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RECONSTRUCTING HOLOCENE CLIMATE CHANGE IN SCOTLAND UTILISING PEAT STRATIGRAPHY AND TEPHROCHRONOLOGY

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ABSTRACT

FACULTY OF SCIENCE

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RECONSTRUCTING HOLOCENE CLIMATE CHANGE IN SCOTLAND UTILISING PEAT STRATIGRAPHY AND TEPHROCHRONOLOGY

by **Peter Guy Langdon**

Changes in Holocene palaeoclimates in Scotland have been reconstructed from seven sites: six ombrotrophic raised mires (Longbridge Moss, Langlands Moss, Temple Hill Moss, Shirgarton Moss, Mallachie Moss, and Craigmaud Moss) and one ombrotrophic blanket mire (Ben Gorm Moss) located across geographical and climatological gradients in Scotland.

The main techniques used for palaeoclimatic reconstruction were plant macrofossil, colorimetric humification, and testate amoebae analyses, which were supported by a radiocarbon based chronology, aided by analyses of spheroidal carbonaceous particles and pine pollen at specific sites. Field stratigraphy was undertaken at each site in order to show that the changes detected within the peat profiles are replicatable. In an attempt to identify synchroneity of climatic phases, tephra isochrones were utilised to allow the precise linking of time-spans between sites. Fourteen tephras were identified from the seven sites, and electron microprobe analysis has been used to geochemically type the tephras to known isochrones from Icelandic eruptions.

Detrended correspondence analysis (DCA) has been used to reconstruct a climatic proxy record from the plant macrofossil data. The testate amoebae data were subjected to a transfer function which produces quantitative reconstructed mire water table values using a calibration program (WACALIB). The results show coherent wet and dry phases over the last *ca.* 5000 years (*ca.* 7500 years at Temple Hill Moss), with the three proxy reconstructions supporting each other extremely well. Significant wet phases were identified at *ca.* cal. AD 1350-1550, AD 870-1150, AD 500-600, 750-940 BC, 1240-1390 BC, 1450-1630 BC, 1700-1900 BC, and *ca.* cal. 2200-2300 BC. Additional wet phases were recognised at Temple Hill Moss at *ca.* cal. 3350 BC, 3900 BC, and *ca.* cal. 4700 BC. Significant dry phases were documented at *ca.* cal. AD 920-1340, AD 350-520, and *ca.* cal. 540-390 BC. Correlations at the time of the Glen Garry and Hekla-4 isochrones revealed a significant difference between climatic phases in the north and south of Scotland, highlighting the possibility of asynchronous changes between northern Scotland and the rest of Britain.

The local extinctions of *Sphagnum imbricatum* in the four sites where it previously grew occurred at different dates, but were all associated with a wet phase.

Spectral analyses of the DCA and humification data revealed identical significant periodicities in both sets of data from four sites, either between 520-580 years or around 280 years. A millennial scale periodicity (1100 years) was identified at Temple Hill Moss, which may be related to an oceanic climate forcing mechanism.

Table of Contents.

Abstract........i

Table of Contents........ii

List of Tables........ix

List of Figures........xi

List of Plates........xvii

Acknowledgements........xviii

Chapter 1. Introduction........1

- 1.1 Background........1
- 1.2 Thesis structure........3

Chapter 2. Previous research........4

- 2.1 Introduction........4
- 2.2 Factors influencing the climatic signal on ombrotrophic peatlands........5
  - 2.2.1 Internal hydrological controls........5
  - 2.2.2 Artificial raising of the water table........7
  - 2.2.3 Anthropogenic disturbances........8
- 2.3 Climatic reconstructions from peat stratigraphy........10
  - 2.3.1 Introduction........10
  - 2.3.2 Peatland pioneers and the establishment of a climatic proxy........11
  - 2.3.3 Rejuvenated interest in peat stratigraphy........14
- 2.4 Holocene climatic changes in Scotland........23
  - 2.4.1 Peat stratigraphy........23
  - 2.4.2 The late-Holocene history of *Pinus sylvestris*: coupled with climatic change and blanket peat expansion?........28
  - 2.4.3 Other evidence for late-Holocene climate change........34
- 2.5 Forcing mechanisms........36
  - 2.5.1 Solar variability........37
  - 2.5.2 Volcanic activity........39
- 2.6 Synthesis - changes in NW Europe and Scotland........41
- 2.7 Conclusions........45

Chapter 3. Methods 48
3.1 Introduction 48
3.2 Tephrochronological approach 48
3.2.1 Introduction 48
3.2.2 Laboratory procedures 51
3.2.2.1 Ashing 51
3.2.2.2 Electron microprobe slide preparation 52
3.2.2.3 Microprobe operating procedure 53
3.3 Site selection and coring strategy 55
3.3.1 Coring strategy 56
3.4 Plant macrofossils 58
3.4.1 Introduction 58
3.4.2 Laboratory procedures 61
3.5 Humification 63
3.5.1 Introduction 63
3.5.2 Colorimetric analysis of peat humification 66
3.5.3 Near infrared reflectance spectroscopy (NIRS) 67
3.6 Testate amoebae analysis 68
3.6.1 Introduction 68
3.6.2 Testate identification 70
3.6.3 Methodology 70
3.7 Spheroidal carbonaceous particle (SCP) analysis 71
3.7.1 Methodology 72
3.8 Radiocarbon dating 73
3.9 Conclusions 77
Chapter 4. Site descriptions 78
4.1 Introduction 78
4.2 Longbridge Moss (LBM) - NY 3053 5694 84
4.2.1 Climate data 87
4.3 Langlands Moss (LAG) - NS 2634 6513 87
4.3.1 Climate data 90
4.4 Temple Hill Moss (TEM) - NT 3113 6613 91
4.4.1 Climate data 91

4.5 Shirgarton Moss (SGM) - NS 2645 6963 94
    4.5.1 Climate data 97
4.6 Mallachie Moss (MAL) - NH 2963 8170 98
    4.6.1 Climate data 98
4.7 Craigmaud Moss (CMD) - NJ 3885 8588 101
    4.7.1 Climate data 104
4.8 Ben Gorm Moss (BGM) - NG 1432 8658 104
    4.8.1 Climate data 105

Chapter 5. Chronology: tephrochronology and radiocarbon dating 109

5.1 Introduction 109
5.2 Tephrochronology 109
    5.2.1 Ashing results 109
    5.2.2 Subsampling of tephra layers 114
    5.2.3 Geochemical results 118
        5.2.3.1 Introduction 118
        5.2.3.2 Longbridge Moss 128
        5.2.3.3 Langlands Moss 129
        5.2.3.4 Temple Hill Moss 130
        5.2.3.5 Shirgarton Moss 133
        5.2.3.6 Mallachie Moss 134
        5.2.3.7 Craigmaud Moss 136
        5.2.3.8 Ben Gorm Moss 139
    5.2.4 Geochemical and spatial variability of tephra distribution 147
    5.2.5 Tephrochronological summary 152
5.3 Radiocarbon dating 153
    5.3.1 Longbridge Moss 155
    5.3.2 Langlands Moss 156
    5.3.3 Temple Hill Moss 157
    5.3.4 Shirgarton Moss 157
    5.3.5 Mallachie Moss 158
    5.3.6 Craigmaud Moss 158
    5.3.7 Ben Gorm Moss 159
5.4 Conclusions 159

Chapter 6. Results....................168
- 6.1 Components of the macrofossil diagrams....................168
  - 6.1.1 UOM....................169
  - 6.1.2 Ericaceae....................170
  - 6.1.3 Monocots undiff....................171
  - 6.1.4 *Eriophorum vaginatum*....................171
  - 6.1.5 *Eriophorum angustifolium*....................173
  - 6.1.6 *Rhynchospora alba*....................174
  - 6.1.7 *Scirpus cespitosus*....................174
  - 6.1.8 *Odontoschisma sphagni*....................175
  - 6.1.9 *Aulacomnium palustre*....................175
  - 6.1.10 *Dicranum scoparium*....................175
  - 6.1.11 *Drepanocladus fluitans*....................175
  - 6.1.12 *Hypnum cupressiforme*....................176
  - 6.1.13 *Pleurozium schreberi*....................176
  - 6.1.14 *Pohlia nutans*....................176
  - 6.1.15 *Polytrichum juniperum/alpestre* type....................177
  - 6.1.16 *Racomitrium lanuginosum*....................177
  - 6.1.17 *Rhytidiadelphus loreus*....................178
  - 6.1.18 Identifiable *Sphagnum*....................178
  - 6.1.19 *Sphagnum* section *Acutifolia*....................179
  - 6.1.20 *Sphagnum imbricatum* spp. *austinii*....................182
  - 6.1.21 *Sphagnum magellanicum*....................184
  - 6.1.22 *Sphagnum papillosum*....................185
  - 6.1.23 *Sphagnum* section *Cuspidata*....................186
  - 6.1.24 *Cenococcum grandiforme*....................187
  - 6.1.25 Charcoal....................188
- 6.2 Peat humification....................191
- 6.3 Testate amoebae....................192
- 6.4 Detrended Correspondence Analysis (DCA)....................196
- 6.5 Testate amoebae transfer function....................202

6.6 Longbridge Moss........204
- 6.6.1 Stratigraphy........204
- 6.6.2 Plant macrofossils........205
- 6.6.3 Humification........207
- 6.6.4 Testate amoebae........209
- 6.6.5 DCA........212
- 6.6.6 Testate amoebae transfer function........212
- 6.6.7 Comparison of the three proxy techniques........213

6.7 Langlands Moss........224
- 6.7.1 Stratigraphy........224
- 6.7.2 Plant macrofossils........224
- 6.7.3 Humification........227
- 6.7.4 Testate amoebae........227
- 6.7.5 DCA........230
- 6.7.6 Testate amoebae transfer function........230
- 6.7.7 Comparison of the three proxy techniques........231

6.8 Temple Hill Moss........241
- 6.8.1 Stratigraphy........241
- 6.8.2 Plant macrofossils........241
- 6.8.3 Humification........244
- 6.8.4 Testate amoebae........245
- 6.8.5 DCA........246
- 6.8.6 Testate amoebae transfer function........247
- 6.8.7 Comparison of the three proxy techniques........247

6.9 Shirgarton Moss........257
- 6.9.1 Stratigraphy........257
- 6.9.2 Plant macrofossils........257
- 6.9.3 Humification........259
- 6.9.4 Testate amoebae........259
- 6.9.5 DCA........260
- 6.9.6 Testate amoebae transfer function........261
- 6.9.7 Comparison of the three proxy techniques........262

6.10 Mallachie Moss....272
    6.10.1 Stratigraphy....272
    6.10.2 Plant macrofossils....272
    6.10.3 Humification....274
    6.10.4 Testate amoebae....275
    6.10.5 DCA....276
    6.10.6 Testate amoebae transfer function....277
    6.10.7 Comparison of the three proxy techniques....277
6.11 Craigmaud Moss....287
    6.11.1 Plant macrofossils....287
    6.11.2 Humification....288
    6.11.3 DCA....288
    6.11.4 Comparison of the two proxy techniques....289
6.12 Ben Gorm Moss....296
    6.12.1 Plant macrofossils....296
    6.12.2 Humification....297
    6.12.3 Comparison of the two analyses....297
6.13 Time series analysis....301
6.14 Summary....305
Chapter 7. Discussion....311
7.1 Climatic change....311
    7.1.1 Climatic phases and ecological thresholds....311
    7.1.2 Coherent phases of change between the peat records....314
    7.1.3 Tephra isochrones - precise correlations....320
    7.1.4 Summary of the palaeoclimatic reconstructions....326
7.2 Comparison and value of the three proxy records....327
    7.2.1 Comparison of plant macrofossil DCA....327
    7.2.2 Comparison of humification records....332
    7.2.3 Comparison of testate amoebae reconstructions....332
    7.2.4 Summary of the three proxies....335
7.3 The role of climate change in species replacement....335
    7.3.1 The role of *Sphagnum imbricatum* in peatland development and climatic change....335

7.3.2 The decline of *Amphitrema wrightianum* 343

7.4 Chronologies and periodicities 345

7.4.1 Oceanicity vs continentality: site sensitivity 347

Chapter 8. Conclusions and future research 349

8.1 Future research 351

Appendix 354

References 364

List of Tables.

Table 2.1 Comparison of the timing of climatic deteriorations inferred from mire wet shifts at a selection of sites across Europe (*ca.* cal. years AD/BC). Adapted from Hughes *et al.* (in press)......42

Table 3.1 Sites investigated, based on a southwest to northeast gradient......57

Table 3.2 ^{210}Pb dates (means) for the three SCP dating features (after Rose *et al.*, 1995)......72

Table 5.1 Mid to late Holocene tephras which have been identified in northern Britain and Ireland and which have published geochemistries......126

Table 5.2 Longbridge Moss LBMT-2: Hekla-4......128

Table 5.3 Langlands Moss LAGT-1:'AD 860 Layer'......129

Table 5.4 Langlands Moss LAGT-2: Glen Garry......129

Table 5.5 Temple Hill Moss TEMT-1: Glen Garry......131

Table 5.6 Temple Hill Moss TEMT-2: Lairg layer (A)......131

Table 5.7 Shirgarton Moss SGMT-2: Glen Garry......133

Table 5.8 Mallachie Moss MALT-1: Glen Garry......134

Table 5.9 Mallachie Moss MALT-2: Hekla-4......135

Table 5.10 Craigmaud Moss CMDT-1: Glen Garry......137

Table 5.11 Craigmaud Moss CMDT-2: Hekla-4......137

Table 5.12 Ben Gorm Moss BGMT-1 (31cm)......139

Table 5.13 Ben Gorm Moss BGMT-2: Glen Garry......140

Table 5.14 Ben Gorm Moss BGMT-3 (81cm)......140

Table 5.15 Ben Gorm Moss BGMT-4: Hekla-4......141

Table 5.16 Summary of the tephra isochrones identified and used in this study......147

Table 5.17 Radiocarbon dates (37) with the full calibrated 2 sigma range......154

Table 6.1 Mean water depths of the species identified in this study on ombrotrophic bogs, from selected references......190

Table 6.2 Moisture classes of testate amoebae......192

Table 6.3 Taxa found in this study based on the taxonomy of Charman *et al.* (in press) and their hydrological requirements (modified from Hendon, 1998)......193

Table 7.1 Wet phases for each site recorded by at least two of the proxy indicators from the site synthesis....317

Table 7.2 Dry phases for each site recorded by at least two of the proxy indicators from the site synthesis....317

Table 7.3 The dominant taxa from testate amoebae analysis at each site....333

Table 7.4 The dates of *Sphagnum imbricatum* growth and extinction from four sites in southern and central Scotland....337

Table 7.5 The dates of the *Amphitrema wrightianum* decline from four sites in Scotland...344

Table 7.6 The periodicities in years discovered from spectral analyses of the humification and DCA axis 1 scores at five raised bogs in Scotland....346

List of Figures.

Figure 2.1 Spectrum of climatic variance (after Mitchell, 1976: with adaptations by Overpeck, pers. comm.)......37
Figure 4.1 Location of sites and climatic sub-regions (adapted from Birse, 1971)......79
Figure 4.2 Annual precipitation for all sites......81
Figure 4.3 Mean annual temperatures for all sites......82
Figure 4.4 Map of Longbridge Moss......85
Figure 4.5 Water balance for Longbridge Moss......87
Figure 4.6 Map of Langlands Moss......88
Figure 4.7 Water balance for Langlands Moss......91
Figure 4.8 Map of Temple Hill Moss......92
Figure 4.9 Water balance for Temple Hill Moss......94
Figure 4.10 Map of Shirgarton Moss......95
Figure 4.11 Water balance for Shirgarton Moss......97
Figure 4.12 Map of Mallachie Moss......99
Figure 4.13 Water balance for Mallachie Moss......101
Figure 4.14 Map of Craigmaud Moss......102
Figure 4.15 Water balance for Craigmaud Moss......104
Figure 4.16 Map of Ben Gorm Moss......106
Figure 4.17 Water balance for Ben Gorm Moss......108
Figure 5.1a Tephra shard abundance at Longbridge Moss......111
Figure 5.1b Tephra shard abundance at Carsegowan Moss......111
Figure 5.1c Tephra shard abundance at Shirgarton Moss......111
Figure 5.1d Tephra shard abundance at Kilhorn Moss......111
Figure 5.1e Tephra shard abundance at Temple Hill Moss......112
Figure 5.1f Tephra shard abundance at Langlands Moss......112
Figure 5.1g Tephra shard abundance at Dalmellington Moss......112
Figure 5.1h Tephra shard abundance at Red Moss of Balerno......112
Figure 5.1i Tephra shard abundance at Mallachie Moss......113
Figure 5.1j Tephra shard abundance at Blar-nam-Fiadh Moss......113
Figure 5.1k Tephra shard abundance at Ben Gorm Moss......113

Figure 5.1l Tephra shard abundance at Craigmaud Moss..113
Figure 5.2a Distribution of Hekla-4 tephra in 1cm peat slices from Longbridge Moss....... 119
Figure 5.2b Distribution of Glen Garry tephra in 1cm peat slices from Shirgarton Moss....119
Figure 5.2c Distribution of AD 860 tephra in 1cm peat slices from Langlands Moss......... 120
Figure 5.2d Distribution of Glen Garry tephra in 1cm peat slices from Langlands Moss.... 120
Figure 5.2e Distribution of Glen Garry tephra in 1cm peat slices from Temple Hill Moss. 121
Figure 5.2f Distribution of Lairg layer tephra in 1cm peat slices from Temple Hill Moss.. 121
Figure 5.2g Distribution of Glen Garry tephra in 1cm peat slices from Mallachie Moss.... 122
Figure 5.2h Distribution of Hekla-4 tephra in 1cm peat slices from Mallachie Moss..........122
Figure 5.2i Distribution of Glen Garry tephra in 1cm peat slices from Craigmaud Moss... 123
Figure 5.2j Distribution of Hekla-4 tephra in 1cm peat slices from Craigmaud Moss.........123
Figure 5.2k Distribution of Beinn Eighe(?) tephra in 1cm peat slices from
Ben Gorm Moss.. 124
Figure 5.2l Distribution of Glen Garry tephra in 1cm peat slices from Ben Gorm Moss.... 124
Figure 5.2n Distribution of Snaefellsjokull(?) tephra in 1cm peat slices from
Ben Gorm Moss.. 125
Figure 5.2m Distribution of Hekla-4 tephra in 1cm peat slices from Ben Gorm Moss........125
Figure 5.3 A diagrammatic summary of 14 profiles showing the tephras at each site and
an idealised sequence for Scotland as a whole (after Dugmore *et al.*, 1995a).......................127
Figure 5.4 Ternary diagram showing the Fe, Ca and K data from LAGT-1 and LAGT-2...130
Figure 5.5 Ternary diagram of the Fe, Ca and K geochemical data from
Temple Hill Moss...132
Figure 5.6 Mallachie Moss tephras.. 136
Figure 5.7 Craigmaud Moss tephras... 138
Figure 5.8 Ben Gorm Moss tephras.. 141
Figure 5.9 Ternary diagram showing the proportion of Fe, Ca and K oxides from
Ben Gorm Moss.. 143
Figure 5.10a Binary plot of FeO against CaO: BGMT-1 and BGMT-3............................. 144
Figure 5.10b Binary plot of FeO against MgO: BGMT-1 and BGMT-3.............................144
Figure 5.10c Binary plot of FeO against TiO_2: BGMT-1 and BGMT-3............................ 144
Figure 5.11 Bi-plot of silica:alkali ratios for isochrone BGMT-3 and Hekla-4 tephras....... 146
Figure 5.12 Bi-plot of silica:potassium ratios for Hekla-4 and Glen Garry tephras............ 148

