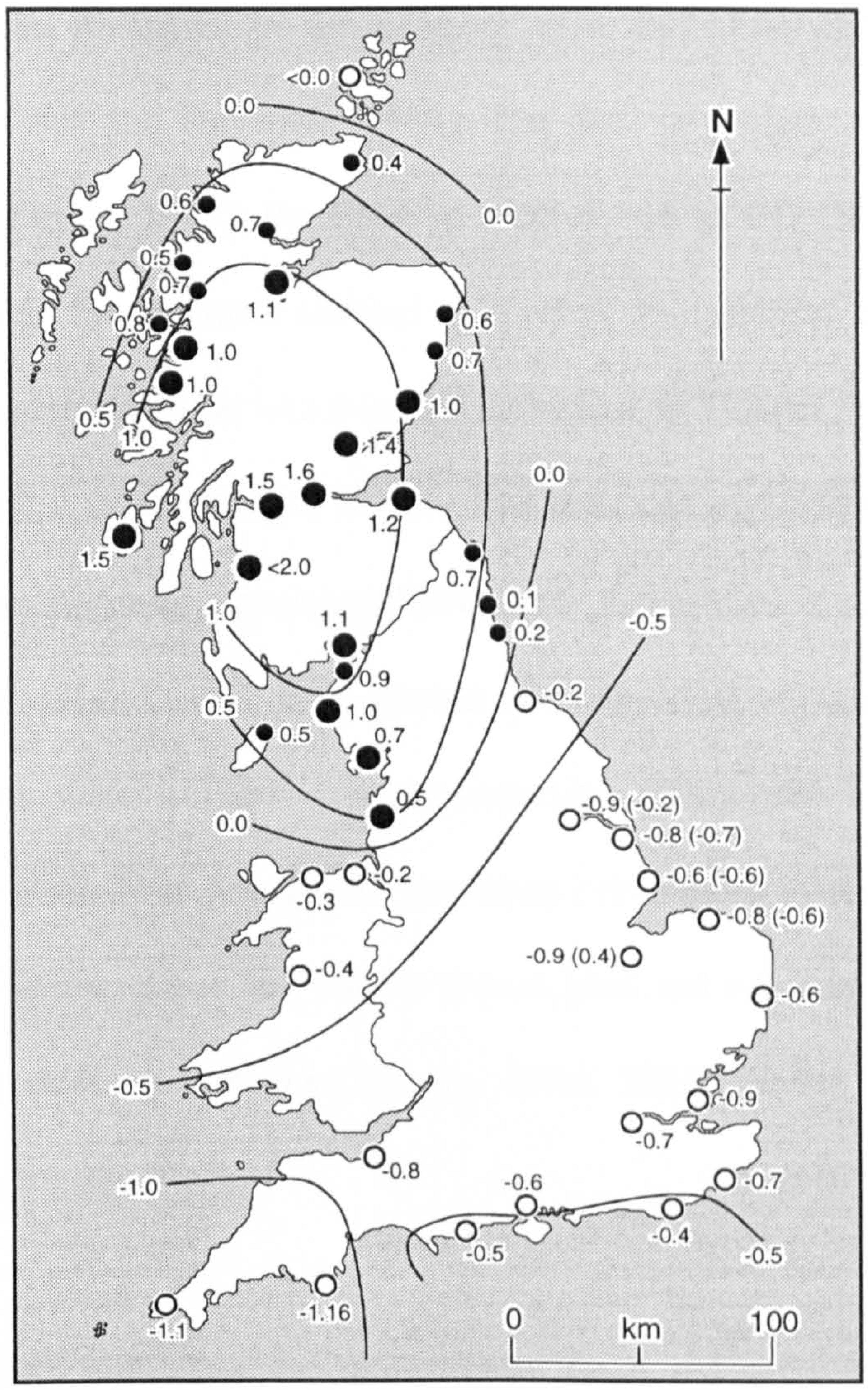


Figure 8.13. Late Holocene crustal movement in Great Britain plus new data from south Devon (mm yr^{-1}, after Shennan and Horton, 2002). Negative sign = subsidence.

8.4 The Holocene coastal evolution of south Devon

A primary objective of this project is to describe the lithostratigraphy of 12 cores from four neighbouring back-barrier systems along the south Devon coast and reconstruct the Holocene coastal environments. The geology of the foreshore around south Devon is predominantly Lower Devonian (Meadfoot Group and Dartmouth) slates and shales, known locally as shillet, e.g., Bantham Sands and Blackpool Sands (Figures 4.3 and 7.3), and Permo-Triassic sandstones, e.g., Slapton Sands (Figure 6.3). Devonian slates, siltstones and sandstones have been altered to form mica and hornblendic schists on the southernmost tip of Devon, e.g., North Sands (Ussher, 1890, 1904; Orme, 1960; Dineley, 1961, Figure 5.3). River channels that have cut into the underlying bedrock during the

Quaternary have been identified in the offshore zone and along the present day south Devon coastline (Eddies and Reynolds, 1988). The Dart is graded to at least –42 m OD (Codrington, 1898; Green, 1949), the Erme to around –46 m OD (Mcfarlane, 1955) and Clarke (1970) records the offshore channel in Torbay to a similar depth. These depths coincide with a submerged cliff at –42.6 m OD identified by Cooper (1948) to the south of Plymouth Sound, and a similar feature in Start Bay recorded by Hails (1975a,b) at a depth of –42.5 m OD. This suggests a prolonged stillstand of sea level at some time during the Quaternary when the south Devon rivers graded to a base level of between –40 and –50 m OD. River channels were infilled then subsequently submerged by minerogenic and organic deposits as a result of post-glacial sea-level rise. Facies of river gravel and head, alluvial sediments, fine sand and silt, barrier gravel, peat and submerged forests are found in and along these submerged valleys (Steers, 1946, 1981; Clarke, 1970; Mottershead, 1971).

The south Devon coast was covered extensively by woodland at the start of the Holocene. Emerging from the uplands of Dartmoor, river valleys were probably densely forested in their lower reaches during this period. Remnants of forests submerged by the post-glacial marine transgression of south-west England have occasionally been uncovered by storms and subsequently identified at low water of spring tides. Healy (1993) documented 34 observations of submerged forest deposits along the north and south coastline of Cornwall. Similar records can be found along Devon's shores (Pengelly, 1865, 1866b, 1868, 1869; Ellis, 1866, 1867; Hall, 1879; Hunt, 1881; Pidgeon, 1885; Inkermann Rogers, 1908; Winder, 1923; Worth, 1934; Roger, 1946; Churchill, 1965a,b; Clarke, 1970). Deposits typically consist of organic-rich sediments and abundant macrofossil remains that have been used to produce radiocarbon dates for the reconstruction of former sea levels, e.g., Morey (1974, 1983a,b) along the Slapton and Blackpool Sands coastline. These samples provide an upper limit of sea level at the time and index points routinely plot lower than those with a brackish or marine provenance do (Figures 8.2, 8.3a and 8.4a). Very few of the submerged forests found along the Devon coast are uncovered at present. However, Pengelly's (1866b) description of a wood that once stood on the shores of Thurlestone, south of Bantham Sands, is one such well-known local example that is frequently clearly visible at low spring tide. At Mothecombe at the mouth of the River Erme another relict forest is visible at MLWST (Thomas, pers. comm. 1997).