Figure 5.13 Hekla-4: tephra loading and silicic content........ 149
Figure 5.14 Glen Garry: tephra loading and silicic content........ 150
Figure 5.15 Longbridge Moss age/depth model........ 160
Figure 5.16 Langlands Moss age/depth model........ 161
Figure 5.17 Temple Hill Moss age/depth model........ 162
Figure 5.18 Shirgarton Moss age/depth model........ 163
Figure 5.19 Mallachie Moss age/depth model........ 164
Figure 5.20 Craigmaud Moss age/depth model........ 165
Figure 5.21 Ben Gorm Moss age/depth model........ 166
Figure 5.22 Shirgarton Moss - SCP stratigraphy........ 167
Figure 6.1 The different *Sphagnum* species found on British ombrotrophic mires (after Daniels and Eddy, 1990)........ 180
Figure 6.2 Taxon-environment response model........ 199
Figure 6.3 Longbridge Moss: stratigraphy........ 215
Figure 6.4 Longbridge Moss - macrofossil diagram........ 216
Figure 6.5 Longbridge Moss - humification data........ 217
Figure 6.6 Longbridge Moss - detrended humification vs time........ 218
Figure 6.7 Longbridge Moss - testate amoebae........ 219
Figure 6.8 Longbridge Moss - plant macrofossils: detrending by segments........ 220
Figure 6.9 Longbridge Moss - plant macrofossils DCA: axis 1 scores vs time........ 221
Figure 6.10 Longbridge Moss - testate amoebae reconstructed water table depths +/- sample prediction errors vs time........ 222
Figure 6.11 Longbridge Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae plotted against time........ 223
Figure 6.12 Langlands Moss: stratigraphy........ 232
Figure 6.13 Langlands Moss - macrofossil diagram........ 233
Figure 6.14 Langlands Moss - humification data........ 234
Figure 6.15 Langlands Moss - detrended humification vs time........ 235
Figure 6.16 Langlands Moss - testate amoebae........ 236
Figure 6.17 Langlands Moss - plant macrofossils: detrending by segments........ 237
Figure 6.18 Langlands Moss - plant macrofossils DCA: axis 1 scores vs time........ 238

Figure 6.19 Langlands Moss - testate amoebae reconstructed water table depths +/- sample prediction errors vs time........ 239
Figure 6.20 Langlands Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae plotted against time........240
Figure 6.21 Temple Hill Moss: stratigraphy........248
Figure 6.22 Temple Hill Moss - macrofossil diagram........249
Figure 6.23 Temple Hill Moss - humification data........ 250
Figure 6.24 Temple Hill Moss - detrended humification vs time........251
Figure 6.25 Temple Hill Moss - testate amoebae........ 252
Figure 6.26 Temple Hill Moss - plant macrofossils: detrending by segments........ 253
Figure 6.27 Temple Hill Moss - plant macrofossils DCA: axis 1 scores vs time........254
Figure 6.28 Temple Hill Moss - testate amoebae reconstructed water table depths +/- sample prediction errors vs time........ 255
Figure 6.29 Temple Hill Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae plotted against time........256
Figure 6.30 Shirgarton Moss: stratigraphy........ 263
Figure 6.31 Shirgarton Moss - macrofossil diagram........ 264
Figure 6.32 Shirgarton Moss - humification data........265
Figure 6.33 Shirgarton Moss - detrended humification vs time........ 266
Figure 6.34 Shirgarton Moss - testate amoebae........267
Figure 6.35 Shirgarton Moss - plant macrofossils: detrending by segments........268
Figure 6.36 Shirgarton Moss - plant macrofossils DCA: axis 1 scores vs time........ 269
Figure 6.37 Shirgarton Moss - testate amoebae reconstructed water table depths +/- sample prediction errors vs time........ 270
Figure 6.38 Shirgarton Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae plotted against time........271
Figure 6.39 Mallachie Moss: stratigraphy........ 278
Figure 6.40 Mallachie Moss - macrofossil diagram........ 279
Figure 6.41 Mallachie Moss - humification data........280
Figure 6.42 Mallachie Moss - detrended humification vs time........ 281
Figure 6.43 Mallachie Moss - testate amoebae........282
Figure 6.44 Mallachie Moss - plant macrofossils: detrending by segments........283

Figure 6.45 Mallachie Moss - plant macrofossils DCA: axis 1 scores vs time........................ 284
Figure 6.46 Mallachie Moss - testate amoebae reconstructed water table depths +/- sample prediction errors vs time.. 285
Figure 6.47 Mallachie Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae plotted against time...286
Figure 6.48 Craigmaud Moss - macrofossil diagram..290
Figure 6.49 Craigmaud Moss - humification data.. 291
Figure 6.50 Craigmaud Moss - detrended humification vs time.. 292
Figure 6.51 Craigmaud Moss - plant macrofossils: detrending by segments........................ 293
Figure 6.52 Craigmaud Moss - plant macrofossils DCA: axis 1 scores vs time.................. 294
Figure 6.53 Craigmaud Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae plotted against time...295
Figure 6.54 Ben Gorm Moss - macrofossil diagram...298
Figure 6.55 Ben Gorm Moss - humification data... 299
Figure 6.56 Ben Gorm Moss - detrended humification vs time... 300
Figure 6.57 Periodogram showing plant macrofossil DCA values from Longbridge Moss over the last *ca.* 4700 years..306
Figure 6.58 Periodogram showing humification values from Longbridge Moss over the last *ca.* 4700 years..306
Figure 6.59 Periodogram showing plant macrofossil DCA values from Langlands Moss over the last *ca.* 4000 years..307
Figure 6.60 Periodogram showing humification values from Langlands Moss over the last *ca.* 4000 years..307
Figure 6.61 Periodogram showing plant macrofossil DCA values from Temple Hill Moss over the last *ca.* 7500 years..308
Figure 6.62 Periodogram showing humification values from Temple Hill Moss over the last *ca.* 7500 years..308
Figure 6.63 Periodogram showing plant macrofossil DCA values from Mallachie Moss over the last *ca.* 4900 years..309
Figure 6.64 Periodogram showing humification values from Mallachie Moss over the last *ca.* 4900 years..309

Figure 6.65 Periodogram showing plant macrofossil DCA values from Craigmaud Moss over the last *ca.* 2000 years......310

Figure 6.66 Periodogram showing humification values from Craigmaud Moss over the last *ca.* 2000 years......310

Figure 7.1 Evidence from plant macrofossil DCA axis 1 scores, humification, and reconstructed water table depths from testate amoebae at Temple Hill Moss from 1600 BC - 0 AD/BC......313

Figure 7.2 Climatic wet phases from all three proxy records at each site...... 318

Figure 7.3 Climatic dry phases from all three proxy records at each site......319

Figure7.4a One sample before the Glen Garry isochrone......321

Figure7.4b The Glen Garry isochrone...... 321

Figure7.4c One sample after the Glen Garry isochrone......321

Figure7.5a One sample before the Hekla-4 isochrone......324

Figure7.5b The Hekla-4 isochrone...... 324

Figure7.5c One sample after the Hekla-4 isochrone......324

Figure 7.6 Plant macrofossil DCA reconstructions for all sites......328

Figure 7.7 Humification plots for all sites...... 329

Figure 7.8 Reconstructed water table depths (mean values) from testate amoebae for all sites......330

Figure 7.9 *Sphagnum imbricatum* percentages against time for all sites where the species was identified...... 336

Figure 10a Longbridge Moss: The extinction of *Sphagnum imbricatum*...... 340

Figure 10b Langlands Moss: The extinction of *Sphagnum imbricatum*......340

Figure 10c Temple Hill Moss: The extinction of *Sphagnum imbricatum*...... 341

Figure 10d Shirgarton Moss: The extinction of *Sphagnum imbricatum*...... 341

List of Plates.

Plate 4.1 Aerial photograph of Longbridge Moss........86
Plate 4.2 Aerial photograph of Langlands Moss........89
Plate 4.3 Aerial photograph Temple Hill Moss........93
Plate 4.4 Aerial photograph Shirgarton Moss........96
Plate 4.5 Aerial photograph Mallachie Moss........100
Plate 4.6 Aerial photograph Craigmaud Moss........103
Plate 4.7 Aerial photograph Ben Gorm Moss........107
Plate 5.1 SEM showing a tephra shard (Hekla-4) from Mallachie Moss........115
Plate 5.2 SEM showing a close up of the vesicles on a tephra shard (Hekla-4) from Mallachie Moss........115
Plate 5.3 SEM showing a tephra shard (Hekla-4) from Mallachie Moss........116
Plate 5.4 SEM showing a highly vesicular tephra shard (Hekla-4) from Mallachie Moss...116
Plate 5.5 SEM showing curved platelets of bubble walls from a tephra shard (Glen Garry) from Ben Gorm Moss........117
Plate 5.6 SEM showing a highly weathered tephra shard (Glen Garry) from Ben Gorm Moss........117

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> “Nature has conducted complex
> but undesigned experiments for
> palaeoecologists to try to unravel.”
>
> H.J.B. Birks (1994)

Chapter 1. Introduction

1.1 Background

This thesis aims to reconstruct climatic change in Scotland from peat stratigraphy over the last *ca.* 5000-7500 years. Ombrotrophic (rain-fed) bogs have been shown to be sensitive to climatic changes as they rely mainly on precipitation for their moisture inputs. Past proxy effective precipitation (precipitation minus evapotranspiration) records from peat stratigraphy can be correlated with instrumental and documentary evidence, thus validating the peat climatic archive (Barber, 1981).

Palaeoclimatic research on peatlands has received a renewed interest over the last decade. Much work has been undertaken on mires in northern England, The Netherlands, and Scandinavia, which is documented in Chapter 2. However, very little peat stratigraphic work on climatic change has been undertaken in Scotland (Lowe, 1993), which is one reason why this region was chosen as a field area for this research. Scotland experiences a complex array of local and regional climates due to the variable topography (Harrison, 1997; Roy, 1997). Average January maximum temperatures can be of the order of 1^0C higher in the extreme southwest of Scotland than on the east coast, with similar differences for minimum temperatures (Meteorological Office, 1989). The geographical location of Scotland also makes it a sensitive site to climatic change, as it can be a 'hinge-line' for the movement of the polar front, which influences climate as weather patterns are trained along the southern flank of this boundary (Lowe and Walker, 1997). This suggests that Scotland could be the location in Britain where a climatic change would be registered first. A southerly shift of the polar front over Scotland has been suggested as being responsible for the Little Ice Age (LIA) in Scotland (Whittington, 1985), and may have been responsible for other climatic fluctuations throughout the Holocene.

Seven sites are selected for palaeoecological analysis in this project, covering a range of geographical and present day climatological regimes. Three main techniques are used in order to detect climatic changes: plant macrofossil analysis, colorimetric humification analysis, and testate amoebae analysis. These techniques are underpinned by a radiocarbon based chronology, supported by tephrochronology, and at specific sites analyses of spheroidal carbonaceous particles and pine pollen have been used to aid the dating of recent peats. The ability of tephra isochrones to precisely correlate climatic changes from multiproxy evidence at a suite of sites is an exciting and original development. Where tephras are present in the stratigraphy the potential synchroneity of climatic changes can be tested, something which cannot be achieved with radiocarbon dated profiles alone.

The techniques used for reconstructing climatic changes from peat stratigraphy rely on the changing surface moisture content of the bog. Temperature clearly has an important affect on these changes, either on evaporation or evapotranspiration. Isotopic analyses have attempted to reconstruct temperature fluctuations from the peat via deuterium ($^2H/^1H$) ratios (Dupont, 1986; van Geel and Middeldorp, 1988). Dupont (1986) considered the deuterium ratios to be influenced by temperature, but van Geel and Middeldorp (1988) found no correlation between the deuterium record from a bog in Ireland and climatic indices developed by Lamb (1977b) over the last 850 years, suggesting that differing plant taxa, specifically vascular plants and Sphagna, influence the enrichment of the isotopes within the peat. A new approach is currently being tested to examine the role of temperature in the peatland climatic record, which is discussed in section 2.4.3.

Many uncertainties still remain concerning the timing, magnitude and spatial scale of climatic changes, and a priority in palaeoclimatic work is to increase the amount and detail of proxy data for the Holocene (Maslin and Berger, 1997). This thesis aims to do this for Scotland, and to compare the results with identified climatic changes from northwest Europe. The synchroneity between climatic changes can be tested using tephra isochrones, and time series have been created in order to identify periodicities within the data. An appraisal of the development and biodiversity of the ombrotrophic bogs studied is also undertaken.

1.2 Thesis Structure

A detailed study of previous research on climatic changes reconstructed from peatlands is presented in Chapter 2. This section also details possible forcing mechanisms which produce Holocene climatic variability. The palaeoecological and chronological methods used are described in Chapter 3, as well as detailing the use of a tephrochronological approach to the research and the site selection strategy. This Chapter is written in reasonable detail, as it is important to understand the processes underpinning each technique. Chapter 4 provides the site descriptions, which includes an outline of the present day climatic conditions. Chapter 5 is dedicated to the chronology of each site. All the geochemical data for the tephra isochrones are presented, as well as inferences about the spatial dispersal of specific tephras. The radiocarbon dates are documented, and used in association with tephras, SCP and pine pollen profiles (where applicable) to create age/depth models for each site. Time series for further analyses were constructed from these models. The palaeoecological results are presented in Chapter 6, which includes a preliminary section on descriptions relating to moisture preferences of the various taxa identified in the thesis. This is vital for the interpretation of plant macrofossil and testate amoebae diagrams. The plant macrofossil data were subjected to detrended correspondence analysis, and a transfer function used on the testate amoebae data. The results are presented in site sections, and a synthesis of the three proxy reconstructions for an individual site is made at the end of each section. A final results section in Chapter 6 details the periodicities found at each site. A discussion of the main results is undertaken in Chapter 7, and the conclusions and suggestions for future research form Chapter 8.

Nomenclature follows Stace (1991) for vascular plants, Daniels and Eddy (1990) for Sphagna, Smith (1980) for non-*Sphagnum* bryophytes, and Charman *et al.* (in press) for testate amoebae.

Chapter 2. Previous research

2.1 Introduction

Ombrotrophic (rain-fed) peat bogs have been shown to contain archives of past environmental information, much of which can be specifically related to changes in climate (Godwin, 1981; Barber, 1981). Ombrotrophic bogs tend to be dominated by *Sphagnum*, which function as cation exchangers (Andrus, 1986), lowering the pH of the ecosystem which results in a species poor environment. The subfossil remains of these species are frequently well preserved, and can be used to infer former mire surface wetness given their sensitivity to specific moisture gradients.

This important terrestrial record has at times been undervalued as a proxy indicator of climatic change, primarily due to misconceptions about the formation of peat (Barber, 1994), which was once thought to grow in a cyclic regeneration pattern (Osvald, 1923). However, this theory was questioned when detailed stratigraphic work was undertaken on peat faces, notably by Walker and Walker (1961), and it was finally refuted by Barber (1978; 1981). The ideas and research with respect to reconstructing climatic changes from peat stratigraphy undertaken up to the 1970's are presented in the second part of this review. In the 1980's and 90's, as interest grew in Holocene research, coupled with the realisation that detailed climatic proxies could be extracted from peat stratigraphy, a rejuvenation of peat stratigraphic research occurred. This work is discussed in the third section.

At the start of the final decade of this millennium very little palaeoclimatic work had been attempted in Scotland using alternative lines of evidence from pollen-stratigraphic data (Lowe, 1993). As stratigraphic studies of plant macrofossils still remain to be better exploited (Huntley, 1996), this would suggest that there is a wealth of information on past climates still to be gathered from peat stratigraphy in Scotland. Whittington and Edwards (1997) endorsed these sentiments, although in the last few years an increasing amount of work has been accomplished in order to address these shortcomings, much of it concentrating on peat stratigraphy (e.g. Chambers *et al.*, 1997a; Anderson, 1998; Anderson *et al.*, 1998; Baker *et al.*, 1999). A detailed study of palaeoclimatic inferences in Scotland from peat stratigraphy is discussed in the fourth part of this review.

Before palaeoclimatic reconstructions from peat stratigraphy can be made, it is necessary to contemplate the factors relating to the origins and development of ombrotrophic bogs which may adversely affect the detection of a climatic signal. Detailed reviews of these factors have been undertaken in the theses of Stoneman (1993) and Mauquoy (1997), so a brief synopsis will be given here (section 2.2). The final section of this review will document the main forcing factors thought to affect Holocene climatic change, followed by a synthesis of northwest European climatic changes over the mid to late-Holocene.

This thesis will use calibrated AD/BC radiocarbon dates wherever possible, but where dates have been quoted in papers in years BP (calibrated or uncalibrated), they will be quoted here as such due to uncertainties with calibration techniques or errors on the original dates. The exception for this rule is in Table 2.1 (in section 2.6), where all the dates have been converted to AD/BC (the above *caveats* notwithstanding) in order to make comparisions between data easier.

2.2 Factors influencing the climatic signal on ombrotrophic peatlands

None of the factors discussed in this section have been considered as serious problems in climatic reconstructions from peat stratigraphy in Britain (Stoneman, 1993; Barber *et al.*, 1994a,b,c; Mauquoy, 1997; Mauquoy and Barber, 1999a,b; Hughes *et al.* in press), with the possible exception of an internal hydrological regime distorting the climatic signal at Ellergower Moss (Stoneman, 1993), which is discussed below. However, it is important to be aware of the possible problems with climatic reconstructions from peat stratigraphy, especially as mires throughout the world may respond differently to climatic changes, and have dissimilar autogenic mechanisms (e.g. Haslam, 1987; Glaser *et al.*, 1997).