Holocene sediments from south Devon cores (Figures 4.6, 5.8, 6.7 and 7.7) show that the bedrock was covered by a mixture of salt marsh and freshwater peat, inter-tidal minerogenic-peat and thin wood facies subsequently overlain by barrier gravel and mudflat deposits. Shoreline position was probably changing constantly under rapidly rising local RSL regimes as on other barrier-dominated coastlines (Carter, 1982a,b; Leatherman, 1983; Orford, 1987; Healy, 1995). Pollen results from North Sands show that some reworking of basal sediments took place at the site but the environment was probably dominated by oak woodland post-8000 years ago. The inundation of basal peat sequences occurred as early as *ca.* 9000 cal years BP on the present day shoreline of Slapton Sands and around 2000 to 3000 years earlier offshore of the North Sands coastline.

Bantham Sands experienced marine conditions as early as *ca.* 8400 cal years BP and Blackpool Sands much later at *ca.* 6800 cal years BP (Figures 4.10 and 7.14). However, these ages are based on a limited number of cores per site and electrical resistivity results (Figures 4.7, 5.9, 7.10 and 7.11) have shown that deeper minerogenic sequences may exist at all sites. These facies may provide evidence of inter-tidal palaeoenvironments that would have formed on the gently sloping rock foreshore and shifting sand flats of pre- and early Holocene south Devon. Clarke (1970) found evidence of similar facies in cores from offshore Torbay *ca.* 9500 BP when he predicted MSL to lay at –22 fm (–40.23 m MSL), below the submerged cliff identified off Plymouth Sound by Cooper (1948) and Start Bay by Hails (1975a,b). In general, at –14.77 m OD south Devon was a marginal marine environment *ca.* 9000 to 8400 cal years ago. The majority of the local coastline below –13.66 m OD was inter-tidal by *ca.* 8100 cal years BP and all sites investigated here show that the marine transgression had reached –6.15 m OD by 6800 cal years BP.

In light of new findings from this thesis the model presented in section 1.4.3 (Figure 1.9, Kraft and Chrzastowski, 1985) is re-considered here and criteria are suggested for the identification of back-barrier environments from lithostratigraphic records. The litho- and biostratigraphical results reveal a complex history for the pre- and early Holocene marine transgression of south Devon. Barrier systems were probably more extensive than today and contained tidal inlets and lagoons (e.g., Clarke, 1970; Carr and Blackley, 1973, 1974; Hails, 1975a,b; Kelland, 1975; Kidson, 1977; Mottershead, 1977; Campbell *et al.*, 1998). Kraft (1971), Belknap and Kraft (1977, 1981, 1985), Niedoroda (1985), Kraft *et al.* (1987), Belknap (1991) and Barnhardt *et al.* (1997) describe a transgressive model of landward barrier movement in which the preservation of back-barrier sediments is controlled by the antecedent topography and depth of shoreface erosion. The ravinement unconformity moves landward and upward, thereby destroying the top of the back-barrier sequence. The best preservation is found in palaeo-channels.

Palaeoenvironments in the three-stage model from Kraft and Chrzastowski (1985, p.644, Figure 9-10) and Belknap and Kraft (1985, p.256, Figure 12) include offshore sand and gravel barriers, tidal inlets, salt marshes, mudflats, and secondary creeks in the back-barrier lagoon (Reinson, 1984). Freshwater lagoon deposits, such as *gyttja* (found in short cores at Bantham Sands and upper facies from Slapton Sands), and supratidal peat marsh and wood is contained in those facies following barrier closure. Overwash facies are indicative of cross-barrier sediment transfer (Morey, 1982b) and net shoreface erosion eventually results in the formation of a headland beach (Kraft, 1979; Kraft and Chrzastowski, 1985).

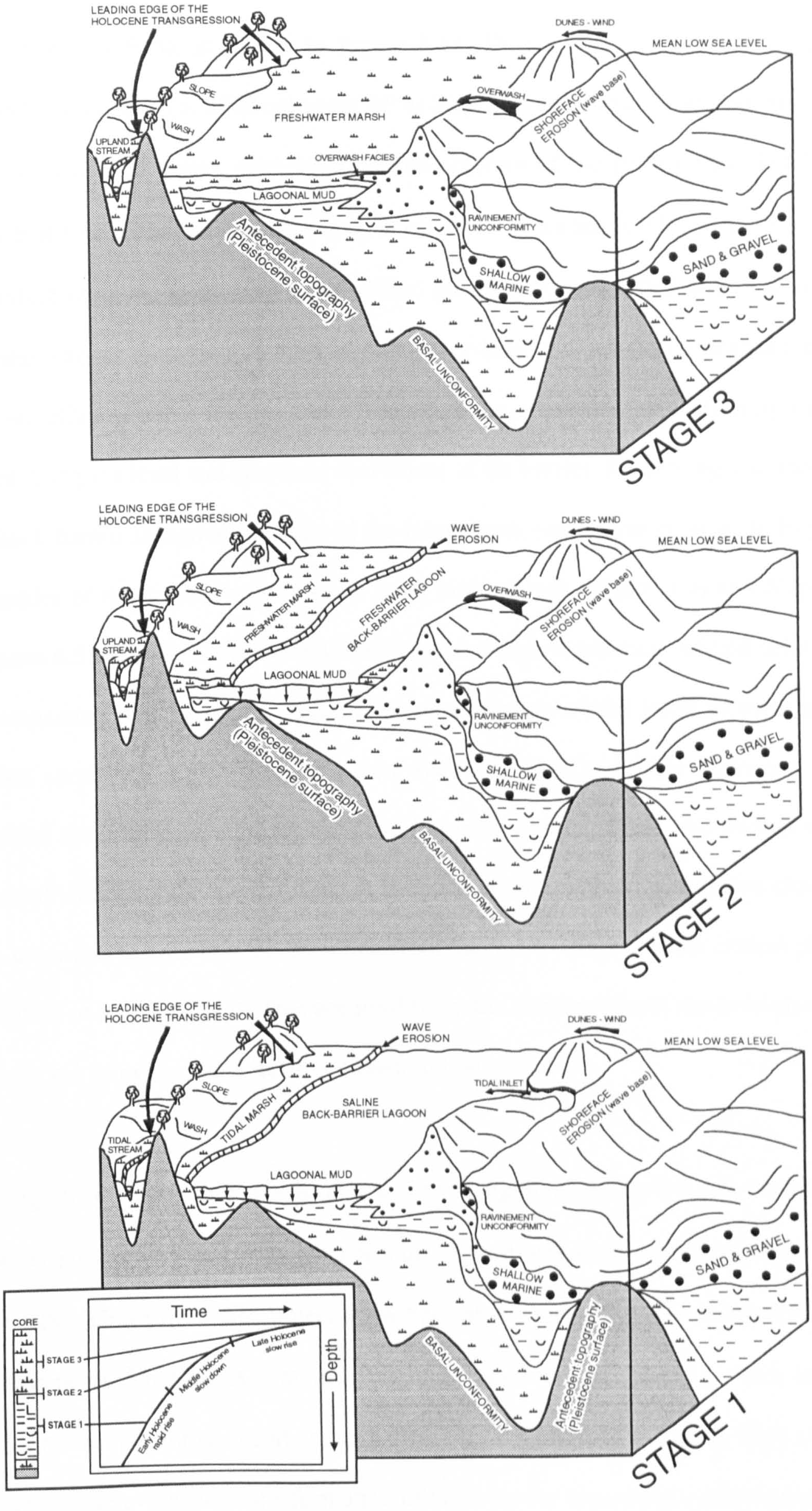


Figure 8.14. A transgressive model of landward barrier movement for south Devon. Redrawn from Kraft and Chrzastowski (1985).