2.2.1 Internal hydrological controls

Mire hydrology is influenced by the relative biophysical properties of the organic accumulations of peat massifs, which are commonly divided into two horizons. Ivanov (1981) defines them as follows: the upper horizon, or *acrotelm*, has a variable water content, high hydraulic conductivity, with periodic aeration and relatively intense biological activity; the lower layer, or *catotelm*, is constantly saturated, with low hydraulic conductivity, and an absence of aeration with only slight biological activity. The catotelm of an intact mire is overlain by the acrotelm. Autogenic changes in the internal hydrology of ombrotrophic bog development could possibly mask climatic changes, as mires have only a limited ability to store excess water in their aerated acrotelm, although this is only thought to affect mires once the equilibrium between bog hydrology and prevailing climate has been altered (Stoneman, 1993).

The water table of a raised bog is domed, and maintained in dynamic equilibrium by impeded drainage (Ivanov, 1981; Ingram, 1982), and although after rain the water would sink quickly through the acrotelm, it would sink very much more slowly through the catotelm. The removal of excess water from the mire surface may occur through subsurface pipes. Although not well understood, Foster and Fritz (1987) suggested that these pipes may have developed due to hydrostatic pressure on the peat floor which followed lines of weakness in the subsurface peat. Subsurface pipes may be responsible for draining the surface of Ellergower Moss (located in the Silver Flowe, Scotland), as the mire is situated in an area of comparatively high precipitation but exhibits a relatively dry sequence (Stoneman, 1993). Rapid surface runoff can also occur, which may initially store the excess water by ponding. As this storage is exceeded, the movement rapidly occurs from one microtopographical low to another, and is eventually discharged into the lagg stream (Boatman, 1983).

When catastrophic flooding of mires occurs, the peat stratigraphy is often able to reflect such events. At Bourtangersveen in the Netherlands, a bog burst is thought to have occurred at *ca.* 2500 BP (Casparie, 1972). Analysis of the stratigraphy revealed evidence of many erosion channels and the sudden replacement of a *Sphagnum* section *Cuspidata* layer by *Sphagnum imbricatum* and *Sphagnum papillosum*. It was thought that the sudden drainage of an unstable mire lake (which was formed due to the climatic deterioration at this time) led to the erosion channels within the mire, which subsequently allowed the colonisation of other Sphagna. The event was found to be relatively short lived compared with the whole peat record.

2.2.2 Artificial raising of the water table

Ombrotrophic bogs are considered to have a raised water table which operates in a detached manner from the groundwater table. However, it has recently been suggested that ombrotrophic bogs can be recharged by an artesian upwelling of groundwater during dry periods, resulting in internal water tracks (Glaser *et al.*, 1997). This model has been developed from bogs which originate on sand and gravel substrates (within the glacial lake Agassiz region of Minnesota), and it is thought that this process should not affect basin mires with a clay substrate, as is common in Britain. However, this process of upwelling groundwater has been further developed by Lamers *et al.* (1999), who suggested that a calcium carbonate rich groundwater influence on raised water tables can react with acids in the upper peat layers forming CO_2. This mechanism promotes higher levels of decomposition to take place, producing relatively high levels of methane and CO_2. The CO_2 diffuses from the peat, stimulating photosynthesis and growth in aquatic and terrestrial Sphagna, and part of the methane oxidises to CO_2. It may therefore be possible that not all the nutrients utilised by ombrotrophic mire flora are received from the atmosphere. However, an increase in CO_2 concentrations has been shown to have little or no effect on the growth of *Sphagnum fuscum*, as it was found that low water levels are more of a limiting factor (Jauhiainen *et al.*, 1997).

Methane is formed in peat as the cellulose of dead plants, which is usually degraded aerobically to CO_2 and water, becomes anaerobically decayed by a consortium of bacteria (Brown and Overend, 1993). Considerable amounts of methane are suggested to be trapped in ombrotrophic bogs, and it is thought that these bubbles (*ca.* 70μm in diameter - Brown, 1997) may reduce the hydraulic conductivity in the lower layers of the peat by blocking the pore spaces, preventing fluid movement, and elevating the water table (Brown *et al.*, 1989). The impeded drainage suggested by Ingram (1982) could therefore reflect trapped gaseous methane in the catotelm (Brown, 1997).

At present there is not yet enough data to suggest whether the amount of methane present in a bog affects the ambient mire water levels. Brown (1997) suggested that more information on the microbial metabolism of peat bogs is necessary before the water movement in mires can be realistically modelled. However, recent research has indicated that the major pathway for methane fluxes to the atmosphere is through *Eriophorum* *vaginatum* tussocks, which exhibit significantly higher fluxes than *Sphagnum* (Frenzel and Rudolph, 1998). This is due to very low rates of methane oxidation associated with *Eriophorum*, and it is suggested that methane bubbles may become trapped in the network of *Eriophorum* roots within a tussock. Subsequently, methane diffuses into the roots and is eventually transported to the atmosphere. Frenzel and Rudolph (1998) noted that the *Eriophorum* roots and rhizomes cover a wide area, often extending below *Sphagnum* lawns, and may therefore drain the methane even from these areas. The problem of methane artificially raising the water table and damping any climatic signal from peat stratigraphy might thus be not as important as was initially thought.

2.2.3 Anthropogenic disturbances

Humans have not only damaged and destroyed large areas of mires, but have also played an important role in the development of the British peatland landscape (Charman, 1997a). The best example is probably the initiation of blanket peats which are frequently associated with human activity from the Mesolithic onwards (Moore, 1993). However, a large amount of damage to mires has occurred over the last few hundred years. Smout (1997) reviews 'Bogs and People since 1600', and comments from accounts written between 1630 and 1730, on the reputation of bogs as a useful resource being particularly high in Scotland. Most of these uses were for fuel, but reclamation was not unusual - an account from 1724 describes parts of Flanders Moss being converted into arable land.

Burning and grazing is likely to have an effect upon bog flora (Chambers, 1997). Burning is considered to be the more serious of these influences, and is discussed in detail in Chapter 6 (section 6.1.25) with respect to macroscopic charcoal remains recovered from peat stratigraphy. Welch (1997) considered the loss of heather and prevention of tree regeneration or colonisation to be the main impacts of grazing by large herbivores on British peatlands, implying that Ericaceae may at times be under-represented in the palaeoecological record due to non-climatic factors.

The cutting of drainage ditches on peatlands can influence the hydrology and affect the water table levels. Van der Molen *et al.* (1992) studied the recent changes in hummock-hollow complexes from Clara Bog, Co. Offaly. They found that drier conditions prevailed between *ca.* AD 1600 - 1880 (which is unusual as many sites register an increase in moisture or cooler conditions at this time often referred to as the Little Ice Age (LIA), *cf.* Barber *et al.*, 1999; Mauquoy and Barber, 1999a; Barber *et al.*, in press), whereas after 1880 the water table rose significantly and increased wetness continued until at least 1927. However, although they considered that human influences were unlikely, surface drains had been cut from AD 1700 - 1750 onwards to drain the peatlands for hunting purposes. After the Great Famine (1845 -1851) the drains became overgrown and filled in, corresponding with an inferred increase in surface wetness conditions soon after. The former suggests that the influence of the drains can be clearly seen in the peat stratigraphy, even though the authors suggested climate as the main mechanism for changes in surface wetness.

An increase in atmospheric deposition of nutrients from anthropogenic activities has also been considered to influence peat growth. Van Geel and Middeldorp (1988) considered the extinction of *Sphagnum imbricatum* at Carbury Bog (Co. Kildare, Ireland) to be related to an increase in dust flux as a consequence of human activity around the bog. They argued that the subsequent abundance of *Sphagnum magellanicum* was possibly due to increased atmospheric deposition of nitrate. There is some evidence to support this theory, as the effects of variable concentrations of nitrate and ammonium on growth of *Sphagnum magellanicum* have been studied by Rudolph and Voigt (1986). They discovered that although for several *Sphagnum* species an optimum nitrate concentration was of 100 μM, concentrations of up to 322 μM were found to be favourable for *Sphagnum magellanicum.* The growth of *Sphagnum magellanicum* was, however, discovered to be rather sensitive to ammonium ions, which has also been found to affect *Sphagnum cuspidatum* (Press and Lee, 1982; Press *et al.*, 1986). Li and Vitt (1994) have also suggested that nitrate loading may significantly affect the establishment of certain moss species, while not affecting others, a feature which has also been noted in *Sphagnum* species by Kooijman and Bakker (1995). More recently it has been recognised that in highly polluted areas, which have high nitrate levels, a further increase in nitrogen deposition resulted in no increase in *Sphagnum* productivity (Aerts *et al.*, 1992). It was suggested that this limitation to plant growth is due to a deficiency in phosphorus, indicating that the N:P ratio for a mire may be crucial to productivity (Aerts *et al.*, 1992; Willams and Silcock, 1997). Hence, an increase in pollution may have promoted the abundance of certain Sphagna, and as *Sphagnum magellanicum* grows faster than *Sphagnum imbricatum* this may have given it a competitive advantage.

An anthropogenically derived signal may therefore be apparent within the last few hundred years of peat growth, although peat based proxy climate records have been positively correlated with instrumental and documentary records (e.g. Barber, 1981; Blackford and Chambers, 1991; Charman *et al.*, 1999). Long records of climatic change can also be ascertained, even from bogs which have been recently cut and drained, e.g. Bolton Fell Moss (Barber *et al.*, 1994b; Hughes *et al.*, submitted).

2.3 Climatic reconstructions from peat stratigraphy

2.3.1 Introduction

Ombrotrophic bogs accumulate peat and maintain their water tables above the local groundwater table, and are thus particularly sensitive to changes in climate as they rely principally on precipitation inputs to maintain and sustain them. Analysis of plant macrofossils within the peat stratigraphy, as well as specific microfossils and changes in humification, enables semi-quantitative measurements of changes in past water table depths to be inferred (e.g. Charman *et al.*, 1999). Hence proxy records of changes in effective precipitation can be derived, which reflects the balance of precipitation remaining after evaporation.

Reconstructing past mire water table depths enables the identification of wet and dry phases within the peat sequences. This allows detailed climatic histories to be reconstructed, which can be correlated through a radiocarbon based chronology to historic-documentary and instrumental climatic changes (Barber, 1981). In undamaged bogs the climatic archive can extend back to *ca.* 8000 years BP (Hughes *et al.*, in press), and time series analysis of the climatic fluctuations can reveal dominant periodicities within the records (Barber *et al.*, 1994b).

Peat bogs are relatively easy to date using radiocarbon techniques, although potential problems do exist (see Kilian *et al.*, 1995). However, there is now an increasing interest in determining rates of change within palaeoclimatic records, which limits the use of bulk radiocarbon dating (Pilcher, 1991). Approaches such as 'wiggle-matching' are therefore necessary (van Geel and Mook, 1989), although the use of this technique can be a costly exercise. Most of the past work on peat bogs has utilised bulk radiocarbon dates, which allows questions related to the general timing of climatic changes to be addressed, but are not precise enough to determine if these changes are synchronous between sites. Techniques such as tephrochronology and dendrochronology are now being used in order to answer such questions.

2.3.2 Peatland pioneers and the establishment of a climatic proxy

Studying changes in peat humification as a proxy for past climatic change has been undertaken since before the start of this century, notably by Blytt (1876) and Sernander (1908). However, misconceptions by Von Post and Sernander (1910) and later by Osvald (1923) led to the development of the theory of cyclic regeneration of peat growth. The fundamental problem with this theory was that at no time had peat stratigraphy been properly studied, as most observations had been concerned with surface features (Backeus, 1990). Conway (1948) suggested that if the climate was found not to be stable but always in some phase of a larger and smaller oscillation then the cyclic regeneration theory could be considered flawed, although it was only when detailed investigations into peat stratigraphy were undertaken that progress was made. It was realised that the positions of hummocks and hollows remained constant over time, thus not supporting the cyclic regeneration theory (Walker and Walker, 1961; Casparie, 1972), and the theory was finally refuted by Barber (1978; 1981) who produced evidence for a replicatable climatic response from 21 profiles from Bolton Fell Moss, Cumbria. For a full review of the falsification of the cyclic regeneration theory see Barber (1981).

One of the first stratigraphical features of mires to be attributed to climate change were recurrence surfaces, the original of which was termed the *Grenzhorizont* - a sharp division between the lower highly humified *Sphagnum-Calluna-Eriophorum* peat and the upper pale fresh *Sphagnum* peat (after Weber, 1900). The *Grenzhorizont* documented by Weber was estimated to be dated between 800-500 BC, and was initially thought to contain a hiatus of up to 1000 years between the horizons, due to a cessation in peat growth initiated by the late Sub-Boreal warmth or dryness as postulated by Blytt and Sernander. The subsequent regrowth of peat was attributed to a sudden deterioration to cool and wet summers (in Barber, 1981).

Granlund (1932) found that recurrence horizons occurred consistently amongst raised bogs in Sweden, probably the most famous of which is the boundary between the 'old' Sub-Boreal and Sub-Atlantic at around 600 BC (Weber's *Grenzhorizont*), which has been found over large parts of northwest Europe (reviewed by van Geel *et al.*, 1996). Granlund (1932) documented other recurrence surfaces which he dated to 2300 BC, 1200 BC, AD 400, and AD 1200. This approach made a considerable impression on Godwin (1946, p.6), who wrote:

> "This simple and attractive picture of a bog-horizon due to climate change and available over big distances as a correlation level has suffered a good deal of recent scrutiny and modification, but there is no reason to doubt its essential validity."

An address by Godwin to the Royal Meteorological Society in 1966 on climate change between 8000 to 0 BC highlighted the main themes being explored at the time. He stressed the inherent difficulties when trying to correlate recurrence surfaces found at various sites in northwest Europe using pollen analyses, and suggested that radiocarbon dating was the next major step in an attempt to determine synchronous climatic changes. However, once radiocarbon dating became available to palaeoecologists it became obvious that recurrence surfaces could not always be correlated synchronously between sites, although the metachroneity of the recurrence surfaces were apparent (Frenzel, 1966). Overbeck *et al.* (1957) obtained radiocarbon dates from recurrence surfaces in German bogs which were dissimilar to Granlund's results, including a date of 100 BC for the main recurrence surface which Granlund had dated at 600 BC. Lundquist (1962) published radiocarbon dates for recurrence surfaces in two Swedish bogs which appeared to differ in age and produce evidence of a time-gap between the old *Sphagnum* and younger *Sphagnum* regrowth (see comments in Barber, 1982). Godwin (1966), however, suggested that the metachroneity of the recurrence surfaces agreed with British findings at Tregaron bog in Wales and Chat Moss, Lancashire. Nilsson (1964a) produced a suite of radiocarbon dates which enabled a chronology to be established throughout the Holocene, documenting the changing growth rates of the peat, and relating them to changing wetness of the mire surface (in Godwin, 1966). The publication of a second paper (Nilsson, 1964b) used pollen diagrams which were used to correlate recurrence surfaces from a number of sites at Ageröds Moss (southern Sweden), and indicated that the humification zones were synchronous (in Dickinson, 1975). It was therefore deduced that periods of relatively moist/cold climates, as well as phases of relatively dry/warm climates could be distinguished from peat stratigraphy.

Much of the early palaeoecological work on peatlands in Ireland was undertaken by Jessen (1949), whose work was later followed up by Mitchell (1956) who published pollen diagrams for a number of Irish raised bogs, and found many recurrence surfaces (between one and three per bog). Mitchell suggested that climatic change may not be responsible for the development of each surface, and he inferred that autogenic factors may be involved. However, as an oceanic climate tends to dominate in western Ireland, many of the bogs exhibit a 'wet' stratigraphy, so that only dry peaks can be detected in the stratigraphy (Barber, 1978). Mitchell did produce evidence for concentrations of wet shifts around 3450 BP, 2750 BP, and 1450 BP, which appear to agree with other palaeoecological records in northwest Europe (see Table 2.1).

As radiocarbon dating of peat deposits became more routine, so the amount of research into peatlands, and specifically climate change grew. Detailed analyses of peat stratigraphy (e.g. Walker and Walker, 1961) gave initial clues that there might be a more refined climatic story in peatlands, specifically in the subtle variations rather than just recurrence surfaces. More detailed analyses of plant macrofossils and microfossils followed (e.g. Casparie, 1972; van Geel 1978), and the correlation of a proxy climatic record from peat stratigraphy with independently derived climatic indices (Lamb, 1965a; 1965b; 1966; 1977a; 1977b) by Barber (1981) rejuvenated interest in the peat archives, and opened the door to the reconstruction of detailed peat stratigraphic proxy climatic records.

2.3.3 Rejuvenated interest in peat stratigraphy

A great amount of research into peatlands, and effort in the generation of palaeoecological data, took place in the Netherlands in the 1970's and 1980's. The inferences generated from such analyses also became more complex. Wijmstra *et al.* (1971) discovered a short-term cycle of 80 years in the *Alnus* curve from Wietmarschen Moor, eastern Netherlands, with the cycle occurring every 4cm throughout the profile (1cm contiguous samples were taken). The 80 year cycle (possibly with a solar influence as it corresponds with the well known 80-100 years Gleissberg cycle – see Frick *et al.*, 1997) was suggested to be responsible for changes in the local water table, and hence *Alnus* trees, by either varying pollen production or by actually altering the composition of the forest. However, assumptions regarding the validity of the cycle must be questioned due to the poor chronological control, which was based on average rates of sedimentation from a nearby core, and subtle variations in peat growth would alter such a short cycle.

With the help of radiocarbon dating, another profile from Weitmarschen Moor - III, van Geel (1972) described *Corylus* pollen cycles with a 32 year frequency. Van Geel (1978) suggested that *Corylus* pollen changes are a good indicator of climatic change as they appear high when hummock forming vegetation is dominant and comparatively low when wet *Scheuchzeria* peat dominated the stratigraphy. The mechanism by which *Corylus* pollen declines in wetter conditions relates to the *Corylus* catkins, which can often degenerate without having produced any pollen in years with very rainy late winters. Macrofossil evidence supported this idea, as hygrophilous *Sphagnum cuspidatum* was found at the same stratigraphic levels as the evidence for *Corylus* decline. However, the *Corylus* pollen curve from a nearby profile at Engbertsdijksveen showed cyclic fluctuations in the order of 150-200 years (van Geel, 1978). The two sites are throught to have reacted synchronously to climatic changes as *Sphagnum imbricatum* growth was initiated on both sites at the same time and the main *Corylus* pollen curve changes are similar between sites (van Geel, 1978). A marked decline in *Corylus* pollen occurred between 1400-600 BC at both sites, and although this has been interpreted as a synchronous decline, the large range of dates which fall within this phase do not necessarily suggest synchroneity. It is possible that this is the reason why different cycles have been identified from the two data sets.

Aaby (1976) was able to secure a good chronology in order to investigate humification changes from peat stratigraphy. He utilised 70 radiocarbon dates of which 55 were from a single 2.5 metre peat profile from Draved Moss, Denmark. The peat record extended for 5500 years, within which Aaby claims to have detected a periodicity of 260 years, with some double periods of 520 years.