A transgressive model of landward barrier movement, adapted from Kraft and Chrzastowski (1985), is shown in Figure 8.14. During Stage 1 sea levels were lower (almost –18 m RSL *ca.* 9000 cal years BP at Slapton Sands) than today and barrier islands and / or sand and gravel spits were located seaward of the present-day shoreline. Core records in this project have shown that, as early-Holocene sea level rose at 5.4 ± 2.1 m/ka, a mixture of minerogenic and cobble-sized gravel was transported from the offshore to nearshore zones along with reworked older inter-tidal and periglacial deposits, e.g., head. The recycling of old sediments took place while the ravinement unconformity was formed under rising sea level and landward movement of the barrier. Reworking also took place in the back-barrier lagoon at the edge of the tidal marsh (see 'wave erosion' in Figure 8.14). Examples of this can be seen in cores from North Sands (Figure 5.6) and Slapton Sands (Figures 6.5 and 6.6). Brackish and freshwater marsh environments formed behind barriers encompassing broad lagoons. These later became narrow tidal inlets as more gravel was moved shoreward under rising seas. At Bantham, pre- and early Holocene sand flats supplied the source material for the dune formations there. These also moved shoreward, eventually choking the marine inlet, as the pace of the marine transgression slowed down. It is important to note that the transition from Stage 1 to Stage 2 is the critical phase in the evolution of Slapton Sands. This occurred when the average rate of sea-level rise slowed to <3 mm yr^{-1} using the curvilinear fit shown in Figure 8.4b.

During Stage 2 the rate of sea-level rise slowed significantly (to 0.9 ± 0.4 m/ka) *ca.* 7000 to 4000 cal years BP and tidal inlets became choked with sediments at approximately –6 to –5 m RSL. Cores from this project show that saline lagoons rapidly changed to freshwater environments. Sand facies containing marine creek foraminifera are overlain immediately by freshwater peat or sand and gravel facies, e.g., Figures 5.17 and 6.13. Only at Blackpool Sands is there evidence of the formation of back-barrier organogenic salt marsh at the time of barrier closure, e.g., Figure 7.13. The advancing barrier-complex, at Slapton Sands in particular, was probably discontinuous until *ca.* 3000 cal years BP (Morey, 1976) by which time many of the Quaternary valley mouths were dammed by shingle. These were subsequently filled in with silts and clays, e.g., Blackpool Sands (Morey, 1976, 1983b), or developed into back-barrier freshwater lake and extensive marsh systems, e.g., Slapton Sands (Hails, 1975a,b). North Sands has the longest record of freshwater peat, deposited during and after barrier closure. Almost 10 m of peat has been formed at the site since *ca.* 7400 cal years BP, the marine inlet closing there possibly as late as 4400 cal years BP. The possible elm decline identified in the peat sequence around –6 m OD at North Sands suggests a date of *ca.* 5300 ^{14}C years BP (Birks, 1989; Peglar, 1993; Parker *et al.*, 2002) for the facies but the data is inconclusive (see section 5.7.2). Bantham Sands contains a shorter, approximately 4 m, record of peat growth but this is mixed with facies of sand and shells transported from the nearby dune system. A freshwater lake, inferred from *gyttja* found in short cores, also existed at the site probably post-4000 cal years ago. Barrier closure at Blackpool Sands and North Sands occurred around the same time suggesting that the smaller former marine basins sealed off at least 1400 years earlier than Slapton Sands.

In Stage 3 the rate of sea-level rise continued to slow down, but data from Hails (1975a,b) and Morey (1976) show a slight increase to 1.3 ± 0.3 m/ka (Table 8.1), and the freshwater lagoons filled in with peat or minerogenic sediments. This stage has been completed at Beesands and Hallsands (Figure 1.12), and the Higher Ley at Slapton (Figure 6.2) is now in an advanced state of terrestrialisation. The Lower Ley coastal section at Slapton Sands is in Stage 2 of the Holocene transgressive model of barrier movement (Figure 8.14) whilst Bantham Sands, North Sands and Blackpool Sands have reached Stage 3. The entire system freezes once the barrier is pinned against the Pleistocene surface and the tidal inlets have been completely choked under rising sea levels in the late Holocene. Future rapid sea-level rise (UKCIP, 2004) will return the system from Stage 3 to Stage 1.

Criteria for the identification of back-barrier environments from facies in this project contribute to findings from coastal models developed along the U.S. Atlantic coast and south-west England, e.g., Kraft (1971), Belknap and Kraft (1977, 1981, 1985), Morey (1980, 1983b), Kraft and Chrzastowski (1985), Belknap (1991), Barnhardt *et al.* (1997) and Jennings *et al.* (1998). Supporting evidence for the geological provenance of sediments is derived from the micro- and macrofossils found within them, e.g., foraminifera, pollen, diatoms, testate amoebae and plant macrofossils (Appendices 1, 2 and 4). Back-barrier facies from Stage 1 (Figure 8.14 and Kraft *et al.*, 1987) include a Pleistocene hard substrate that is overlain by fluvial sediments (Figure 6.6), a basal freshwater and / or salt-marsh peat (Figures 5.4 and 5.5) and possibly reworked marsh and mudflat sediments (Figure 5.6). Lagoonal sequences (Reinson, 1984) comprise slate, shale, sandstone and gravel and minerogenic fines with marine shells (Figure 6.5). Laminated sediments are also found in facies indicative of inter-tidal lagoonal environments (Figure 6.4).

Sequences from Stage 2 (Figure 8.14) consist of intercalated salt-marsh facies in thick marine minerogenic sections (Figures 4.4 and 4.5), entire brackish marsh facies (Figure 7.13) and occurrences of thin supratidal minerogenic or organic lenses (Figure 7.14). Sand facies include washover sheets and channel-fill deposits (Figure 6.5 and Reinson, 1984) and those re-deposited in freshwater marsh facies (Figures 5.7 and 5.19). Sequences indicative of barrier closure are marine sand and gravel overlain by freshwater peat and aeolian derived sand, e.g., Bantham Sands, minerogenic fines overlain by freshwater peat at North Sands and terrestrial gravel at Slapton Sands, and organogenic peat overlain by minerogenic fines and gravel at Blackpool Sands. The late Holocene, Stage 3, facies comprise lacustrine silt and clay (*gyttja*) at Bantham and Slapton Sands, wetland peat marsh at North Sands and minerogenic sediments at Blackpool Sands. Dune facies are also found in late Holocene sequences at Bantham Sands and overwash gravel deposits in lacustrine sediments from Slapton Sands.

8.5 Methodological issues

This project has raised many methodological issues related to the careful evaluation of possible errors in the assignment of sea-level index point chronology and altitude. The primary aims of this project have been achieved by pursuing several objectives with a methods-based approach (section 1.3). From this, a number of significant advancements have been made and limitations encountered that require further discussion. This section presents the most important of those and offers suggestions for further improvement in obtaining as reliable a sea-level history as possible under current research techniques.