Other research around this time in the Netherlands tended to concentrate on macrofossils, as well as microfossils. Casparie (1972) undertook research from Bourtanger Moor, near the Dutch/German border. He suggested that the synchronous invasion of *Sphagnum imbricatum* and *Sphagnum papillosum* on a highly humified peat surface could have been due to an increase in precipitation at *ca.* 2000 BC, indicating a change to more oceanic conditions. Van Geel (1978) was also able to infer various shifts in climate since the Atlantic period from an examination of the macrofossil and microfossil stratigraphy of the Engbertsdijksveen (ENG I), and he also pioneered research into fossil fungal spores and palynomorphs, which may have a major influence in palaeoecology in the future.

Another profile from the Engbertsdijksveen (ENG VII), taken over 1km away from the core analysed by van Geel (1978), has been investigated by Dupont and Brenninkmeijer (1984). A decline of *Corylus* pollen occurred at 820 BC, and was inferred to be in good agreement with ENG I. The suggestion of a climatic deterioration at this time was also indicated by isotopic records - the $^{2}H/^{1}H$ and $^{18}O/^{16}O$ ratios of cellulose. Comparisons between ENG I and ENG VII also revealed simultaneous expansions of *Sphagnum* section *Cuspidata* and *Calluna.* However, Dupont and Brenninkmeijer (1984) suggested that ENG VII was wetter than ENG I at 800 BC, and that changes in *Sphagnum* assemblages could not always be easily correlated between profiles, suggesting local ecological variability.

Palaeoclimatic inferences from peat stratigraphy in Britain were still ongoing in the 1970's, although much of the research on mires concentrated on detailed structural and hydrological features of bog surfaces (e.g. Boatman and Tomlinson, 1973; 1977). Dickinson (1975) analysed a number of cores from Rusland Moss, Cumbria, in order to test whether the recurrence surfaces identified in the sequences were synchronous.

Dickinson noted that subtle changes in humification could be detected within the peat sequences (sampled with a Russian corer), although the analyses lacked detailed macrofossil identification, and there was little interpretation in terms of climatic changes. However, Dickinson did propose that the recurrence surfaces identified *were* at synchronous levels throughout the bog, even though some of them displayed more gradual changes in humification. The main changes in humification were dated to 740 BC, AD 400, and AD 1145, which compare well with other reconstructions (Table 2.1).

The main wet phases identified from 21 profiles by Barber (1981) at Bolton Fell Moss, Cumbria, occurred between AD 900-1100, AD 1320-1485, and AD 1745-1800. Dry phases were also suggested, although they were harder to identify with any certainty, as the onset of a dry phase could result in a cessation in peat growth or peat erosion. Having stated this, the main dry periods recorded at Bolton Fell Moss were between AD 90-600, the AD 800's, the AD 1200's, and the early AD 1500's. Further work carried out on macrofossil data from Bolton Fell Moss by Barber *et al.* (1994b) suggested that fluctuations in moisture operated over a 800 year periodicity, with an ocean-driven cycle being the possible driving mechanism.

Stoneman (1993) also worked on Bolton Fell Moss, and produced evidence for climatic deteriorations between AD 930-1120 (similar to Barber, 1981), and around 840 BC and 530 BC, either of which could be similar to the event which occurred across NW Europe around 2650 BP (van Geel *et al.*, 1996). Research on Bolton Fell Moss has been undertaken by Mauquoy (1997), who found a periodicity from macrofossil changes of 260-280 years, which is directly comparable with the cyclicity found by Aaby (1976) of 260 years. Recent work at the site (Hughes *et al.*, submitted) compared the 1994 core (BFMJ - Barber *et al.*, 1994b) with another core taken from within 100 meters (BFML). Several wetter phases are apparent in both profiles, the most prominent of which commenced at *ca.* cal. 4700-4500 BP. This event was clearly one of the most significant events of the mid-late Holocene as it has also been reported in fluctuating lake levels (Digerfeldt, 1988; Yu and Harrison, 1995) and from other mire data (Korhola, 1995; Anderson, 1998). Other climatic deteriorations have been identified by Hughes *et al.* (submitted) around *ca.* cal. 3000-2800 BP, 1900 BP, and *ca.* cal. 1000 BP. Futhermore, detailed climatic reconstructions for the last 1500 years have been undertaken on Bolton Fell Moss using testate amoebae analysis (Woodland *et al.*, 1998).

Dupont (1986) constructed temperature and rainfall records for the Holocene using palaeoecological and isotope analysis from the Meerstalblok, part of the former extensive Bourtangerveen in the Netherlands. A semi-quantitative curve of moisture conditions was calculated from the palaeoecological data which employs simple weighted averaging with respect to the moisture tolerance of the species. An element of subjectivity is involved with this method, as some taxa have bimodal moisture distributions (specifically *Sphagnum imbricatum* and *Sphagnum* section *Acutifolia*), although the method does give a good first approximation for climatic changes. The analyses identified a number of climatic phases, several of which corresponded to phases observed in the records from Engbertsdijksveen (*cf.* van Geel, 1978; Dupont and Brenninkmeijer, 1984). Dupont found a trend towards drier vegetation between 4500-4000 BP, becoming wetter between 3700 and 3500 BP. A good correlation is also apparent between both sites which suggest a wet phase between 3500-2500 BP. Dupont also found a periodicity of 206 years in the temperature data, and tentatively suggested that the cycle could be linked to solar variability and climate, but stated the need for more data before drawing firmer conclusions.

Work undertaken at Store Mosse in southern Sweden was able to show a change in raised bog stratigraphy which appeared indicative of increasing humidity (Svensson, 1988a). This has been dated to *ca.* 2400 BP and is comparable with Granlund's (1932) recurrence surface III and is indicative of the general climatic deterioration found over NW Europe around this time (van Geel *et al.* 1996). Another recurrence surface was indicated by a '*Magellanicum* bog stage' which replaced the '*Fuscum-Rubellum* bog stage' and was radiocarbon dated to *ca.* 1200 BP. Svensson (1988b) suggested that the climate may have become more oceanic during the development of the mire, as the peat forming communities showed an upward trend comparable with the present day east-west gradient in bog communities.

Changes in humification from ombrotrophic peatlands have also yielded valuable proxy climatic data. Nilssen and Vorren (1991) sampled 14 ombrotrophic mires in northern and central Norway and subjected them to colorimetric humification analysis. Their data appears to correlate well with the results from Aaby's 1976 work, and significant phases of lower humification, indicating wetter climates were identified between 3850-3750 BP and 2950-2750 BP. Recent research concentrating on colorimetric changes in humification in Britain has also been undertaken, notably by Blackford (1990), Blackford and Chambers (1991; 1995), and Chambers *et al.* (1997a), and in conjunction with other proxy indicators by Chiverrell and Atherden (1999), Mauquoy and Barber (1999a), and Baker *et al.* (1999), which are discussed later.

Blackford and Chambers (1991) provided evidence for a Dark Age (1400 BP) climatic deterioration from five sites in a transect from western Ireland to North Yorkshire, which can also be seen in many other palaeoecological reconstructions (see Table 2.1), and is within the same timescale as the AD 536 volcanic event (Baillie, 1994 and section 2.5.2). Blackford and Chambers (1995) compared two radiocarbon dated blanket peat humification records from near Letterfrack, Co. Galway, western Ireland over the last 1000 years. The broad similarities between the datasets suggested a climatic origin for the humification changes. Wetter (or cooler) periods were recorded between *ca.* AD 1410-1540 and AD 1660-1720. Drier (or warmer) periods were registered between *ca.* AD 1230-1380, AD 1560-1650 and AD 1730-1780. The results from Blackford and Chambers (1995) show similarities with historical records and other proxy records for the Medieval Warm Period (MWP) and Little Ice Age (LIA). Blackford and Chambers proposed that the palaeoclimatic records from Letterfrack could be correlated with sunspot activity, notably the Maunder minimum (AD 1650-1715) and Spörer minimum (AD 1420-1530), and suggested that the peat record was in general agreement with solar activity between AD 1650 and 1900. However, the earlier part of the record provided no clear correlation with the Wolf minimum (AD 1280-1340).

Many palaeoecological investigations have been undertaken on the raised bog complexes in southern Finland, but due to the flat topography most of the mires have developed through paludification. Studies have therefore been more concerned with analysing the development of peat growth before trying to decipher the climatic signals that may be in the peat stratigraphy (see Korhola, 1992; 1994 and Mäkilä, 1997). As paludification of the mires occurred, vertical growth would have slowed, suggesting that peat growth, both vertical and lateral, appeared not to have proceeded at a uniform rate, making climatic inferences difficult (Korhola, 1992). However, a more 'peaceful period' of mire growth (mainly paludification) occurred between 6000-4000 BP at many mires in southern Finland, and this has been interpreted as a dry climatic phase (Korhola, 1992; 1994). A more humid climate has been suggested for the formation of slightly humified peat between cal. 3300-2700 BP and during the last 1300 years (Mäkilä, 1997).

A regional approach has been adopted by Korhola (1995), in order to identify phases of climatic change from peat stratigraphy in southern Finland. He studied 71 radiocarbon dated basal peats, known to have originated by paludification, from 23 different sites in southern Finland. A period of more intense peat formation between *ca.* cal. 4300-3000 BP was identified, and as this period appeared strikingly similar to lake level changes in southern Sweden, an increase in humidity at this time was suggested. A period of increased dryness has been identified earlier in the Holocene, starting at about cal. 7300 BP. This period corresponded with a less active phase of peatland initiation, and has also been documented in decreasing lake levels (Digerfeldt, 1988). This dry phase ended at about *ca.* cal. 5600-4400 BP in southern Sweden (Digerfeldt, 1988; Harrison *et al.*, 1993), which coincided with the intensive peat formation which began at cal. 4300 BP. Allogenic forcing (climatic change) has therefore been suggested as a possible mechanism (Korhola, 1995).

A recent study by Almquist-Jacobson and Foster (1995) of raised bog development in central Sweden found that fen development leading to an accumulation of *Sphagnum* peat occurred between 5000-4000 BP, during relatively dry phases. This dry phase fen-bog transition was also found in British mires by Hughes (1997), although the event generally occurred earlier in the Holocene in Britain. This highlights the differences between the climatic phases suggested by the Finnish mires and British mires, and it is possible that the mires in Finland are not so climatically sensitive.

The recent peatland palaeoclimatic work in Britain appears to have been mainly undertaken in northern England, and more recently in Scotland (documented in section 2.4). Complementary to the work carried out on Cumbrian raised bogs, extensive research into palaeoclimates from the blanket peats in the Pennines has been undertaken by Tallis (1994; 1995; 1997). Plant macrofossil analysis at Alport Moor (Tallis, 1994) found relatively high levels of *Sphagnum* leaves in just over half the samples recorded. The majority were *Sphagnum* section *Acutifolia*, although *Sphagnum* section *Cuspidata* were also recorded, signifying wet phases, and *Sphagnum imbricatum* and *Sphagnum magellanicum* were found associated with active hummock building. Tallis found that changes in *Sphagnum* richness, when radiocarbon dated, compared well with the climatic phases of Lamb (1977b). However, care must be taken when interpreting an increase in *Sphagnum* cover with respect to climate change. Other factors influence vegetational changes in blanket mire systems, notably the topography (and its relative retention of impacting water), pollution (Ferguson and Lee, 1978; 1980) and wildfires, possibly resulting in permanent loss of *Sphagnum* cover (Tallis, 1987; 1994). Studies on the nearby Featherbed Moss were used to interpret the development of the erosional system on Alport Moor. Tallis (1994) interpreted a climatic phase shift from dry to wet conditions at 1400 BP, which has also been recorded in the same region by Blackford and Chambers (1991) and Mauquoy (1997). There also appeared to be strong evidence for a much drier climate between AD 1150 and 1300. Before this period there seems to have been dry bog vegetation at *ca.* 2800 BP, which developed into a wet surface by *ca.* cal. 200 BC, providing similar dates to the climatic deterioration around 2650 BP (van Geel *et al.*,1996).

At other blanket peats in the Pennines, Tallis (1995; 1997) has used the remains of *Racomitrium lanuginosum* macrofossils (Holme Moss and Over Wood Moss), and *Empetrum nigrum* pollen records to infer climatic signals. *Racomitrium lanuginosum* displays oceanic characteristics (Lindsay *et al.*, 1988); however, competition from other mire flora, mainly Sphagna, tends to restrict its distribution to areas with a lower water table. Therefore, its presence indicates high humidity (wet climate) with a dry bog surface, potentially indicating in the palaeoclimatic record a change from relatively dry conditions to a wetter climate. The *Racomitrium* levels in the Pennines are synchronous with a rise in *Plantago* pollen, which has been dated by reference to historical dates (Tallis and Switsur, 1973) and by correlation with radiocarbon dates from nearby profiles at Alport Moor and Featherbed Moss (Tallis and Livett, 1994). Two *Racomitrium* zones occurred at these sites, the first appeared to be between *ca.* AD 1250 - 1450. This would have been associated with the documented shift to wetter climates after *ca.* AD 1300 and appeared to tie in with the recognised early Medieval Warm Period of AD 1150 - 1300 (Lamb, 1985). The upper zone has been dated to *ca.* AD 1700 - 1750, implying dry climatic conditions at around AD 1700, just before a phase shift to a wetter climate after this date.

Records of *Empetrum nigrum* pollen have been assembled for 18 blanket bogs in the southern Pennines (Tallis, 1997). *Empetrum nigrum* prefers to live in areas with reduced water table levels, and as the pollen is dispersed only over short distances, its occurrence in abundance in the pollen record can be viewed as a proxy for climate change. Surface quadrat sampling confirmed that high values of *Empetrum* pollen were indicative of purely local presence and most likely to be from hummock forming microtypes. The main episodes of high *Empetrum* pollen occurred prior to *ca.* 860 BC and between *ca.* AD 1100-1250. The first date indicated a warm period before the climatic deterioration seen over NW Europe at 2650 BP, and the second corroborates the *Racomitrium lanuginosum* record of Tallis (1995) as an indicator of the Early Medieval Warm Period.

Much of the dating for Tallis' work relies on relatively few radiocarbon dates, and correlations with pollen curves relating to historical dates, resulting in a certain amount of subjectivity. The fact that the dates identifed by Tallis correlate with independently derived climatic phases may be quite correct and emphasise the strength of the palaeoclimatic record, but it may also be a case of 'suck in and smear' (Baillie, 1991), signifying the need for a better chronology before validating the proxy climatic records.

Recent work on reconstructing climatic changes from peat stratigraphy in Cumbria, by Mauquoy and Barber (1999a), presented a replicated 3000 year proxy climate record from Coom Rigg Moss and Felecia Moss. The study recorded ten periods of effective precipitation through detailed plant macrofossil and colorimetric humification analyses.

The main wet phases were dated using a chronology of radiocarbon dates which was supported by pine pollen correlations, and the subsequent age/depth model was able to interpolate significant dates. Both proxy records showed periods of increased effective precipitation around *ca.* cal. 760-710 BC, 590-520 BC, 180-130 BC, 30 BC - AD 80, AD 210-360, AD 550-670, AD 920-1060, AD 1110-1260, AD 1400-1470, and *ca.* cal. AD 1770-1800. Coom Rigg Moss has also been studied by Charman *et al.* (1999), who compared the plant macrofossil and humification records with testate amoebae reconstructions from replicate cores. One of the cores (CRMI) recorded very little significant changes in the macrofossil content, as they were dominated by monocotylodons and UOM, with very little Sphagna and surprisingly virtually no *Sphagnum imbricatum*. The second core (CRMIV) showed greater variability in all three proxies, suggesting the area the core was taken from was much more sensitive to hydrological changes throughout its growth (*sensu* Barber *et al.*, 1998). Wet shifts were identified in all of the proxies from CRMIV at *ca.* cal. 1550 BC, 950-750 BC, 200 BC, AD 400, AD 600, AD 950, AD 1100, AD 1350, AD 1500, and *ca.* cal. AD 1850, which show some agreement with the wetness changes identified by Mauquoy and Barber (1999a). Charman *et al.* (1999) indicated that moisture changes from the CRMI record can be correlated with CRMIV for the last 1500 years with some confidence. The earlier periods show less agreement, although the overall implication was that the replicability was best over the last 1000 years.

This work has been complemented by the research of Hughes *et al.* (in press) who documented a full Holocene peat archive from Walton Moss, Cumbria. Wet shifts have been registered as commencing at *ca.* 7800 BP, 5300 BP, 4410-3990 BP, 3500 BP, 3170-2860 BP, 2320-2040 BP, 1750 BP, 1450 BP, 300 BP, and *ca.* 100 BP. Notable comparisions can be made between the sites, especially the wet shift in the Dark Ages, around 1400 BP, which has also been identifed by Blackford and Chambers (1991), and documented by Baillie (1994; 1995), who suggested the 'event' may have had a volcanic origin (see Section 2.5.2). The Walton Moss record also exhibits periodicites of *ca.* 1100 years and *ca.* 600 years between wet shifts, which has been suggested as being broadly comparable with the 1470 ± 500 year mean pacing of Holocene Ice Rafted Debris in the north Atlantic (Bond *et al.*, 1997). This indicates that the climatic changes could be linked to changes in oceanic circulation (Barber *et al.*, 1994b; Anderson *et al.*, 1998).

2.4 Holocene climatic changes in Scotland

2.4.1 Peat stratigraphy

The detailed pattern of terrain in Scotland can produce significant variations over short distances, for example substantial areas of upland massifs afford little shelter, enhancing any local factors which ameliorate microclimate (Morrison, 1983). This will influence protection from the wind, orographic rainfall, local cloud cover and hence insolation. It is therefore possible to view past and present Scottish climates as being intensely regional (Lowe, 1993), a factor which has been documented by the regional differentiation of woodland succession and consequent decline throughout the Holocene (Tipping, 1994).

Peat has formed over large areas in Scotland throughout the Holocene, and at present occupies 10.4% of the land surface (Taylor, 1983). Ombrogenous bogs generally developed on lowland sites, specifically plains, valleys or basins. Blanket peats tended to start forming from *ca.* 4000 BP (Bennett, 1984; Dubois and Ferguson, 1985), and did so in abundance on the western side of Scotland. Here the climate is so humid that there is a pronounced tendency for peat to form where the slope lies at less than 15°, as at this angle the ground moisture remains almost constantly in excess of that lost by evaporation (Ratcliffe, 1964).

Early research into Scottish mires was undertaken by Lewis (1905; 1906; 1907; 1911), who analysed stratigraphy and macrofossil assemblages, and related their fluctuations to past climates. Lewis's sites were re-examined by Samuelsson (1910), who stated that two distinct pine stump layers, which Lewis had identified, did not exist, and also doubted many of Lewis's macrofossil identifications. Many of the findings from these early works are contradicted by Pears (1968), who suggested that one or both reached erroneous conclusions, probably due to the enormous area they studied (with comparatively few sites per region), and the variations in local topography studied between basins and surrounding gentler slopes (Pears, 1968).