8.5.1 Autocompaction of sediments

Samples collected from unconsolidated or non-basal sediments that are used to contribute to the long-term sea level history of south Devon have been corrected for sediment autocompaction. Sea-level indicators have also been identified from sediments lying directly on top of hard substrate, i.e., basal samples. This use of 'corrected' and 'compaction-free' SLIPs is a primary advancement in the evaluation of vertical errors commonly found in sea-level chronologies. Previous discussion (section 8.1) has shown that Holocene SLIPs currently available for south Devon (e.g., Hails, 1975a,b; Morey, 1976), Cornwall (Healy, 1993, 1995), and a significant number of those available for the remainder of south-west Britain (Heyworth and Kidson, 1982), are from intercalated sediments. Resultant local and regional Holocene sea-level histories may therefore have plotted lower than they otherwise should.

Samples were corrected for autocompaction using the Paul and Barras (1998) model and both significant results and serious methodological issues arose from working with back-barrier core sediments. In organic-rich sequences vertical sediment displacement is almost 2.2 m (Figure 5.23) and up to 1 m in minerogenic-peat sequences (Figure 7.17) around 8 m and 2 m below current ground level respectively. Sequences containing large gravel particles, typical of back-barrier facies, produce the least amount of sediment consolidation, <0.2 m (Figure 5.26). Results around 0.2 m commonly include minerogenic facies within 1 m or so of the basal substrate, and significant increases in the rate of change occur at contacts with peat facies (Figure 5.23). These results show that uncorrected 'basal' (within a metre of the substrate) SLIPs are not likely to significantly alter the altitude of a sea-level curve upon correction, unless the facies is entirely basal peat. However, samples extracted more than 1 to 2 m above bedrock can raise SLIP heights as much as 0.2 to at least 2.2 m through thick Holocene sequences. Correction for autocompaction is especially important when rates of sea-level rise are slow, e.g., after 7000 cal years BP (Table 8.1). Many such uncorrected SLIPs from similar facies are to be found in the current literature for south-west England.

Methodological issues arising from this work have served to highlight some limitations despite the significant advancements made here. Problems occur when incomplete sedimentary sequences are extracted. This is generally due to the drilling system, skill of the crew and the nature of the sediments being worked with. Compaction of facies resulting from the percussion drilling technique, core loss (including non-retrieval of sediment) and withdrawal of heavily saturated sediments into air (de-watering) are prime examples of this. Models are currently not sophisticated enough to fully extrapolate values and are based on long-established geotechnical formulae themselves not currently tested by all sediment types. For example, pure organic peat routinely records high and wide ranging liquid (Atterberg, 1911) limits values and it is difficult to determine consolidation coefficients for this sediment type. Similarly, groundwater levels may have previously been deeper at sites resulting in greater values for effective stress and coefficients of consolidation. However, these factors will result in any correction applied to intercalated sequences being too low rather than too high but serve to bring samples closer to their original height of deposition.

8.5.2 Indicative meaning

Many of the sea-level index points currently available for south-west Britain have not accurately been related to a reference tidal level. Methodologies include the approximation of indicative meaning from pollen zones, archaeological remains, dendrochronologies, and plant fossils including submerged forest remains and horizons above or below washover facies (see Heyworth and Kidson, 1982). More commonly used sea-level indicators such as foraminifera, diatoms, testate amoebae and ostracods have been used by a number of authors to identify SLIPs but transfer functions are not often used to quantify indicative meaning. In fact, a number of SLIPs screened by the Sea-level Research Unit at the University of Durham (Environmental Research Centre, 2003) are assigned an approximate indicative meaning based on palaeoenvironmental information supplied by the original author (Horton, pers. comm. 2003). Haslett *et al.* (1997) suggest that authors not assigning indicative meaning to their samples cannot consider them as accurate sea-level index points.

This project has sought to show that employing a quantitative method of determining vertical SLIP errors provides the researcher with more reliability in the interpretation of indicative meaning. Healy (1993, 1995) determined indicative meaning in a qualitative way from pollen and diatoms identified in his Marazion Marsh and Trewornan samples. This thesis argues that methods similar to that employed by Horton *et al.* (1999), Edwards and Horton (2000), Edwards (2001) and Gehrels *et al.* (2001) offers a more precise determination than that of Healy (1993, 1995) and others. For this purpose, the Erme Estuary and Frogmore Creek have proved useful locations for relating modern foraminiferal analogues and fossil assemblages to a common tide level. This has been achieved over a significantly greater tidal range than previously available in the literature, and identifies four distinct zones from which an indicative meaning for those assemblages can be assigned with good precision. Horton (1999), Horton *et al.* (1999a,b), Edwards and

Horton (2000) have used mudflat assemblages to assign indicative meaning to UK-wide datasets but a significant gap in the data exists in south-west England. This project ably fills that gap with a contemporary dataset extending from –2.6 to +2.6 m MTL (Figure 3.3). The WA-PLS regression model (the most appropriate model to use for a gradient length = 3.6) determined $r^2 = 0.93$ for the dead assemblage and the transfer function is able to predict sea-level changes with very good precision (RMSEP = ±0.29 m and is the best estimate of vertical error from the model data). Results from this study are comparable to those of Horton *et al.* (1999a, see section 3.5).

If the palaeoenvironmental setting identified from core sedimentary facies is not replicated by the modern analogue then fossil faunal assemblages are likely to be different also. Although the contemporary inter-tidal dataset from Gehrels *et al.* (2001) has been extended to –2.6 m MTL in this project the number of samples collected below MTL is not as great as those obtained above it (Figure 3.3). Results from mudflat samples in Bantham Sands cores illustrate a potential problem with the quantification of indicative meaning lower down the inter-tidal frame. Values of –7 to –4 m MTL are common for the indicative meaning of samples containing very diverse calcareous assemblages (Figures 4.11 and 4.13). Total numbers are not low in the Frogmore Creek modern analogue samples, ranging from over 120 (around –2.6 m MTL) to 1000-3000 (between –2.6 m MTL and MTL) animals per sample. Species diversity per fossil sample is generally half that of modern samples and numbers are lower in the uppermost fossil samples. If some species of foraminifera, e.g., *E. macellum*, *E. crispum* and *A. mamilla*, are over-represented in fossil sediments compared with modern samples the transfer function will calculate an indicative meaning that is significantly below the range of contemporary assemblages (section 3.5). Increasing the resolution of contemporary samples below MTL would undoubtedly improve the modern analogue, given that the modern marine setting is similar to the palaeoenvironmental one. At the very least, the inferred sub-MTL indicative meaning is in the right direction for cores BA-97-1 and 2 but simply may not be as deep as those reconstructed from 'barrier-closure' facies in other cores.

8.5.3 Electrical resistivity surveying

Core sections extracted from back-barrier systems have assisted in the reconstruction of Holocene palaeoenvironments along the south Devon coastline. However, it is notable that the small number of boreholes taken from each site (2 to 4) leaves a significant amount of each cross-section unexplored. Electrical resistivity surveying was originally carried out to fill in these subsurface gaps but required ground-truthing from the boreholes themselves. The method is also proposed as a pilot survey technique for identifying thick Holocene sedimentary sequences that would provide the longest record of sea-level change. The potential for this technique is well known in civil engineering (Griffiths and King, 1981; Dahlin *et al.*, 1996; Bjulemar, 1997; Dahlin *et al.*, 1999; Airey and Bates, 2000) but not so commonly used for palaeoenvironmental investigations.