Since these early insights, very little work appears to have been undertaken on Scottish peat stratigraphy in relation to climatic changes, and the research emphasis has been on pollen analysis (Lowe, 1993). Recent evidence for climatic change from pollen analysis has been fully discussed in the reviews by Birks (1977), Walker (1984), Lowe (1993) and Tipping (1994), with some more recent examples detailed in section 2.4.3. The second section on Holocene climatic change in Scotland concentrates on both macrofossil and microfossil evidence from *Pinus sylvestris* (section 2.4.2), as its expansion and subsequent decline in the mid to late-Holocene may well have been connected with climatic changes and the expansion of blanket peat.

Plenty of ecological studies have been carried out on Scottish mires during the last century, notably during the last 40 years in the Silver Flowe bog complex (e.g. Ratcliffe and Walker, 1958; Boatman and Tomlinson, 1973; Boatman, 1977). Some studies recorded mire stratigraphy (e.g. Durno, 1957; Boatman and Tomlinson, 1977; Boatman, 1983), although very little attention was given to climatic changes.

It appears that detailed peat stratigraphic work in relation to climatic change has only been undertaken in Scotland from the late 1980's onwards, since the falsification of the cyclic regeneration theory by Barber (1981). From a transect encompassing 18 mires between western Ireland and northeastern Poland, Haslam (1987) analysed the changes in macrofossil content across the main humification change, which enabled proxy climatic curves to be constructed for the period between 4500-500 BP in northwestern Europe. At Flanders Moss in Stirlingshire, Scotland, Haslam documented a clear division between the lower humified and upper non-humified stratigraphy. Macrofossil analyses revealed a more complex situation, as the hygrophilous *Sphagnum* section *Cuspidata* could be found in abundance in the lower, more humified, peats. Haslam dated the wet shift at the main humification change to 3480±50 BP, whereas Turner (1965), in a pollen stratigraphic study, dated the lower humified *Sphagnum* peat to 2712±120 BP. The difference in dates was explained by Haslam as the possible resultant of local hydrological factors, which may have affected Turner's marginal site by influencing fresh *Sphagnum* peat initiation.

Stoneman (1993) studied a suite of 10 mires in northern England and southern Scotland across a comparable temperature regime and marked rainfall gradient, utilising plant macrofossil and humification analyses. He inferred significant climatic phases when shifts in bog surface wetness appeared to coincide from three or more sites, and rejected the hypothesis that the spread of shifts were purely random. Those mires that did not provide evidence of climatic shifts at the same time may have had their sensitivity to climatic shifts masked by the presence of *Sphagnum imbricatum* or *Sphagnum* section *Acutifolia* which can both grow over a range of hydrological conditions (see section 6.1). This highlights the need for multiproxy data in such investigations (Barber *et al.*, 1998; Charman *et al.*, 1999). Stoneman was only able to utilise three radiocarbon dates per core, and used an age/depth model to interpolate other dates. The suggested phases of wetter bog surface wetness can be seen in Table 2.1.

As well as raised mires, much of the recent palaeoclimate work on Scottish bogs has concentrated on blanket mires, as these tend to dominate in the wetter regions, specifically along the northwest coast and Hebridean Islands. Chambers *et al.* (1997a) derived a proxy climate record for Talla Moss in the Scottish Borders which was out of phase with changes in the pollen record from the same site. Thus it was inferred that major shifts in peat humification could be attributed to external forcing. A markedly wet phase commenced at Talla Moss at *ca.* 540 BP, which seemingly correlated with the start of the LIA, and this also heralds the Spörer minimum for solar sunspot activity (*cf.* Blackford and Chambers, 1991). Other wet shifts were identified at *ca.* 3455 BP, 2600 BP, 1930 BP and *ca.* 1095 BP. Spectral analysis revealed a cyclicity in the proxy climate record of 210 years, a figure which is comparable with the cyclicity of 206 years identified by Dupont (1986) and the 260 year cyclicity found by Aaby (1976) from raised mires in Denmark.

Two recent papers by Anderson *et al.* (1998) and Anderson (1998) provide a number of inferences about climatic changes in Scotland throughout the mid to late-Holocene. Anderson *et al.* (1998) investigated five sites in northern Scotland which showed evidence of a major transition in climate, radiocarbon dated to between 3900-3500 BP. The mires used in the study were topogenous bogs which formed in basins. Anderson (1998) suggested that such bogs can be sensitive to climatic change if they are predominantly ombrotrophic and isolated from nearby streams. However, this approach can be problematic, as it is difficult to identify climatic changes if streams had been inflowing into the basin earlier in the Holocene, possibly influencing the sensitivity of the mire to climate change. Some of the inferences made from microfossil analyses may also be ambiguous. Anderson (1998) states that, although the interpretation of fluctuations in Cyperaceae pollen is not straightforward, because these species can grow over a very wide range of moisture conditions (see Section 6.1), a large change from Ericaceae pollen to an increase in Cyperaceae pollen could be interpreted as a shift from drier to wetter bog conditions. However, other evidence is needed to corroborate this assumption, as the change in pollen assemblages noted by Anderson (1998) may just be indicative of ecological changes on the mire, with the sedge pollen indicating a dominance of *Scirpus cespitosus* or *Eriophorum vaginatum*, both of which can grow on dry mire surfaces. These potential problems notwithstanding, major changes in surface wetness from macrofossil and microfossil analyses have been identified by Anderson *et al.* (1998) and Anderson (1998).

The major transition in climate between 3900-3500 BP was implied to be synchronous with changes inferred from other regions (Anderson *et al.*, 1998). Proving that changes are synchronous can be problematic, especially when employing radiocarbon dates alone, as potential regional lags in the climatic system may not be resolvable. The nature and magnitude of the climatic shift, however, did seem to be highly significant. Whilst summarising data from many sites, Barber (1982) suggested that there had been a major climatic change over Europe between 3950-3850 BP (see Table 2.1).

Other climatic shifts identified by Anderson *et al.* (1998) and Anderson (1998) included three other regional shifts to wetter bog conditions and two regional shifts to drier conditions. Best estimates for the wet phases spanned *ca.* cal. 5120-5070 BP, 3340-3270 BP and *ca.* cal. 940-800 BP. Dry shifts were estimated between *ca.* cal. 4330-4120 BP and *ca.* cal. 1480 to 1340 BP. This last phase shift is especially interesting as at 1400 BP there appears to have been a major wet shift in northern England (Blackford and Chambers, 1991; Tallis, 1994; Mauquoy, 1997). Using pollen evidence from *Alnus glutinosa*, Bennett and Birks (1990) suggested that climatic changes do not appear to have been synchronous between Scotland and England. It therefore seems possible that the discrepancy between the climatic shifts of Anderson *et al.* (1998) and dates for climatic changes in northern England could be due to asynchronous events between Scotland and northern England that Bennett and Birks (1990) suggested from their pollen data. However, even though Anderson (1998) suggested that humification shifts may have occurred within a hundred years, the lack of precision associated with radiocarbon dating (Pilcher, 1991) may well account for the difference of 'only' a hundred years between the dates.

The climate in Scotland will be particularly influential in areas where vegetation grows at the limit of its range, and where stresses within vegetational communities will be high. One of these areas is the Cairngorms, where the climate is the dominating natural force governing the environment (McClatchey, 1996). A water shedding blanket bog in the northern Cairngorms, Moine Mhor, was found to contain well preserved plant macrofossils, and subsequent analyses suggested that the surface wetness of the bog had fluctuated repeatedly over the last 2000 years (Barber *et al.*, 1999). The proxy effective precipitation record from the Cairngorms could also be correlated with another similar reconstruction from a raised bog at Fallahogy, Northern Ireland, the largest change of which was driven by climate in the period AD 1700-1850, which was interpreted as the LIA (Barber *et al.*, in press). It was suggested that when the two bogs were in phase with each other they were responding to a spatially coherent temperature signal, whereas when they were out of phase they may have been responding to a non-coherent rainfall regime.

Peat humification records have recently been compared with another terrestrial palaeoclimate proxy, in the form of stalagmite luminescence data, in northwestern Scotland (Baker *et al.*, 1999). The two records have been correlated over the last 2500 years, and have been found to compare well with historical climatic events, showing wetter conditions at 150-400 BP and 500-600 BP which coincide with the Maunder and Spörer sunspot minima and climatic deteriorations recognised by Blackford and Chambers (1991) and Chambers *et al.* (1997a). Precise correlations between records is difficult however, as the errors on both the AMS and U-Th dates are ±100 years. Taking into account the dating errors, spectral analysis has been performed, and a clear periodicity of 80-125 years was identified in both records, and was potentially attributed to solar variability, notably the well known 80-100 year Gleissberg cycle (Frick *et al.*, 1997). Baker *et al.* (1999) also speculated whether the wet phases could be solely due to a more southerly track of the polar front as a result of solar variability, or if this mechanism could be combined with the internal ocean-climate variability.

2.4.2 The late-Holocene history of *Pinus sylvestris*: coupled with climatic change and blanket peat expansion?

Proxy climatic indicators are able to demonstrate the rapid warming at the start of the Holocene, but the scale of climatic variations during the rest of the Holocene is harder to evaluate. This becomes especially apparent during the late Holocene, when potential anthropogenic effects on the vegetation in Scotland make it difficult to establish possible climatic changes, especially where the landscapes tended to be unwooded (Vorren, 1986). In the Cairngorms, about 3000 years ago forests declined regionally as a result of anthropogenic pressures, and peatland ecosystems increased in areal extent (Bennett, 1996). This decline in forest, and subsequent increase in peatland ecosystems occurred in many other areas in Scotland, and especially the Western Isles and Shetland (Bennett *et al.*, 1990; Bennett *et al.*, 1992). However, these vegetation changes were not all necessarily due to anthropogenic changes, especially as in many places (particularly the more remote islands) vegetation changes were taking place when there is no evidence for human occupation (see Bennett *et al.*, 1990). This implies that another factor(s) was influencing vegetation changes, and the main overriding mechanism to be considered must be climate.

Possibly one of the best proxy indicators of environmental change in Scotland during the late-Holocene may be derived from analyses of the macrofossil and microfossil remains of *Pinus sylvestris* (Scot's Pine) and their relationship with blanket peat formation. *Pinus sylvestris* is potentially a good indicator of environmental change because it appears to have expanded rapidly during the mid-Holocene, possibly in two phases (Bridge, *et al.*, 1990), but underwent a sudden decline around 4000 BP (Bennett, 1984; Bridge *et al.*, 1990), perhaps due to the initiation or rejuvenation of blanket peat formation.

The natural indigenous pinewoods of Britain may be regarded as being confined to Scotland, although they have been widely planted in the British Isles (Carlisle and Brown, 1968). Pinewoods can tolerate a wide range of climatic conditions, and it is possible that the northern distribution is limited by low temperatures during either pollination or fertilization (Carlisle and Brown, 1968), and this has been found to be more important than rainfall in influencing growth (Grace and Norton, 1990). *Pinus sylvestris* prefers light, freely drained soils, and although they have been planted successfully on deep peats, they tend to suffer on wetter sites which are well exposed and elevated, and susceptible to fire (Carlisle and Brown, 1968).

The fossilised remains of pine stumps have been found in blanket mires throughout Scotland. These were first studied in detail in the 1960's and 70's by Pears (1968, 1969) and Birks (1975), after the inital surveys by Lewis (1905-7; 1911) and Samuelsson (1910). Pears attempted to reconstruct the altitudinal limits of the tree line by mapping the distribution of pine macrofossils preserved in the blanket peats of the Cairngorms. Two layers of pine stumps were identified, although there was a lack of dating control on both layers (Pears, 1975a), and the assumption that the pine remains at the highest altitudes are synchronous at each site remained problematic (Pears, 1975b). Research by Birks (1975) in the Cairngorms arrived at similar conclusions, confirming that the pine (and birch) remains did not form synchronous horizons. Birks suggested that in some instances the pines had been killed by increased waterlogging, whereas at another site the cause of death appeared to have been fire, leading Birks to infer that there were no overriding climatic changes controlling bog development from *ca.* 7000-4000 BP. However, a review of pine stumps in blanket peats throughout Scotland by Birks revealed a striking group of dates between 4500-4000 BP which corresponded with substantial declines in pine pollen from a number of sites. This suggested a possible climatic deterioration at this time, potentially associated with a increased expansion of blanket peat (Birks, 1975).

The former growth of pine on peat bogs has been interpreted as indicating that the upper layer of peat was sufficiently aerated due to dry surface conditions, whereas the preservation of pine stumps indicated somewhat wetter conditions in the surrounding peat (Birks,1975). However, many of the recorded stumps have been dated after *Pinus sylvestris* was thought to have declined from pollen records. Gear and Huntley (1991) presented two theories for the presence of pine stumps. They first suggested that at the time of preservation there was only a sparse population of *Pinus sylvestris*, growing only in particularly favourable sites, and therefore little pollen was available for preservation, but the stumps were preserved during favourable conditions. Their second theory suggested that regional pine forests developed only for brief periods of time, which had been overlooked by conventional pollen studies. Gear and Huntley (1991) supported this argument by illustrating a fine resolution pollen stratigraphic record (containing an interstratified pine stump in the peat exposure), and found that pine pollen was greater than 30% of the terrestrial pollen sum for only 68mm in a 123mm set of samples, with the greatest values spanning only 7 and 15mm of peat respectively.

Isotopic analysis on macrofossils of *Pinus sylvestris* has been used to suggest a relationship with precipitation and pine growth. Dubois and Ferguson (1985) used cellulose from pine stumps for deuterium isotopic analysis, with extremely low deuterium values indicating periods with very heavy rainfall, which were identified as 'pluvial phases'. Dubois and Ferguson (1985) found that pluvial phases occurred around 7300 BP, 6200-5800 BP, 4200-3940 BP, and at about 3300 BP. No pines younger than 3200 BP have been found above 530m, and Dubois and Ferguson (1985) suggested that trees at lower altitudes were not preserved as the destruction of less dense wood occurred before the trees could have been enveloped by peat. The onset of a pluvial phase at around 4200 BP suggested that blanket bogs became rejuvenated in the Cairngorms, and developed on a large scale, as the soil became waterlogged due to increased precipitation, and initiated the increase in peat formation (*cf.* Birks, 1975). A spread of blanket bogs in northern Britain at *ca.* 4000 BP, possibly initiated by widespread climatic change, which severely reduced the range of *Pinus sylvestris*, has also been suggested by Bennett (1984).

Similar dates for the expansion of blanket peat have been suggested by Bennett *et al.* (1990), who proposed that a major spread of blanket peat may have begun around 5500 BP on Southern Uist of the Western Isles; and that by 4300 BP the blanket peat was dominant, as by this time the abundance of tree and shrub pollen had decreased from 60% to 10%, and was replaced by pollen of *Calluna*, Gramineae, and Cyperaceae. The status of *Pinus sylvestris* on Southern Uist, Western Isles, was described as uncertain.

Bridge *et al.* (1990) used the pluvial episodes described by Dubois and Ferguson (1985), and compared them with major rises in pine pollen. It appeared that pine grew successfully on mire surfaces in the western Highlands from 6800-4000 BP, although during this time there were probably variations in density, size and longevity of the trees, possibly in relation to changing environmental conditions. This was emphasised by local variations in pine history found by Bridge *et al.* (1990), including a major expansion of pine woodland at *ca.* 6800-6600 BP, followed by a significant reduction in pine cover between 6600-4970 BP, and a second expansion of pine around *ca.* 4970-4660 BP. Two distinct phases of pine growth were also recorded between *ca.* 6500-3000 BP at low altitudes from Rannoch Moor, western Scotland (Ward *et al.*, 1987). Bridge *et al.* (1990) found that there was a time lag between pine reduction and climatic change, with the troughs in macrofossils seeming to follow the pluvial episodes. This could have indicated that the pine woodlands were significantly reduced by a climatic change, or that the macrofossils have not been preserved, or are yet to be discovered. However, it was suggested that the density and distribution of pine varied in Scotland probably as a result of variation in precipitation levels.

A brief phase of local pine forest between *ca.* 4500-4000 BP, from a mainly treeless landscape, in the Flow Country, northern Scotland, has also been noted by Charman (1994), although the reasons for this sudden growth of pine are unclear from the pollen data. Charman suggested that it may be indicative of the mire surface drying out before 4500 BP, enabling pine to colonise. The sudden decrease in pine at 4000 BP is in accordance with other records (Bennett, 1984; Bridge *et al.*, 1990), with local pine growth in the area having ceased around 3920 BP, and pine had disappeared almost completely by *ca.* 3400 BP.

Changes in growth of pine have also been documented from Scandinavia. Kullman (1987) used *Pinus sylvestris* megafossils to infer that the pine tree-limit in southern Sweden was stable from before 8000 BP, and remained at high altitudes until 4000-3500 BP. Kullman argued that climatic shifts of relatively low magnitudes are not able to change the ranges of pine, implying that if climate was responsible (rather than biological processes) then the event must have been of considerable magnitude. This is similar to the evidence presented by Anderson *et al.* (1998) of a major climatic event between 3900 - 3500 BP. Dendrochronological inferences have also been made from Scandinavia using *Pinus sylvestris* chronologies, which have been able to show that the LIA was not a synchronous event throughout northwest Europe (Briffa *et al.*, 1990). Recent research documented warm summers in eastern Norway around AD 1510 (the same date as a historical tephra eruption from Hekla which has been identified in Scotland), AD 1770, and AD 1940, whereas cool summers occurred around AD 1600, AD 1710, and AD 1800 (Kalela-Brundin, 1999).

Peat development on the Shetland Islands seems to have occurred slightly later than 4000 BP. A study from Lunnasting indicated that at 3900 BP there was a temporary decline in woodland, followed by a recovery for *ca.* 400 years (Bennett *et al.*, 1992). By 3120 BP, however, Bennett *et al.* (1992) showed that the landscape appears to have been treeless, and was dominated by acidic heathland and blanket peat. Birks and Line (1992) suggested that the decline in palynological richness from 3100 BP may have been due to a simplification of the vegetational mosaic during the development of blanket peat. Bennett *et al.*, (1992) point out that the changes in vegetation may not be due to climatic change, but that soils may have become leached, then podsolized, initiating peat growth. Thus the increase in mire vegetation from 4800 BP could be due to natural acidification, and even though there is no archaeological evidence for human settlements before 5000 BP, anthropogenic effects since then could have affected the magnitude of any naturally occurring processes.