In general, lithologies with low resistivity have high conductivity such as water saturated zones, porous rock and marsh peat. Lithologies with high resistivity have low conductivity and contain little water such as very compacted clay, weathered rock (highly variable resistivity) and lithified rock. Factors affecting the determination of the resistivity of lithologies include porosity, permeability, water saturation, temperature and salinity (Hallenburg, 1984). It is clear that back-barrier systems may not offer the perfect environment in which to undertake such a survey although data generated in this project have revealed some useful trends. Figure 8.15 shows that the resistivity of peat and fine minerogenic sediments overlap to a greater extent than gravel and the finer sediments. The resistivity of common sediments and minerals is very variable, e.g., Milsom (1996) lists clay with values of 1-100 Ωm, gravel at 100-600 Ωm and weathered bedrock at 100-1000 Ωm. Kearey and Brooks (1984) gives approximate values for clay of 1-50 Ωm and as much as 50-10000 Ωm for shale. The practice of using median resistivity values to interpret the data is perhaps less widespread but may serve to identify sedimentary facies more easily in this project. The weathered bedrock surface is expected to have a median resistivity of around 100 Ωm, coarse sand and gravel approximately 60 Ωm, minerogenic fines about 45 Ωm and peat roughly 30 Ωm (Table 8.6).

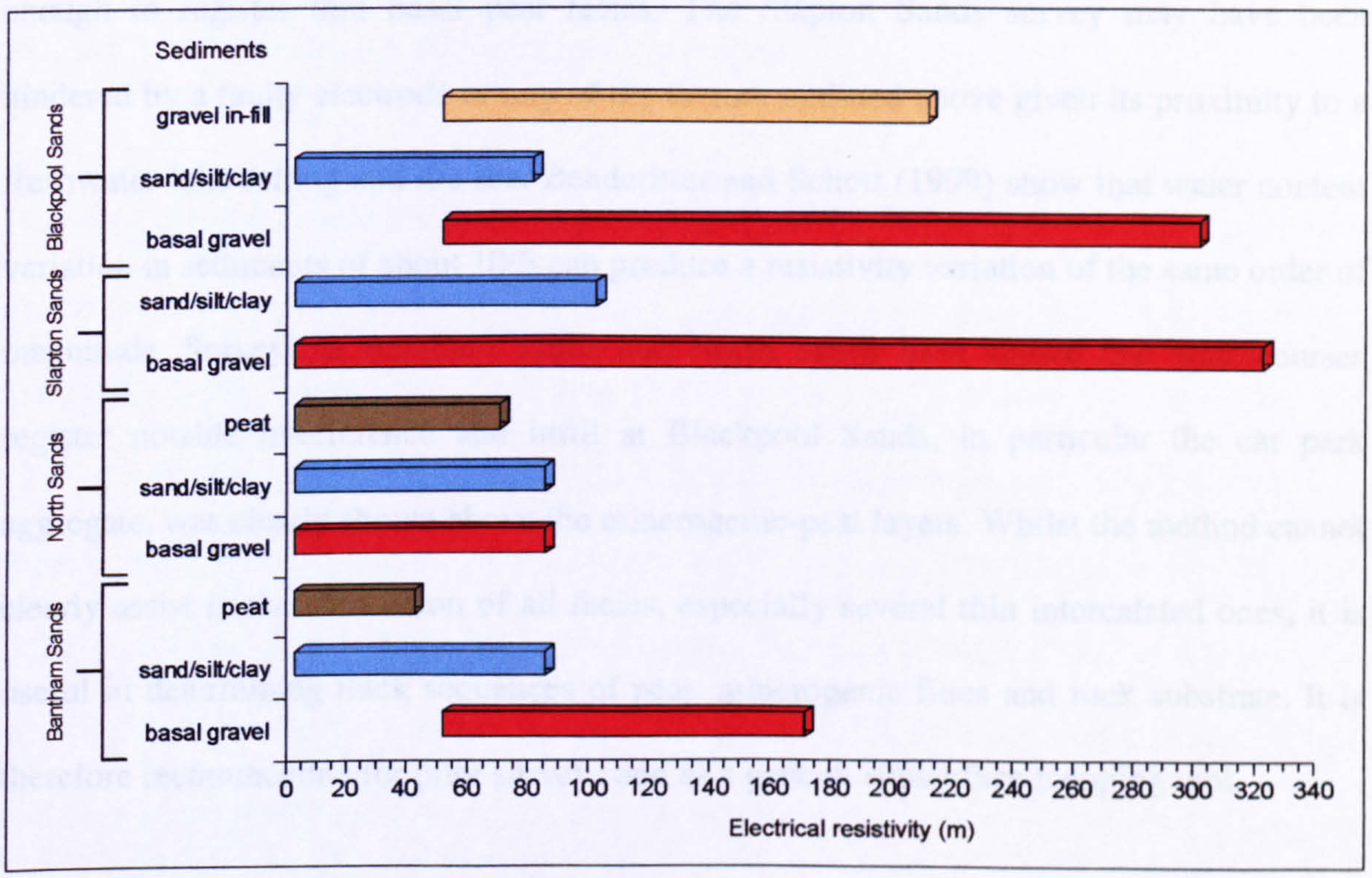


Figure 8.15. The resistivity ranges of back-barrier sediments from south Devon.

Site	Basal gravel (weathered bedrock)	Minerogenic fines (sand, silt, clay)	Peat	Barrier gravel	Gravel infill	Coarse sand and gravel
Bantham Sands	85	42	20	-	-	60
North Sands	42	42	34	-	-	47
Slapton Sands	160	50	-	727	-	-
Blackpool Sands	150	40	-	-	105	75

Table 8.6. Summary of the median electrical resistivity values at all sites (units in Ωm). Based on the ground-truth method using cores overlying shore-parallel, shore-normal and shore-diagonal surveys.

The method has been most useful in determining the approximate depth of the weathered bedrock surface and thus the deepest, and shallowest, Holocene sedimentary sequences. The vertical extent of the minerogenic fines has been more difficult to determine. Readings may have been affected by groundwater and salinity levels at some sites and where thin sand lenses can be found intercalated with peat and fines (Hallenburg, 1984). Peat sequences routinely record lower values but the survey equipment is not sophisticated enough to register thin basal peat facies. The Slapton Sands survey may have been hindered by a faulty electrode or any of the factors outlined above given its proximity to a freshwater lake setting and the sea. Benderitter and Schott (1999) show that water content variation in sediments of about 10% can produce a resistivity variation of the same order of magnitude. Surveys at Bantham Sands and North Sands have shown that watercourses register notable interference and infill at Blackpool Sands, in particular the car park aggregate, was clearly shown above the minerogenic-peat layers. Whilst the method cannot clearly assist in the correlation of all facies, especially several thin intercalated ones, it is useful in determining thick sequences of peat, minerogenic fines and rock substrate. It is therefore recommended for pilot surveys and as a general subsurface mapping tool.

8.5.4 Sea-level indicators

The quantification of indicative meaning and the reconstruction of coastal palaeoenvironments rely on the recognition of specific sea-level indicators. Foraminifera are commonly used protozoa for which standard methods have been developed to determine vertical error margins for SLIPs (Scott and Medioli, 1980a; van de Plassche, 1982). However, sedimentary sequences from south Devon show that back-barrier systems often contain very complex facies resulting from rapidly changing palaeoenvironments. In general, two main problems were identified during the investigation stages. Firstly, very low numbers (<5 per 5 ml) of foraminifera were counted in some crucial facies, i.e., the basal or upper contact. Secondly, there is speculation that some facies may have suffered displacement, e.g., Blackpool Sands cores (section 7.6). Evidence for some sediment reworking has been confirmed in the basal section of core NS-97-3 (section 5.7.2), and this mechanism could affect foraminiferal assemblages preserved in the sediments. Increasing the sampling resolution improved both the biostratigraphical reconstruction (e.g., Figure 7.13) and identified basal facies where a number of calcareous foraminifera were found intact (e.g., Figures 7.12 and 7.14). Sudden "jumps" in the indicative meaning of the sediments are indicative of discontinuous sequences.

Post-depositional alterations to foraminiferal assemblages have been discussed in studies of the surface and infaunal distribution of these animals (Scott and Medioli, 1980a; Goldstein, 1988; Goldstein and Harben, 1993; de Rijk, 1995; Massey, 1998; Saffert and Thomas, 1998), particularly the density and diversity of species assemblages. Holocene coastal environments around south Devon were very dynamic and salt marshes and mudflats existed in close proximity to rivers or freshwater lakes. Bacterial oxidation of agglutinates can occur in sediments drying out above the level of groundwater (Smith, 1987) or in the very high marsh, and may have taken place at North Sands and Blackpool Sands where numbers of animals change markedly across some facies. Ozarko *et al.* (1997) suggest that species such as *M. fusca* are particularly susceptible to bacterial degradation in fossil sediments but de Rijk (1995) proposes that *M. fusca*, *T. inflata* and *J. macrescens* have robust tests and are well preserved in cores. However, de Rijk (1995) also found that *J. macrescens* and other agglutinates are sensitive to compaction at depth. This is interesting to note because some specimens were found at varying stages of collapse in sequences in this project. Nonetheless, to what extent such conditions can cause the total breakdown of significant numbers of animals is not known.

Some basal sequences indicate that coastal environments evolved rapidly under quickly rising seas and transportation of tests may have occurred. This is a potential problem in index point 17 where calcareous foraminifera are found in very low numbers in what appears to be a freshwater marsh facies. Goldstein *et al.* (1995) suggested that preservation problems occur with both calcareous and agglutinated tests given the "marked environmental variability that typifies most salt marshes". Some modern environments have also revealed the remarkable adaptability of some species. Phleger and Walton (1950) and Culver and Buzas (1981) found *J. macrescens* in the marine bay and shelf environments whilst Matera and Lee (1972) discovered *T. inflata* on the tidal flats of Long Island, USA, strongly correlated with grain size. On the other hand, agglutinated species such as *P. ipohalina* may have disintegrated over time because their tests are comprised of loosely cemented, poorly sorted coarse grains (de Rijk, 1995). The more robust agglutinates do not generally suffer selective preservation of tests and thus do not hinder the palaeoenvironmental interpretation of biofacies. It is clear that whilst there is evidence that certain conditions may have adversely affected the post-depositional preservation of foraminifera there is much to suggest that this may not have been the case.

8.5.5 Identifying salt-marsh facies

Salt-marsh facies are normally identified in the lithostratigraphical record by the significant increase in organic content and darker colouring of the sediment. Marsh facies should be easily distinguishable from terrestrial, fresh-water or marine minerogenic facies at the Troels-Smith (1955) analysis stage of the coastal sediments. However, this was not the case with some cores in this project, e.g., NS-97-3 (Figure 5.6) and may be indicative of problems often encountered with back-barrier sediments in south Devon. Kraft (1971) and Belknap and Kraft (1985) identify a range of problems associated with the preservation of the lithostratigraphic record from barrier coastlines, e.g., reworking, fluvial erosion and overwash (section 8.4). It is interesting to note that Healy (1993, 1995) concluded the early Holocene marine environments of south Cornwall have left behind similarly complex lithostratigraphic records due to the rapidly changing nature of coastal barrier-lagoon systems.

In Start Bay, south Devon, marsh sediments have been eroded and reworked (Hails, 1975a) at the leading edge of the Holocene transgression thus are not well preserved in cores extracted from the offshore environment (Hails, 1975b). Belknap and Kraft (1981) also identified poor preservation of coastal marsh sediment in facies from back-barrier systems. However, some basal facies in this project do not appear to have suffered many of the problems identified by these authors. Morey (1976, 1983a,b) suggests that the basal facies of Slapton Sands represent an idealised sequence but that the barrier-closure organic phases are interrupted by washover facies (Burrin, 1982; Carter, 1982b; Jennings and Smyth, 1982). Carter (1982b) suggests that such events should be easily distinguishable in the sedimentary record and this is clearly seen in the barrier-closure facies of the Slapton Sands cores (Figure 6.7).

The sediments of core NS-97-3 were investigated for trends in carbon content (Figure 5.20) to identify the potential salt-marsh facies reconstructed from foraminiferal analyses (Figure 5.18) and justify the radiocarbon dating of those facies. The objective was to replace the foraminiferal sampling strategy with a more efficient method. Organic facies should be expected to contain more carbon than the minerogenic facies, e.g., de Rijk (1995) found the highest organic carbon percentages (20-30%) near the upland marsh and the lowest (2-9%) near the inter-tidal marsh. However, this is not the case in the reworked fossil basal sediments of core NS-97-3 where zones dominated by agglutinated foraminifera are very low in carbon content (<1%) and those dominated by calcareous species are higher (up to 4%). The method was less efficient than the foraminiferal sampling strategy and demonstrated that bulk sediment sampling is necessary to obtain sufficient carbon for AMS ^{14}C dating of sediment with low organic content.

Age reversals in the sedimentary sequence indicate that basal marsh deposits from core NS-97-3 have probably been reworked with older material from offshore (Figure 5.6). However, a probable "double sample" collected in error during coring is identified at the base of core NS-97-3 and this is reflected in the lithostratigraphical (Figure 5.6), biostratigraphical (Figures 5.18 and 5.29) and chronological (Figures 5.28 and 5.30) evidence from the sediments (section 5.7). Nonetheless, it would seem that the problems of reworking at the leading edge of the Holocene transgression (Belknap and Kraft, 1981, 1985) being shown in the fossil record (Hails, 1975a,b) have occurred at North Sands. Some irregularities in the composition of pollen assemblages indicate reworking of the sediment sequence in the basal section (reflected also in the double sample section) of core NS-97-3. Moreover, the pollen sequence indicates that basal and in-core sediment ages should be younger than *ca.* 8000 cal years BP (post-alder rise) but seven samples were between *ca.* 11,200 and 8200 cal years BP (Table 5.3). Further analysis shows that at least four of these samples are unacceptable as sea-level index points and the remainder from the core are considered borderline inclusions to south Devon's RSL history (section 5.7).

Chapter 9

Conclusions

9.1 Original aims

The aims of this Ph.D. thesis are to:

1. Establish a record of early and middle Holocene relative sea-level change for the south Devon coast.
2. Test current geophysical models of ice retreat and Earth rheology against the south Devon Holocene relative sea level history.
3. Calculate rates of late Holocene and ongoing crustal motion along the south Devon coastline.
4. Reconstruct the Holocene palaeoenvironments of the evolving south Devon coastline.