It was not always the case that *Pinus sylvestris* was able to dominate forested areas of Scotland before declining at around 4000 BP. At Claish Moss in northwest Scotland *Pinus* never achieved dominance in the pollen spectra, as *Quercus* and *Ulmus* tended to dominate (Moore, 1977). Moore (1977) suggested that the resistance of forests to the dominance of *Pinus* may be due to the oceanicity of the western sites. Bennett (1984) noted that at *ca.* 9000 BP competition from other species (for example, *Corylus avellana* type, *Ulmus*, and *Quercus*) restricted the habitats available to *Pinus sylvestris*. This may also have been the case in stopping *Pinus sylvestris* colonising Shetland, the remote outer islands and parts of southwest Scotland, as *Quercus* may have arrived before *Pinus* *sylvestris*, and formed sufficiently dense woodland to prevent colonisation from *Pinus sylvestris* (Bennett, 1984). Today *Pinus sylvestris* is only native to the Scottish Highlands, and where it grows it is often the dominant forest tree (Bennett, 1984).

Other methods have also been used in order to extract environmental information with respect to the pine decline. Tephra studies have been used in order to ascertain if the Hekla-4 eruption had a causal affect on the decline of pine, and also to provide greater precision when dating the decline (Blackford *et al.*, 1992; Hall *et al.*, 1994b; section 3.2.1). Dendroclimatological studies can also assign detailed information to specific subfossil wood samples, such as fire episodes (e.g. Chambers *et al.*, 1997b) and periods of environmental stress, potentially related to climatic change (e.g. Lageard *et al.*, 1999). Investigations of bog pines in Scotland were initially limited to the last few centuries for detailed climatic reconstructions (e.g. Hughes *et al.*, 1984) and floating chronologies (e.g. Ward *et al.*, 1987), but pine macrofossils from Cheshire have recently been cross-matched to subfossil oak records allowing a more accurate dating of pines on bogs, which will aid precise correlations and comparisons of proxy-climate data between sites (Lageard *et al.*, 1999).

The above evidence indicates that the decline of *Pinus sylvestris* was not synchronous across Scotland, although evidence does exist for the initiation of peat growth around *ca.* 4000 BP. However, even if this event is synchronous across Scotland, and this is suggested as unlikely (Moore, 1993), it may not have been due to increased rainfall, but lowered rates of evaporation (Whittington and Edwards, 1997). A set of modelling simulations by Kutzbach and Guetter (1986) concentrated on the changing orbital parameters and their relationship with climate over the last 18,000 years. They have suggested that between 12,000-6000 BP the tilt of the earth's axis was greater than at present and perihelion took place in the northern summer. An increase in anticyclonic conditions would have occurred, and although winters would have been cooler, the summers would have been warmer than present, with a comcomitant increase in evaporation rates (Whittington and Edwards, 1997). Thus, by 4000 BP, as the orbital parameters were changing, comparative evaporation rates may have been lowered, resulting in peat formation where local conditions allowed.

2.4.3 Other evidence for late-Holocene climate change

Geomorphic evidence can also be used to infer climatic changes in the late-Holocene. Ballantyne (1991) commented that many authors assume that landscape instability is automatically associated with climatic deterioration, especially the onset of enhanced erosion in the Scottish Highlands being attributed to climatic deterioration during the LIA. Landscape instability has also been used to suggest climatic cooling during the LIA from areas in Ireland (P. Coxon, pers. comm.). However, Innes (1983) used lichenometry to post-date some debris flows in the Scottish Highlands to after the LIA, and suggested that during the nineteenth and twentieth centuries an increase in debris flows in the Highlands resulted from land use changes (including increased grazing and burning practises), and not climatic changes and progressive weathering. Hinchliffe (1999) examined the debris reworking from relict talus slopes in northern Skye, and through an analysis of buried palaeosols, found that different sectors underwent reworking at *ca.* cal. 5900-5600 BP, 2300-1700 BP, and *ca.* cal. 700-500 BP. No evidence for anthropogenic burning or management of local vegetation was found, and in the absence of any direct causal relationship between climatic change and reworking (Ballantyne, 1991), the coincidences between the dates and inferred climatic deteriorations at these times require further research (Hinchliffe, 1999).

Sedimentological (including colorimetric humification) and pollen stratigraphical techniques from a raised bog at Burnfoothill Moss, East Dumfriesshire, have produced results from which potential climatic changes could be inferred, notably a climatic deterioration around the proposed time of the LIA (Tipping, 1995). A very wet phase has been identified just before *ca.* 400 cal. BP, interpreted by Tipping as the initial stages of the LIA, which is preceded by a dry phase between *ca.* 600-400 cal. BP, possibly an indication of the MWP. Other climatic shifts identified from this site included a wet shift at cal. 5250 BP, followed by a more substantial wet shift at cal. 4000 BP (coherent with many other studies), and also a dry shift at cal. 1900 BP, which ends at *ca.* cal. 1200 BP. Tipping identifies a potential weakness in his approach, as some of the shifts from the proxy indicators appeared to contradict each other, stressing the need for a multiproxy approach. Another problem in his approach occurred when each proxy indicated a unidirectional shift, but not synchronously. Tipping suggested this may be due to natural lags in response time within the peat system, which has implications for intra site correlations.

There has been a large amount of interest in determining periodicities in recent years, and hence forcing mechanisms, within Holocene palaeoecological data (from Aaby 1976 through to Baker *et al.*, 1999). Oliver *et al.* (1997) presented detailed analyses on a pollen profile from a peat–filled kettle hole at Pickletillem near Fife in eastern Scotland. The analyses revealed strong spectral peaks, with frequencies that corresponded to the stratigraphical structure in the core. This suggested that the pollen assemblages vary cyclically, from woodland to grassland alternating in a quasi-cyclic manner. Significant periodicites were identified at 360 years and 800 years, the latter of which corresponds with that estimated by Barber *et al.* (1994b), and the former can be associated with the period found by Wjimstra *et al.* (1984). The potential significance of these cycles is discussed in section 2.5.

Evidence for late-Holocene climatic change in Scotland may also be derived from lake sediments. The Cairngorms are host to many enclosed lakes, one of which, Lochan Uaine, has been used to construct climatic changes using a wide range of physical, chemical and biological methods. A low amplitude cycle of organic matter production over the last 4000 years with a mean periodicity of 200 years was identified by Battarbee *et al.* (1996). It was hypothesised that the cycles were controlled by variations in temperature which dictated the length of the ice-free season, and influenced the amount of organic matter generated. Brooks (1996) constructed a 3000 year record of changes in chironomid assemblages from Lochan Uaine, which was used as a proxy for environmental change. A significant change in the assemblage was recognised around 2500 BP, associated with changes in trophic conditions or climatic cooling. Brooks states that sampling at a finer resolution and comparisons with the results from other proxy records should elucidate a more complete picture of climatic change in the Cairngorms. Holocene chironomid reconstructions have rarely been used in Britain, and it is suggested that by utilising a temperature proxy indicator for which a transfer function has been derived, such as chironomids (Lotter *et al.*, 1999; Olander *et al.*, 1999; Brooks and Birks, 1999), coupled with mire surface wetness reconstructions (e.g. Woodland *et al.*, 1998; Mauquoy and

Barber, 1999a; Charman *et al.*, 1999; Hughes *et al.*, in press) a greater understanding of the proxy signal in peat bogs and the climatic forcing mechansisms could be achieved.

2.5 Forcing mechanisms

Scientists are always searching for causal mechanisms for climatic changes, especially when there is a suggestion that cyclicity can be detected in a palaeoenvironmental reconstruction. A number of forcing mechanisms are thought to have influenced Holocene climatic changes, the scale of which has been referred to as "Little-ice age-scale variations" (Bartlein, 1988). The most commonly identified 'external' mechanisms are solar variability and volcanic activity, which are thought to modulate 'internal' mechanisms, such as natural oscillators (e.g. oceanic circulation) and internal feedbacks.

Forcing mechanisms are thought to operate on different temporal scales, and a spectrum of climatic variance was presented by Mitchell (1976) which documented forcing mechanisms operating at different frequencies (Figure 2.1). The 'Gap' identified by Mitchell related to millennial scale changes, and has been suggested as driving Holocene climates. A forcing mechanism which operates over these frequencies has recently been identified in North Atlantic sediments and corresponds to a cyclicity of 1470 ± 500 years (Bond *et al.*, 1997). The mechanism has been linked to thermohaline circulation in the North Atlantic, coinciding with increases in Ice Rafted Debris. During deglaciation the cycle is thought to have regulated iceberg discharges into the North Atlantic, regulating the production of North Atlantic Deep Water. A model simulation by Bond *et al.* (1997) also suggested that large amplitude oscillations in thermohaline circulation can be induced by increasing fresh water fluxes to the ocean, which can be related to Northern Hemisphere ice sheet decay. The importance of the ocean in modulating climatic variability has been known for some time, although the aspects of the sun-climate relationship, and the influence of the oceans, is still the subject of much ongoing research (e.g. van Geel *et al.*, 1999).

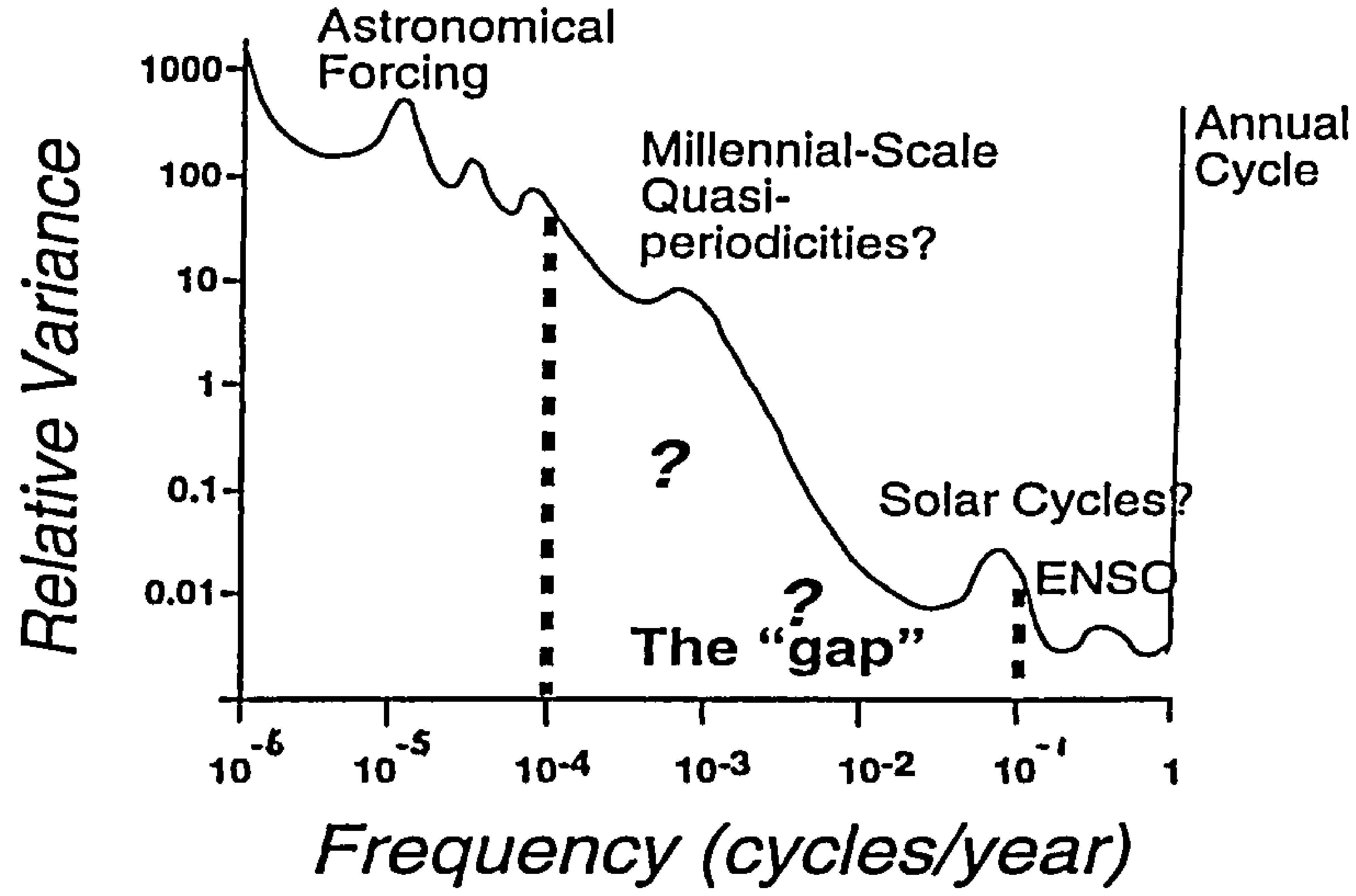


Figure 2.1 Spectrum of climatic variance (after Mitchell, 1976, with adaptations by J. Overpeck, pers. comm.).

2.5.1 Solar variability

Variations in the solar radiation that reach the lower atmosphere and the earth's surface are thought to affect the circulation of the atmosphere and oceans. Variations in the surface appearance of the sun have been apparent since the times of Galileo and have allowed long records of solar activity to be established. Long periods of constant variation have been recognised, for example anomalously low solar activity (low sunspot numbers - dark patches on the sun which burn at only *ca.* 5000^{0}K instead of the more usual temperature of *ca.* 6000^{0}K) have been identified throughout the last 1000 years, with notable episodes such as the Maunder, Spörer and Wolf Minimums (Blackford and Chambers, 1995).

Correlations between solar variability and climate change have been investigated by utilising the isotope ^{14}C as a proxy for solar variations (e.g. Stuiver *et al.*, 1995; 1997). The isotope ^{14}C should be of the same proportions in living matter as in atmospheric CO_2. However, variations in ^{14}C have been determined from past levels of CO_2 which have been caused by geomagnetic and solar modulation of the cosmic ray flux (Stuiver and Braziunas, 1993). Hence records have been determined for the changes in ^{14}C over time, and various periodicities of solar variability have been discovered. Van Geel *et al.* (1996; 1998) have correlated the climatic deterioration which occurred across northwest Europe around 2650 BP with a sudden and sharp rise in ^{14}C. If this correlation is validated by other proxy records and known climatic regimes then it could appear that ^{14}C levels could act as a climate proxy.

Other attempts have been made to correlate ^{14}C anomalies with climatic records. Wigley and Kelly (1990) compared climatic records from both hemispheres with ^{14}C anomalies and found a statistically significant correlation. However, they used a simple energy-balance climate model to show that the mean reduction of solar irradiance, when ^{14}C was at a maximum, would have to have been between 0.22 and 0.55% over a period of roughly 200 years to cause these cool periods. Kelly and Wigley argued that these changes are up to an order of magnitude larger than have been observed by satellite measurements. However, Lean *et al.* (1995), using a long term solar variability record calculated from the Schwabe (11 year) irradiance cycle, suggested that total solar irradiance during the Maunder Minimum may have been reduced by 0.24%, a figure just within the boundaries suggested by Wigley and Kelly (1990).

Karlén and Kuylenstierna (1996) used climate data from reconstructed glacier fluctuations and tree-line shifts over Scandinavia to compare with an index of solar variability (^{14}C anomalies). A lag of 150 years was used for the pine tree record as they can often survive for some time after the onset of cooler climates, and this showed that lower temperatures in Scandinavia correlated with peaks in ^{14}C 17 out of 19 times over the last 9000 years. This suggests that solar variability is an important mechanism for determining climate change, even if it is coupled to other mechanisms, such as oceanic circulation.

Stuiver and Braziunas (1993) constructed two $^{14}CO_2$ records to test spectral relationships, including a single year record from 0-450 cal. BP, and a Holocene bidecadal record from 0-11,400 cal. BP. The single year record displayed periodicities of 2-6 years, suggesting El Niño Southern Oscillation (ENSO) perturbations, and periodicities of 10-11 years, suggesting solar modulation (the 11-year sunspot cycle) or North Atlantic thermohaline circulation instability. The Holocene bidecadal record identified a periodicity of 512 years, which is also thought to be related to thermohaline circulation instability, with North Atlantic deep water forming near the start of the Younger Dryas interval rather than at the termination. A 206-year cycle has also been identified, which has been attributed to solar modulation, and is possibly modified by a climatic (or oceanic) response. These cycles are similar to the ones found by Aaby (1976) from raised bogs in Denmark, as well as the periodicities found by van Geel (1978), Dupont (1986), Barber *et al.* (1994b), Mauquoy (1997), and Chambers *et al.* (1997a).

2.5.2 Volcanic activity

Large volcanic eruptions are known to affect the global climate for at least one or two years (Baldwin *et al.*, 1976; Rampino and Self, 1982). A prime example is the eruption of Mount Pinatubo in June 1991 which caused the largest perturbation this century to the particulate content of the stratosphere (McCormick *et al.*, 1995). This almost immediate climatic forcing is generally accepted, but the evidence for long term change is weaker (Crowley and Kwang-Yul Kim, 1993). It would, however, appear possible that eruptions larger than experienced in modern history may have affected the climate for longer, as suggested by evidence from dendrochronology (Baillie, 1994). The mechanism leading to narrow growth rings on trees after such an event is a short term shift to cooler, wetter conditions associated with a negative pressure anomaly over the British Isles (Kelly and Sear, 1984). It would therefore be possible that these events could have been recorded as climatic deteriorations in peat stratigraphy, as the short term wetter/cooler conditions may have pushed specific species over an ecological threshold (*cf.* Conway, 1948).

The mechanisms by which volcanoes are able to cool the global climate have been modelled from real examples, and the eruption of Mount Pinatubo from June 1991 onwards has been one of the most studied. The eruption introduced large amounts of sulphate aerosols into the stratosphere, which resulted in a significantly correlated decrease in stratospheric ozone, thus reducing the amount of energy available to heat the lower atmosphere, producing a cooling effect (Kinnison *et al.*, 1994).

Past volcanic eruptions have been documented as acid spikes in ice cores, and as very narrow tree ring widths in dendrochronological records. The first major link with volcanic events and climatic change occurred when large peaks in sulphates were identified from the Camp Century and Crête ice core records in central Greenland, and compared to a temperature index, showing that clustered eruptions had a considerable cooling effect on the climate (Hammer *et al.*, 1980). The results also appeared to correlate well with the California tree ring record of LaMarche (1974). More detailed correlations in recent years have been undertaken between acid spikes in ice cores and volcanism throughout the Holocene (e.g. Zielinski *et al.*, 1994).

Further research into tree ring data suggested that at various points throughout the late-Holocene, trees had been damaged by severe frosts (indicated by narrow rings), which had been apparently caused by stratospheric dust veils produced by major volcanic eruptions (Baillie, 1995). Specific eruptions appear to have been large enough to cause regionally/globally cooler conditions (Scuderi, 1990), with specific events detected in tree ring records and documentary records from around the globe, notably the Santorini eruption in the early 1600's BC (Baillie and Munro, 1988 - the exact date is still argued over - Baillie, 1995), and an eruption at AD 536 (Baillie, 1994). It is known that significant cooling can occur within months of a major volcanic eruption (Kelly and Sear, 1984; Sear *et al.*, 1987), so it was not surprising to find evidence in the tree ring records the following year after a documented eruption. However, it is not only atmospheric temperature which appears to be affected by large volcanic events. Recent research has also documented an increase in storminess that follows volcanic eruptions from a 200 year record of gale frequency at Edinburgh (Dawson *et al.*, 1997).

The two dates identified as cooling events from the dendrochronological records at AD 536 and in the early 1600's BC (see Baillie, 1995) can also be identified as major deteriorations in climate from many peat stratigraphic records (Table 2.1). The chronology from peat bogs is far less precise than the dendrochronological records, and so direct correlations cannot be made. However, if there were large volcanic eruptions at these times, which resulted in major global cooling events, there is no reason why evidence for cooling cannot be inferred from peat stratigraphy, as bogs are thought to respond to temperature changes in a coherent manner over large spatial scales (Barber *et al.*, 1999).

2.6 Synthesis - changes in NW Europe and Scotland

An investigation into proxy climatic records from peat stratigraphy has shown that some of the phase shifts are well defined across northwest Europe (Table 2.1). Peat stratigraphic climatic shifts for the late-Holocene have been previously summarised by Barber (1982). He suggested a climatic deterioration to have occurred at *ca.* 3950-3850 BP, followed by a more pronounced wet shift at *ca.* 3450-3350 BP. The main climatic deterioration in the late-Holocene (excluding the LIA) appears to have been between *ca.* 2850-2550 BP, and as can be seen above this date is picked out by the majority of climatic reconstructions from peat stratigraphy. Barber (1982) also suggested that there is some evidence for further decline around *ca.* 2050 BP. This appraisal from 1982 took the majority of dates for these phase shifts from sites in northern England, Ireland, and on the continent (especially Scandinavia and the Netherlands) because this is where most of the climatic reconstructions on raised bogs had been undertaken. As Barber (1982, p.110) comments:

> "In particular, there is a need for more continuous data from north-west Scotland, Ireland and Wales to compare these more oceanic areas with the work from Cumbria, Denmark and the Netherlands."

From the data collected since 1982, it is now apparent that much more detail can be discerned from climatic reconstructions from peat stratigraphy, culminating in local and more regional palaeoclimatologies. Without precise dating it is currently impossible to say whether the changes are synchronous, but specific regional differences can be identified in Table 2.1. There appears to be multiple evidence for a shift to wetter, more humid conditions soon after 4000 BP, most likely between 3900-3500 BP, although the exact timing for this climatic deterioration appears to be smeared, probably as a result of the dating techniques. It is worth noting that there is no evidence in Scotland for an earlier climatic deterioration between 4450-4250 BP (2500-2300 BC), although this climatic phase is well documented in records from northern England and northwest Europe. There is very strong evidence for the well accepted climatic deterioration which occurred at the end of the Subboreal, *ca.* 2650 BP, throughout northwest Europe, with the exception of northern Scotland. It may be possible that the climate was too wet in Scotland and the

Table 2.1 Comparison of the timing of climatic deteriorations inferred from mire wet shifts at a selection of sites across Europe (ca. cal. years AD/BC). Adapted from Hughes *et al.* (in press).

Region	Author	AD					BC							Cycles
Scotland NW Scotland	Baker *et al.* 1999	1550-1800 1350-1450												80-125
Moine Mhor	Barber *et al.* 1999	1670-1810												
5 basin sites NW Scotland	Anderson 1998			1010-1150						1390-1320 1950-1550		3170-3120		
Talla Moss Scottish Borders	Chambers *et al.* 1997a	1410		850		250 20	320	650		1510				210
southern Scotland	Tipping 1995a	1550			750		100				2050	3300		
Scottish Borders and Cumbria	Stoneman 1993	1335-1615		930-1120		60-150	130-10	530-270	840-730 1235-1080	1690-1600				
Cairngorms	Dubois & Ferguson 1985									1600	2850-2430		5150-4700	
N England Walton Moss	Hughes *et al.* in press	1850 1600			500	200	370-90		1220-910	1550	2460-2040	3350 3950	5850	1100 600
Bolton Fell Moss	Hughes *et al.* submitted			950		50			1050-850		2750-2550			
Coom Rigg Moss	Charman *et al.* 1999	1850 1500	1350 1100	950	600 400		200		950-750	1550				
Coom Rigg Moss Felecia Moss	Mauquoy and Barber 1999a	1770-1880 1400-1470	1100-1260	920-1060	550-670	210-360	30-AD 80 180-130	590-520	760-710					
Bolton Fell Moss	Barber 1981, Barber *et al.* 1994b	1740 1450		950 780	650		250		950	1650	2400			800
Rusland Moss	Dickinson 1975		1145		400				740					
UK transect	Blackford & Chambers 1991				550-740									
Ireland W Ireland	Blackford & Chambers 1995	1410-1540 1660-1720												
Ireland	Mitchell 1956				500				800	1500				
Netherlands Bourtangersveen	Dupont 1986					10				1350 1700	2500	3350		206
Engbertsdijksveen	Van Geel 1978, Van Geel *et al.* 1996								900 1070 850-760	1800	2400	3500 3900 4500	4850 5200	150-200
Denmark Draved Mosse	Aaby 1976	1500	1290 1090		450	250	350		1050	1450	2050 2350 2650	2900 3100 3450		260
Norway 20 bogs	Nilssen and Vorren 1991	1530	1230 1100	810	550	270 20	280	500	850 1170	1420 1830	2330 2740	3400		
Sweden	Granlund 1932		1200		400				650	1200	2300			
NW Europe	Haslam 1987			800		100		600	1100		2250			

peat bogs were not sensitive to such a change. A wet shift is also evident from virtually all sites with palaeoclimatic reconstructions at 1400 BP, but this shift is not apparent in any of the evidence from Scotland. In fact, Anderson *et al.* (1998) suggest a dry shift from *ca.* 1500 BP in northwest Scotland. Therefore either the climate was changing synchronously very rapidly across northern Britain, or there appears to be a difference in climatic shifts between northwest Europe and Scotland at this time. More data are clearly needed in order to test these hypotheses.

The LIA and MWP appear to be represented in most of the palaeoclimatic records discussed. There is some suggestion that the LIA was a global phenomena, certainly wider than just Europe; evidence for these climatic regimes has been found as far from Europe as the northern Sargasso Sea (Keigwin, 1996). However, evidence shown in Table 2.1 suggests that the record of the LIA did not necessarily occur over northwestern Europe at the same temporal resolution. Barber (1981) suggested that the periods of wettest bog surfaces in the last millennia occurred between AD 1320-1485 and AD 1745-1800, which showed a striking agreement with Lamb (1977b). In Ireland there is evidence to suggest two major phases, initiated between AD 1410-1540, with a second pulse following between AD 1660-1720; a dry phase occurring in-between (Blackford and Chambers, 1995). Mauquoy (1997) suggested that the LIA in Cumbria occurred between AD 1420-1800, while Chambers *et al.* (1997a) suggested that the wet phase which occurred in the Scottish Borders at AD 1410 could be the start of the LIA. In Dumfriesshire, Tipping (1995) suggested the initial stages of the LIA occurred just before 400 cal. BP (*ca.* AD 1550), a little later than suggested for the Borders and further south into northern England. In Fennoscandia, the LIA only seems to have occupied a period from AD 1570-1650 (Briffa *et al.*, 1990). The fact that two stages of the LIA are picked up in Ireland and further east into northwestern Europe the cool climatic event becomes harder to detect, infers the crucial role of the North Atlantic on the climate of northwestern Europe, although regional variations occur.

Whittington (1985) suggested that from the start of the fifteenth century to the nineteenth century Scotland experienced its coldest weather since the last of the Devensian ice was present, with the exception of a warmer period between AD 1500-1550, a period which was also detected in Cumbria by Barber (1981). Whittington (1985) attributes this cold weather to the circumpolar vortex developing meridional loops, which when well developed can detach from the circumpolar vortex and form slow moving or static cells of high or low pressure. In turn, these cells can dislocate the westerly airflow, and dislocate the polar front along which the mid-latitude depressions travel and which cool the seas of the North Atlantic. Whittington (1985) proposed that these conditions became 'normal' during the LIA. Lamb (1977) stated that during 80-85% of summers between AD 1688-1700 the paths of centres of cyclonic depressions passed over or near Scotland (56^o - 62^o N) causing violent storms.

The LIA may only have been confined to a period of 80 years in Fennoscandia (Briffa *et al.*, 1990), but there was also little evidence for the MWP in this region. Although the MWP phase is found in many of the proxy records, it may not have been a globally synchronous event. As dendrochronology is able to date proxy records to an annual, and seasonal, resolution, further work has been able to show that over the Urals in Siberia summers were cool in the 11th and 12th centuries (Briffa *et al.*, 1995), times which have been shown to be warm in Fennoscandia. Regional disparities also occur over Britain; evidence since Barber (1981) has shown that a drier (or warmer) phase around the MWP occurred in Ireland between AD 1230-1380 (Blackford and Chambers, 1995) whereas in the Pennines a drier/warmer period is recognised between AD 1100-1250 (Tallis, 1997). Mauquoy (1997) identified a drier period between AD 1110-1290, which appeared similar to the record from the Pennines, within the boundaries of ^{14}C dating. However, in Dumfriesshire, Tipping (1995) identified a dry phase between 600-400 cal. BP (*ca.* AD 1350-1550), which occurred much later than in areas further south. It would therefore seem that before regional forcing mechanisms can be fully identified, more evidence is needed on the spatial and temporal variability of climatic events we know to have existed, but still do not yet have enough evidence to depict their (a)synchroneity.

Just being able to document and synthesise these changes is not always enough. With the increase in computational power over the past two decades, modelling climatic changes over the Holocene has become possible and more studies are being undertaken (see Kutzbach and Guetter, 1986; Kutzbach and Gallimore, 1988; and more recently Valdes and Glover, 1999). COHMAP (Cooperative Holocene Mapping Project) was set up in the 1970's in order to compare atmospheric GCM's with quantitative palaeoclimatic reconstructions (see Wright and Bartlein, 1993). The next phase of palaeoclimatic modelling is now PMIP (Palaeoclimatic Modelling Intercomparison Project), which has been set up to enable the comparison of atmospheric general circulation models (AGCM's) with palaeoecological reconstructions (e.g. Masson *et al.*, 1999). With more available palaeoecological data, and with dating and correlations of these proxy records becoming more precise, coupled with ever increasing computational power, it is hoped that in the future modelling will be undertaken which will give better insights into the forcing mechanisms of the climate throughout the Holocene.

2.7 Conclusions

The evidence cited above shows that climatic changes can be recognised from proxy evidence throughout northwest Europe. The evidence for climatic changes in Scotland fits in with many of the phase changes observed in Europe, although there are notable anomalies. Most evident is the observed wet shift in northern Britain at 1400 BP which has been recognised as a dry phase in Scotland. However, it has been stressed that there is a relatively small amount of peat stratigraphical palaeoecological data relating to climatic changes for Scotland during the Holocene compared to the rest of northwest Europe. Building on the established regional climatic shifts, the next step is to try and precisely date these climatic events, in order to build up a more detailed pattern of climatic change over Northwest Europe during the Holocene. The magnitude of these changes, coupled with the increasingly observed local nature of some climatic shifts, seems now to be of the utmost importance when trying to understand the climatic system, past and present.

This thesis aims to test and add to the palaeoclimatic data from Scotland using a multiproxy approach. The use of tephrochronology will enable the synchroneity of climatic changes to be tested across a range of sites in Scotland covering a large climatic gradient. The main themes identified from this review, and hence addressed by this project, can be listed as follows:

- It is clear from the above study that there is a need for more peat stratigraphical palaeoclimatic data from Scotland (*cf.* Lowe, 1993). Seven sites are analysed in detail in this thesis. The wealth of palaeoecological data attained from these sites should enable climatic changes to be inferred throughout the last *ca.* 5000-7000 years over a local and potentially more regional scale. There is still plenty of scope for a greater coverage of palaeoecological data in Scotland in order to compliment the work carried out in this thesis.

- This thesis aims to test whether the well known climatic deteriorations which occurred in northwest Europe during the mid to late-Holocene can be identified from peat bogs in Scotland. Some of the data in Table 2.1 suggests that these climatic changes in northwest Europe may have occurred at different times in Scotland, and therefore potential leads and lags (within the limits of the chronologies) can be tested.

- Scotland has been shown to be sensitive to shifts in the polar front, as documented by the severe LIA conditions. Although the LIA can also be detected from analysis of peat stratigraphy in northern England (Barber, 1981; Mauquoy and Barber, 1999a), the higher latitude of Scotland should allow more subtle changes in the polar front, and hence climate, to be detected. There may therefore be differences in the nature and timing of climatic changes between northern and southern sites within Scotland, as well an oceanic influence affecting western sites compared with eastern sites.

- There is a need for more precise dating and correlation of peat sequences, and in the absence of wiggle-matching within this study, tephra isochrones have been used. Tephra searches were carried out on every site studied, enabling the possibility of precise correlations between a suite of sites, and the climatic changes deduced from the palaeoenvironmental analyses.

- Further research into periodicities within peat stratigraphic records is a specific aim of this thesis, including the subsequent identification with a possible forcing mechanism. There is now a need for the magnitudes of these climatic changes to be ascertained, which is now possible with the testate amoebae transfer function (described in Chapter

6 and discussed further in Chapter 7), as this models the moisture response in a linear fashion. This offers scope for comparisons of the same event at different sites, and possibly providing quantitative data for the palaeoclimatic modelling community.

- It is also an aim of this thesis to appraise the development and biodiversity of the ombrotrophic bogs studied. Many changes in species, notably extinctions, are directly linked to climatic changes, and exploring these relationships further will aid future interpretations of peat-stratigraphic research.

Chapter 3. Methods

3.1 Introduction

The previous section has reviewed the nature and extent of past climatic changes throughout northwest Europe, as detected from the peat stratigraphic archive. The methods used within this thesis for determining past mire water table fluctuations are now discussed. They incorporate the analysis of plant macrofossils, colorimetric humification, and testate amoebae. In order to correlate precisely the palaeoclimatic reconstructions between sites tephra isochrones have been utilised, and the detection of these within the stratigraphy was fundamental in the site selection process. Tephra isochrones also aid the radiocarbon based chronology, acting as 'pinning points' within the chronology (Barber *et al.*, 1999).

3.2 Tephrochronological approach

3.2.1 Introduction

The detection of tephra layers within peat stratigraphy offers great potential in palaeoecological reconstructions, as the airfall events represent a discrete moment in time. If these tephra horizons can be found at multiple sites, the tephra layers can be correlated to develop tephra isochrones (Dugmore *et al.*, 1995a). The utilisation of the tephra isochrones as a dating method based on the identification and correlation of tephra layers was first termed 'tephrochronology' by Thorarinsson (1944).

Three types of tephra exist: pumice, sand and ash, of which ash is the most fine-grained, and is therefore the most easily transported by the atmosphere. It is these ash falls which commonly form the tephra layers identified in sediments located far from the source volcanoes. In northern Britain, peat deposits have been found to contain fine-grained Icelandic tephra of varying concentrations, which commonly occur as distinct millimetre-scale horizons (Dugmore *et al.*, 1995a).

Raised peat bogs are excellent retainers of tephra layers, as the inorganic content within them will be primarily atmospheric fallout, implying a continuous record of atmospheric sedimentation will be stored within the bogs. The first documented discovery of Holocene Icelandic tephra within mainland Britain was in blanket peat at Altnabreac, situated in the peatlands of Caithness, Scotland (Dugmore, 1989). Following on from Dugmore's pioneering work, discrete, well-defined layers of Holocene tephra have been discovered at many other sites in Britain (Bennett *et al.*, 1992 (Shetland); Pilcher and Hall, 1992 and Pilcher *et al.*, 1996 (Ireland); Dugmore *et al.*, 1995a and Dugmore *et al.*, 1996 (Scotland); Pilcher and Hall, 1996 (northern England)). Locating and dating tephra horizons in peats has led to environmental interpretations of Holocene landscapes (Hall *et al.*, 1993; Hall *et al.*, 1994a), and suggested that environmental change may not always be climatically forced (Blackford *et al.*, 1992; Hall *et al.*, 1994b; Edwards *et al.*, 1996; Hall *et al.*, 1996). It has been suggested that the *Pinus* decline in Scotland could partially have been attributed to volcanic acid loading from the Hekla-4 eruption (Blackford *et al.*, 1992). However, it appears that in Ireland there is no link between the *Pinus* decline and Hekla-4 eruption, with the *Pinus* decline occurring before the tephra deposition around 4000 BP (Hall *et al.*, 1994b; Dwyer and Mitchell, 1997).

Tephra shards can be subjected to geochemical analysis as a means of typing each horizon with a geochemical signature. The glass chemistry within a single eruption is usually either approximately homogeneous or displays a distinct trend, allowing correlation of tephra shards along an isochronous layer (Westgate and Gorton, 1981). Thus, a tephra isochrone from a single airfall event can be traced spatially over large areas, depending on local levels of deposition. Dugmore *et al.* (1992) were able to show that the geochemical signatures for tephra from the Hekla-4 eruption were replicatable in Iceland and Scotland, proving that long distance transport and long time burial in peat does not alter the geochemical properties of the shards. This allows the prospect of precise correlations between sites across geographical and climatological gradients, and hence precise correlations between peat derived palaeoclimatic records.

The use of tephras as a means for precise correlations between sites relies on certain principles being true. Hunt and Hill (1993) suggest that the main assumptions behind tephrochronology are that the ash is deposited instantaneously (in geological terms) and that each isochrone possesses a unique geochemical signature. It has been shown above that the latter of these assumptions is generally valid, with the geochemical trend in individual isochrones representing compositional changes during an eruption (Thorarinsson, 1950), with the content of silica decreasing through time during major eruptions (Thorarinsson, 1967). It is generally accepted that ash layers were deposited instantaneously, although, spatial variations in fallout can occur, and this often results in each tephra isochrone displaying a patchy distribution (Dugmore *et al.*, 1995a).

Perhaps the most important aspect of utilising tephra isochrones for palaeoecological reconstructions is the lack of mobility of glass shards through the peat profile. Dugmore and Newton (1992) have shown that peak influxes of tephra are frequently restricted to less than 1cm of peat stratigraphy, giving a dating accuracy for the base of the tephra to *ca.* 15 years. Following up this argument, tephra shards from the Hekla AD 1510 eruption were studied at Loch Portain, (North Uist, western Scotland) and were found to be concentrated between 12 -15cm from the top of the monolith (Dugmore *et al.*, 1996). Detailed sampling revealed two peaks in shard concentrations, and geochemical analysis revealed two significant populations of tephra. It was therefore suggested that the likely diffusion of an individual tephra throughout the peat profile would be on the millimetre rather than centimetre scale (Dugmore *et al.*, 1996). Supporting this, Hall *et al.*, (1994b) demonstrate how 74% of Hekla-4 tephra from Sluggan Bog is restricted to a peat horizon 5mm thick, and 90% restricted to an 8mm band.