Objectives employed to achieve these aims have raised scientifically significant methodological issues that are summarised here. Sea-level index points generated from standard and relatively new techniques have been carefully evaluated for age and altitudinal errors. This involved a number of additional and refinement stages in the reconstruction process. Steps taken to improve the reliability of sea-level data obtained from fossil core records have resulted in conclusions that should be of interest to the wider scientific community and, in some cases, the general public.

9.2 Original hypotheses

Data derived from this Ph.D. thesis are used to test the following two hypotheses (each followed by the result):

1. The Holocene RSL history for south Devon is in agreement with existing data from the Holocene SLIP database for south-west Britain, and rates of crustal motion calculated

from new SLIPs along the south Devon coast are similar to those calculated by Shennan and Horton (2002) from the existing SLIP database.

The early and mid-Holocene data from south Devon agree closely with existing SLIPs from south Devon and Cornwall and rates of crustal motion are at least 0.07 m/ka lower than the best estimate of Shennan and Horton (2002).

2. Geophysical models are reliable predictors of Holocene RSL change in south Devon, but Peltier's model, which predicts a glacio-isostatic forebulge, is in closer agreement with the geological data than Lambeck's model.

Models are in close agreement with early Holocene data but overestimate the middle to late Holocene RSL history. Peltier's model is the better predictor of the general trend in Holocene RSL from robust empirical data and Lambeck's model from all south Devon SLIPs. A collapsing forebulge mechanism is invoked south of the former British Ice Sheet.

9.3 Holocene sea-level changes in south Devon

Thirty-three SLIPs were obtained from cores extracted at Bantham Sands, North Sands, Slapton Sands and Blackpool Sands. Twenty-four provide a reliable early to mid-Holocene RSL history and the remaining SLIPs are subject to irreconcilable age or height errors. The "accepted" data improve the current Holocene SLIP database (2003) for the south-west peninsula and southern Britain by 23% and increase south Devon's database almost fivefold. Sea level has risen 13 m along the south Devon coastline between *ca.* 9000 and 4400 cal years BP. The data agree closely with index points from Hails (1975a,b) and Morey (1976) in south Devon and Healy (1993, 1995, 1996) in west Cornwall. Early Holocene RSL rates are ~5.4 ± 2.1 m/ka and mid- to late Holocene rates are ~1.1 ± 0.2 m/ka and are in close agreement with those from the literature. RSL profiles produced from south Devon and west Cornwall SLIPs include an undulating one similar to Heyworth and Kidson (1982) and Jennings *et al.* (1998), and a smooth curvilinear one that is similar to Clarke (1970), Hawkins (1971, 1979), Morey (1976), Healy (1995) and others.

9.4 Testing the geophysical model predictions for south Devon

Lambeck's (1997) local predictions from his (1993a,b, 1995) GB-3 model are in close agreement with the early Holocene empirical data (*ca.* 9000 to 7800 cal years BP) but overestimate mid- to late Holocene RSL by as much as 3 to 5.5 m (*ca.* 7000 to 4500 cal years BP). Lambeck's (2000) modifications have increased the misfits during the early Holocene but improved predictions during the mid- to late Holocene. Peltier's (2000) local predictions from his (1998) ICE-4G model underestimate early Holocene RSL by up to 3 m and overestimate the mid- to late Holocene data by a similar amount. Peltier's (2002) modifications have improved the fit during the early Holocene but overestimated mid- to late Holocene RSL further. Both Lambeck and Peltier underestimate significantly basal SLIPs from core NS-97-3 but Lambeck (1997) agrees very closely with the uppermost SLIPs from that core. The predictions that best describe the disagreement between the models *ca.* 6700 cal years BP is Peltier (2002). Overall, Lambeck (2000) predicts closely the RSL profile of all "accepted" SLIPs from south Devon. The profile generated by ICE-4G (Peltier, 1998) is closest to that produced by the more robust empirical data. Decreasing the rate of ice melting from *ca.* 7000 cal years BP onwards and continuing the melting for times subsequent to *ca.* 4000 cal years BP would improve the fit during the mid- to late Holocene (Peltier, pers. comm. 2003). In general, comparisons with the modelled data support a collapsing forebulge mechanism along the coastline of south-west England.

9.5 Crustal movement in south Devon

The south Devon SLIP database published by Shennan and Horton (2002) is increased from a total of 5 to 29 sea-level index points. Their (rejected) limiting dates and north

Devon SLIP (No. 86) are excluded from the total contribution to rates of crustal movement in south Devon. The data, SLIP 31 from Blackpool Sands and published SLIPs younger than *ca.* 4500 cal years BP (Hails, 1975a,b; Morey, 1976), produce a net rate of relative land subsidence of –0.64 to –1.14 m/ka and a best estimate of on-going late Holocene crustal subsidence of –1.16 m/ka. This is in close agreement with the rate of land subsidence for south Devon of –1.23 m/ka from Shennan and Horton (2002). The best estimate of crustal subsidence is reduced further to –1.12 m/ka if recent global sea-level rise (UKCIP, 2004) is taken into consideration. A second order polynomial fit to the more robust SLIPs from this project could produce a lower estimate of on-going late Holocene coastal subsidence.

9.6 The Holocene coastal evolution of south Devon

Data agree with the findings of Clarke (1970), Hails (1975a,b), Morey (1976, 1983a) and others that a barrier-lagoon complex existed off the present-day coast of south Devon during pre- and early Holocene times. Minerogenic sediment filled south Devon's river channels and freshwater marsh and salt marsh bordered shallow marine lagoons *ca.* 9000 to 7800 cal years BP. Salt marsh formed offshore of the North Sands coastline at least 2000 to 3000 years earlier. Shallow marine conditions dominated by *ca.* 8100 cal years BP (below –13.7 m OD) to *ca.* 6800 cal years BP (at –6.15 m OD). Inter-tidal conditions existed at the sites until at least 4400 cal years BP. A back-barrier lake (sediments identified in pilot cores) developed at Bantham Sands before it filled in with freshwater peat and sand from the nearby dune system. A freshwater lake and marsh developed behind Slapton Sands, and Blackpool Sands basin filled in with silts and clays. Barrier-closure occurred at Blackpool Sands *ca.* 4400 cal years BP and a freshwater peat marsh formed at North Sands around the same time. Barrier-closure of lagoons and freshwater bodies was probably complete along south Devon's coastline by *ca.* 3000 cal years BP (Morey, 1976). Slapton Sands is in Stage 2 of the Holocene transgressive model of barrier movement (adapted from Kraft and Chrzastowski, 1985) whilst Bantham Sands, North Sands and Blackpool Sands have reached Stage 3. Future rapid sea-level rise (UKCIP, 2004) will return all systems to Stage 1. The identification of back-barrier environments from facies in this project contributes significantly to findings from coastal models developed elsewhere in the Northern Hemisphere.

9.7 The electrical resistivity of back-barrier sediments

Weathered bedrock has a resistivity of roughly 100 Ωm, coarse sand and gravel approximately 60 Ωm, fine minerogenic sediment about 45 Ωm and about 30 Ωm for organic peat. The surface 9 m of barrier gravel at Slapton Sands recorded 727 Ωm, surface gravel infill at Blackpool Sands 105 Ωm, and around 60 Ωm for intercalated coarse sand and gravel facies at all sites except Slapton Sands. The positioning of core NS-97-4 on the shore-parallel survey is in doubt as the weathered bedrock facies has a resistivity of 42 Ωm, a result expected for fine minerogenic facies. Surveys carried out at Bantham Sands and North Sands show that watercourses produce significant interference and similarly the proximity of lake and sea to the Slapton Sands barrier. Results show that the method can usefully distinguish between the weathered rock substrate and finer sediments.

9.8 Autocompaction of back-barrier sediments

The autocompaction of back-barrier sediments was calculated using the Paul and Barras (1998) geotechnical model. Fine minerogenic sediments in core NS-97-2 are vertically displaced by 0.07 m near the base and 1.1 m at the minerogenic-peat contact. This increased to a maximum of 2.2 m in the overlying minerogenic-peat sequence below the contact with organic peat. Basal sediments in core NS-97-3 have been displaced 0.04 m above the base and 0.2 m at the contact between gravel and fine-minerogenic-peat. Above the base of core BS-97-2, minerogenic-gravel has been displaced by 0.02 m to almost 1 m in the overlying minerogenic-peat facies. Results in core BS-97-3 ranged from 0.04 m to

>0.7 m in the uppermost sediments. Pure organic peat facies have liquid limits values >200%, outside the usual range for the coefficient of compression index, resulting in corrections being too low for these facies. Results show that current researchers are underestimating the effect of autocompaction on sub-surface sediments.

9.9 A new foraminiferal training set for sea-level reconstruction in south-west England

Surface sediments from the Erme estuary and Frogmore Creek in the Salcombe-Kingsbridge estuary were sampled for foraminifera. The contemporary data from this project, Gehrels *et al.* (2001) and White (2001) provide an inter-tidal (–2.6 to +2.6 m MTL) transfer function for south-west England from which sea-level changes are predicted with high precision (RMSEP = ±0.29 m). The dead assemblage provides the most powerful predictor of indicative meaning (r^2 = 0.93) and the WA-PLS regression model is the most appropriate to use (gradient length = 3.6). *Haplophragmoides* spp., prefers the uppermost 1.5 m of the inter-tidal frame, *Reophax* spp., *E. scabra* and *E. oceanensis* lives around MTL and *B. elegantissima*, *E. margaritaceum* and *T. angulosa* below MTL. *J. macrescens* and *M. fusca* dominate between MTL and +2.6 m MTL and *A. beccarii* and *H. germanica* between –2.6 and +1 m MTL. *T. inflata* is a useful indicator of environments above +1.2 m MTL.

9.10 The problems with back-barrier sediments

9.10.1 Sea-level indicators

Thirteen AMS ^{14}C ages were determined from sediment containing <30 foraminifera but eight of those were from cores NS-97-3 and BS-97-1, providing all but one (core NS-97-4) of the rejected SLIPs. All samples dated by AMS ^{14}C were supported by contiguous (1 cm) sampling that produced consistent numbers of animals throughout the facies. Many of the sediment samples contained gravel and the animals are likely to have been partially or fully broken down during post-mortem transport, e.g., *J. macrescens*. The possible dissolution of calcareous tests is a serious consideration at the interpretative stage (Green *et al.*, 1992, 1993; Scott and Medioli, 1978) but sufficient evidence, e.g., large numbers of test linings, was not found to suggest that this had taken place. The re-deposition of foraminifera at the edge of the palaeo-marsh is likely to account for the anomalous indicative meaning of index point 17 and foraminifera have been displaced in the basal facies of core BS-97-1. Samples containing low numbers in the modern training set were excluded from quantitative analyses. Basal peat samples from North Sands did not contain significant numbers of testate amoebae.

9.10.2 Identifying and radiocarbon dating fossil salt-marsh facies

Back-barrier sediments from cores in this project are generally characterised by high minerogenic and low organic content. Carbon results from core NS-97-3 did not clearly identify differences in organic content between salt marsh and mudflat facies reconstructed from foraminiferal analyses in predominantly gravel sediment. Basal zones dominated by agglutinated foraminifera contained <1% carbon and those containing calcareous species <4% carbon. Around 9 to 17% carbon was recorded in the upper minerogenic-peat and organic peat facies where no gravel was found. These results agree more closely with de Rijk's (1995) findings that the highest organic carbon percentages (20-30%) are found near the upland marsh and lowest (2-9%) near the inter-tidal marsh. The method is less efficient than the foraminiferal sampling strategy explained in chapter 2 but demonstrated that sediment with low organic content must be sampled in bulk to obtain sufficient carbon for AMS ^{14}C dating. Sediments from core NS-97-3 contain older carbon (*ca.* 11,200 to 9000 cal years BP) and age reversals occur throughout the sequence. A probable "double sample" of the basal section, collected in error during coring, is also identified from the lithostratigraphical, biostratigraphical and chronological analyses of core NS-97-3. Age/depth plots show that some SLIPs from this core are acceptable for sea-level studies.

However, the long- and short-term RSL chronology and salt marsh accumulation rates cannot be calculated reliably from this core. Pollen analyses indicate some sediment reworking and that the basal sediments should be younger than *ca.* 8000 cal years BP (post-alder rise). The AMS ^{14}C dating of bulk sediment from basal gravel sequences is therefore not recommended.

9.11 Future work

This thesis has achieved the aims stated in sections 1.2 and 1.4.3 but areas for further research remain. Only three late Holocene (last 4000 cal years BP) SLIPs have been produced from south Devon. Recent investigations by the author have shown that an indicative meaning can be obtained from Thurlestone Sands back-barrier (Figure 1.12) sediments that may produce local SLIPs of mid- to late Holocene age. A higher resolution RSL history would improve estimates of crustal movement and test geophysical models more rigorously during that period. Increasing the sampling resolution of the foraminiferal training set below MTL would improve the reconstruction of indicative meaning for fossil inter-tidal samples. A geotechnical analysis of peat sediments is possible and would enhance the model output. The radiocarbon dating of sediments from the upper marine-freshwater interface would identify ages of barrier-closure and further elucidate theories of Holocene coastal evolution in south Devon. Furthermore, the Thurlestone system may have reached Stage 3 of the coastal evolution model much later than sites investigated in this project.

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Appendices

Contents (CD-ROM):

APPENDIX 1 LITHOSTRATIGRAPHIC RECORDS

1a Sediment description of fossil coastal sediments using Troels-Smith (1955)

1b Key to core logs (adapted from Troels-Smith, 1955)

1c Site core logs (1 m sections)

1d Photographic record of core logs (entire cores and sub-sections)

APPENDIX 2 FOSSIL FORAMINIFERA RECORDS

2a Raw and percentage counts for all foraminiferal assemblages

APPENDIX 3 CARBON RESULTS

3a Carbon percentage values for core NS-97-3

APPENDIX 4 POLLEN RESULTS

4a Raw and percentage counts for pollen assemblages from core NS-97-3

APPENDIX 5 GEOTECHNICAL DATA

5a Geotechnical theory and standard methods

5b Laboratory results

5c Model results

APPENDIX 6 GEOPHYSICAL MODELLING DATA

6a Numbers for geophysical modelling runs by Lambeck (pers. comm. 1997, 2000) and Peltier (pers. comm. 2000, 2002)

APPENDIX 7 MODERN ANALOGUE DATA

7a Raw and percentage counts of live and dead foraminifera