The need for replicatable data in order to assess the impact of volcanic activity on mid-Holocene climates in Ireland has recently been documented by Caseldine *et al.* (1998), who used pollen and humification analyses to reconstruct environmental change. Three profiles taken from a raised mire at West Corlea show a general change to increased mire wetness at *ca.* 2300 cal. BC across the period of tephra deposition, which has also been noted by Dwyer and Mitchell (1997). However, comparisons between the three profiles do not suggest a major synchronous flooding event immediately following tephra deposition. It is suggested that mineral and tephra redeposition occurred due to pool formation.

A number of crucial points can be deduced from the study of Caseldine *et al.* (1998). Firstly, it seems imperative that data are replicatable with respect to tephra horizons, highlighting the need for detailed field stratigraphy when undertaking any palaeoecological analysis (see Barber *et al.* 1998). Secondly, a multiproxy approach is beneficial when trying to establish the synchroneity of signals, and will aid the interpretation of potential noise within a climatic signal. Finally, it is important to analyse tephra concentrations at centimetre, or even millimetre scales, in order to make detecting synchronous changes more reliable. Caseldine *et al.* (1998) only used contiguous 4cm samples for tephra analysis, which creates problems in Ireland as it appears that two tephra layers have been deposited around 4000 BP (Dwyer and Mitchell, 1997). If the peak concentration of tephra shards could have been restricted to only a centimetre or two, identifying possible separate volcanic events and correlating changes in humification could have been made easier.

3.2.2 Laboratory procedures

3.2.2.1 Ashing

The ashing technique of Pilcher and Hall (1992) was used to locate tephra horizons as it was the quickest known method for analysing long cores. Contiguous 5cm^3 samples of peat were ashed in nickel crucibles at 600^{0}C for four hours and then left to cool. The samples were washed in 10% HCl in order to remove any soluble inorganics, then washed in distilled water. Each sample was sieved through a 24μm mesh to remove any small fractions of detritus and washed again in distilled water. Material was then mounted on slides in Histomount synthetic mounting medium and analysed at x100 and x400 magnification. The shards were identified by their characteristic morphology which is highly vesicular or curved platelets of bubble walls, often with sharp edges, as well as isotropism under plane polarised light. A few problems with the identification and counting of shards were discovered. The ashing technique does not always destroy all the organic material (especially where the peat is rich in monocot or ericaceous remains). Thus some shards may have been obscured on certain slides, and therefore not counted. Certain samples still contained so much material after ashing that not all of it was placed on the slide, therefore possibly under-representing the total amount of shards present at a certain level. Generally, shards under 20μm were not counted as they were often too small to be identified with certainty.

3.2.2.2 Electron microprobe slide preparation

Before work could be undertaken on the electon microprobe the shards had to be isolated from the peat. The ashing technique cannot be used for this as the high temperatures used alter shard geochemistry, specifically that of the alkalis. An acid digestion technique (Persson, 1971; Dugmore *et al*, 1992) was therefore used to isolate the mainly silicious material from the peat. $3cm^3$ samples of peat were taken from the same depth as the centre of the original $5cm^3$ sample which contained the largest shard concentration for a given tephra layer. Once the acid digestion was complete, the samples were pipetted in solution onto an evaporating dish in order to concentrate the silicates.

Slide preparation for electron microprobe work is an important procedure, and it is suggested that considerable time and care is spent over the procedure (P. Hill, pers. comm.). Slides were first frosted using a 600 grit carborundum powder, and then cleaned in an ultrasonic bath. An epoxy resin (in this case araldite and a hardening resin at a ratio of 9:1) was placed on the frosted slide, and the dried silicious material (containing tephra shards) was dropped via a spatula onto the resin and mixed in thoroughly. The slide was then left to harden on a hot plate for approximately 5 hours. Ideally the hardened resin should be a thickness of between 200-300µm above the slide surface.

The following stage involved grinding down the samples to between 70-100µm above the slide surface, which exposes some of the tephra shards at the resin surface. It is generally recommended to use either 1000µm or 1200µm carborundum paper (P. Hill, pers. comm.). 1200µm paper was used on these samples as the resin was only *ca.* 100-150µm thick on most slides, so there was not a substantial amount of resin to grind down, increasing the chances of losing shards through plucking during the grinding process.

Once ground down to a suitable thickness the slides were washed in petroleum ether in an ultrasonic bath. The final stage of preparation was to polish the surfaces of the slides using a diamond paste. Flat polished surfaces are necessary to prevent the absorption of X-rays by scratches in the surface of the sample (Boygle, 1994). The first polish was carried out using a 6µm diamond paste which gave a good overall polish but still left a few visible grooves on some of the material. Thus a second polish was carried out using a

1μm diamond paste which erased the majority of grooves and left as smooth a polished surface as is necessary for probe analysis.

The final procedure of slide preparation was to give the slides a carbon coat in order to provide a conducting surface layer, and to enable a path for the probe current (Reed, 1993). The edges of the slides were covered in colloidal graphite, leaving a small amount connecting the sample with the edge of the slide, thus maximising contact between the slide and the samples. The slides were then ready to be used on the electron probe for microanalysis.

3.2.2.3 Microprobe operating procedure

The most common and effective tool to determine the geochemical composition of volcanic glass is the electron microprobe (Westgate and Gorton, 1981; Larsen, 1981). The advantages over other methods, such as X-ray flourescence (XRF), is that it can be used where material is limited. Electron probe microanalysis (EPMA) is grain discrete, and therefore allows parts of glass shards to be targetted where they have not undergone alteration, or contain vesicles or microphenocrysts (Dugmore, 1989; Hunt and Hill, 1993). Thus fresh faces of shards can be selected for analysis, reducing the potential of poor analysis totals.

The electron microprobe discharges a beam of electrons approximately 1-2μm in diameter towards a glass shard, which then discharges X-rays, which are detected by crystal spectrometers within the probe. The X-ray energy of the sample is unique to each element, and its intensity is proportional to the amount of that element (Hunt and Hill, 1993). The X-rays collected by the spectrometers therefore allow a geochemical signature of each volcanic shard to be derived. Once a population of shards from an isochrone have been analysed, the geochemical character of the tephra layer can then be determined and correlated with the geochemistry of a known eruption (e.g. Hekla-4) or a known tephra isochrone (e.g. Glen Garry).

The electron microprobe at Edinburgh University (Cambridge Instruments Microscan V) was used for EPMA on tephra shards from each site. A standard wavelength dispersive analytical technique was employed with a voltage of 20 kV, beam current of 15nA, and a beam diameter of 1-2μm. It is worth noting that the volume from which X-rays are generated (which produce the analyses) as a result of the beam interaction with the sample, is a pear-shaped/spherical volume some 3μm deep and across. This in reality makes it difficult to analyse anything smaller than 5μm, with 10μm being difficult, and anything over 20μm not generally being a problem.

Once the electron beam had been turned on it was left to stabilise on a copper block for at least 20 minutes. The beam position was then verified and the spectral drift was checked. Once the beam was stabilised the standards of the elements were checked. The background counts were checked at the start of each session and then counts for the major elements - potassium (K), sodium (Na), silicon (Si), titanium (Ti), magnesium (Mg), manganese (Mn), iron (Fe), calcium (Ca), and aluminium (Al) were checked.

The final stage of probe preparation is to check the andradite standard which is of known composition, with over 200 analyses having been undertaken on the probe (P. Hill, pers. comm.). If the standard falls within the known error range then the analyses of samples can proceed. The andradite standard was checked approximately every 12 analyses to ensure that the beam was still performing correctly.

A single analysis on a tephra shard consisted of counting nine major elements using two spectrometers, with a single count lasting 10 seconds. Every analysis performed was recorded by a computer, and due to the volatilisation of Na, this element was checked at the start and finish of each analysis. Thus any mobility during the overall analysis can be assessed. Corrections were made for counter dead time, atomic number effects, flourescence and absorption using a ZAF correction program based on Sweatman and Long (1969).

3.3 Site selection and coring strategy

As detailed in the previous section a tephrochronological approach was used as one of the main site selection criteria. Hence one of the pre-requisites for each site selected was that it should contain at least one tephra isochrone, but preferably two, before any palaeoecological analysis would be undertaken.

A large study area such as Scotland presents many potential sites to choose from. The selected sites should be sensitive to changing climates, or at least be representative of a specific climatic zone within the study area (see Chapter 4). Therefore, it was attempted to study a cross-section of Scottish sites, reflecting present day climatological gradients, from the northwest to the northeast, and also from the southern Borders to the more northern central areas. The very north of Scotland was avoided, as many of the mires are blanket peats, often having complex hydrological regimes which can make the detection of a climatic signal difficult (Charman, 1995).

The option of studying paired sites was dismissed, as Mauquoy and Barber (1999a) showed that similar climatic reconstructions can be ascertained from such an approach. Using single sites increased the potential of detecting regional climatic signals over Scotland, and allowed different climate zones to be compared, from the hyper-oceanic of the west coast to the more continental climates of the northeast Cairngorms and Grampian plains.

Locating undisturbed sites from these regions was essential for reconstructing a complete palaeoclimate record. Many of the raised bogs in Scotland have been damaged, either cut for commercial purposes, recently afforested, or been drained causing shrinkage of the peat. Many of the larger peatlands were initially rejected for study, as field visits and reports from Scottish Natural Heritage (SNH) suggested they were not in an undisturbed state.

Fourteen sites were visited, with twelve sites being initially selected for analysis. Extensive field visits, coupled with data provided in SNH reports and laboratory tephra analysis enabled seven sites with distinct tephra horizons to be selected for further research. All fourteen sites originally investigated are listed in Table 3.1.

3.3.1 Coring strategy

An initial core was taken for tephra analysis from each site using a Russian 50 x 5cm pattern corer to a depth of 5 metres where possible. The Hekla-4 isochrone was chosen as the cut-off point for climatic reconstructions as it has been well dated (2310 ± 20 cal. BC, Pilcher *et al.*, 1995) and most ombrotrophic mires in Scotland contain at least 4000 years of peat. Taking an initial core for tephra analysis alone enabled material to be conserved from the master cores, making sure there was sufficient peat left for palaeoecological and radiocarbon analyses.

If the site was accepted for inclusion in the project after preliminary tephra analysis, a master core was taken using a 30 x 9cm Russian sampler (Barber, 1984), with 5cm overlaps between cores. Monoliths (40 x 10 x 10cm tins) were taken from the surface peats at Shirgarton Moss and Longbridge Moss. Ben Gorm Moss displayed exposed sections of blanket peat which, the whole profile of which was sampled with monolith tins. The master cores were taken next to the initial 'tephra' core, from the same sensitive part of the mire, usually a *Sphagnum magellanicum/papillosum* lawn located on the crown of the mire. Profiles from near the mire edges are more likely to contain a shorter, possibly less complete palaeoclimatic record. Having sampled the mire, the cores were wrapped in carbon-stable plastic bags and sealed with tape. On return to the laboratory the cores were stored horizontally in a fridge at 4^0C.

Only one master core per site was obtained for palaeoecological analysis due to time constraints for laboratory work and the need to study a number of sites. This approach therefore raises questions about replicability of data, a basic tenet of science, which has been comprehensively addressed by Barber *et al.* (1998). They suggest that to achieve the maximum climatically forced variation reflected in the stratigraphy, detailed field sampling should ideally be first undertaken. The preliminary field sampling has been shown to be necessary as the geographical centre of a bog does not always contain the deepest, and hence most complete palaeo-record, as found at Walton Moss, Cumbria (Barber *et al.*, 1998). Field stratigraphy was undertaken at each site with a 100 x 5cm

Table 3.1 Sites investigated, based on a southwest to northeast gradient

Site	Grid Reference	Intact Area (hectares)	No. identified tephra isochrones	Site Accepted/ Rejected	Comments
Longbridge Moss	NY 053 694	125	2	Accepted	Original extensive Lochar mosses have been afforested with conifer plantations, although Longbridge Moss remains intact with good raised peat stratigraphy.
Carsegowan Moss	NX 428 589	76	1	Rejected	Very small amounts of tephra. Previously studied by Stoneman (1993) who noted that the flora is highly modified, probably due to burning from sparks produced by the nearby railway line.
Langlands Moss	NS 634 513	25	2	Accepted	One half of the mire was forested which has now been cleared. Some shallow drainage ditches on the surface, and modified flora consisting of high levels of Ericaceae, although surface appears intact. Two well defined tephras including a historical ash layer at around 30cm depth.
Dalmellington Moss	NS 466 064	51	0	Rejected	The site has been reported to be burnt each year by SNH, although high levels of *Sphagnum* are maintained on the mire surface. No tephra was found.
Temple Hill Moss	NT 113 614	43	2	Accepted	Mire with dry stratigraphy but no signs of drainage ditches and contains typical raised mire floral assemblages with abundant *Sphagnum*. Two tephras are present.
Red Moss of Balerno	NT 162 635	12	0	Rejected	The site has been heavily drained, with large drainage ditches present and a dry stratigraphy of dark, amorphous peat. No tephra was found.
Shirgarton Moss	NS 647 962	58	2	Accepted	Site appears intact with only small drains on the surface and abundant *Sphagnum* coverage. Two tephras are present.
Kilhorn Moss	NS 620 962	33	1	Rejected	Situated next to Shirgarton and also displays abundant *Sphagnum* assemblages, although less tephras are present so the site was rejected.
Mallachie Moss	NH 963 170	*ca.* 80	2	Accepted	Large area of raised peat with cuttings around the edge, although the centre of the mire appears undisturbed, with swathes of *S. magellanicum*. Two tephras have been located in the stratigraphy.
Blar-nam-Fiadh	NH 834 537	138	0	Rejected	Good raised peat stratigraphy, although no tephra was found.
Rora Moss	NK 040 520	165	N/A	Rejected	Site heavily cut for commercial purposes. Previously surveyed by Tipping (1996). No cores were taken for further analysis.
Turclossie Moss	NJ 884 574	63	N/A	Rejected	The site is described by SNH as an intermediate mire, and displays a dry stratigraphy. Cuttings have been taken from the mire edge. No cores were taken for further analysis.
Craigmaud Moss	NJ 882 586	17	2	Accepted	Although only a small site, Craigmaud appears intact. The stratigraphy is well humified, although some plant macrofossils could be seen in the field. Two tephras are present.
Ben Gorm Moss	NG 432 659	*ca.* 10	4	Accepted	Water shedding uncut blanket site with some peat hagging. Contains at least four tephras.

Russian pattern sampler, in order to locate areas which contain the most *Sphagnum cuspidatum* pool layers and greenish-yellow pool muds. Detailed field stratigraphy is presented in Chapter 6. A profile taken from this sensitive area should thus exhibit maximum climatic variation, as opposed to coring through a persistent hummock, which could record a 'complacent' climatic signal (*sensu* Lowe, 1993).

3.4 Plant macrofossils

3.4.1 Introduction

Plant macrofossils are regarded as those subfossil remains which can be recognised without the aid of high-powered microscopy (Moore and Bellamy, 1974; Birks and Birks, 1980). The analysis of plant macrofossils need not be just an accompaniment to microfossil data (Middeldorp, 1986) but it can also provide an independent approach to the reconstruction of environmental conditions (Lowe and Walker, 1997). The study of plant macrofossil remains is a key aspect of this thesis, and their interpretation is fundamental for understanding changes in climate from peat stratigraphy (Barber, 1981). Gorham and Janssens (1992, p.123) comment on the use of bryophytes in palaeoecology as follows:

> "It is evident from autoecological as well as stratigraphical results that bryophytes respond rapidly to changing minerotrophic and hydrological conditions in peatlands and are excellent tools for the palaeoenvironmental reconstruction of peatland development and succession."

The bulk of the identifiable material in autochthonous peats is dominated by bryophyte remains, and can usually be identified to species level (Janssens, 1983). In British ombrotrophic mires these macrofossil components are usually dominated by *Sphagnum* remains. *Sphagnum* grows apically from capitula while dying at the base, at a rate determined by the species. Over time the dead mosses are altered chemically and structurally during decomposition to form peat (Johnson *et al.*, 1990; Clymo, 1991).

The changing species composition of *Sphagnum* within a hummock-hollow gradient on a raised bog is recognised as a product of water tolerance (Andrus *et al.*, 1983; Vitt and

Slack, 1984). Their organisation along specific ecological gradients is related to cation-exchange capacity, desiccation tolerance, desiccation resistance, water-holding capacity, drying rates and photosynthetic response at differing water contents (Clymo and Hayward, 1982; Andrus, 1986). These various characteristics result in individual species occurring at specific heights above or below the water table (Ratcliffe and Walker, 1958). However, it has also been suggested that hummock species have broad distributions within the hummock-hollow gradient compared to narrower distributions of hollow species (Andrus *et al.*, 1983). This idea is supported by evidence from British raised mires as hollow species died when transplanted to a hummock, whereas a hummock species survived when transplanted to a hollow (Clymo, 1970). This therefore highlights the need for interpreting the whole species assemblages from fossil data, rather than relying on specific 'indicator' species.

The decay rate of peat is crucial for preserving plant macrofossils. Decay weakens the plant matter, and as more dead matter is added to the surface the structure collapses (Clymo, 1991). During this collapse the bulk density increases maybe 4-fold and the hydraulic conductivity decreases by several orders of magnitude (Clymo, 1992). Differing rates of decomposition within mire systems will lead to differing rates of preservation, which may ultimately alter the relative abundance of the original bog flora. Studies from bog hummocks with a 30-40cm thick acrotelm in Sweden suggest that *Sphagnum* may be exposed to aerobic conditions for over a hundred years and may lose up to 80% of its mass before being passed through to the more slowly decaying anaerobic catotelm (Malmer and Holm, 1984). This value is in general agreement with the data presented in the review by Clymo (1984). This suggests that peat forming in hollows should exhibit much lower carbon losses as the acrotelm is much thinner (*ca.* <10cm) and the residence time in this layer is much shorter before it is added to the catotelm.

However, it has been found that decomposition rates in hollows can be nearly twice that of hummocks, suggesting that the rate of peat accumulation should be higher in hummocks (Rochefort *et al.*, 1990; Johnson *et al.*, 1990). There are, however, suggestions that the differing rates of decay are species controlled. It has been noted that *S. acutifolium* and *S. cuspidatum* decomposed at about twice the rate of *S.*

papillosum (Clymo, 1965). *S. cuspidatum* also decays faster in hollows than *S. fuscum* does in hummocks, where conditions in hummocks are more favourable for decay (Johnson and Damman, 1991). The explanation for the comparative lower decomposition of *S. fuscum* on the hummocks was probably due to a combination of low pH, higher uronic acid content (differing chemical characteristics to other species), or drier peat (Rochefort *et al.*, 1990). It would appear that this tends to keep the mire in equilibrium, with the hummocks accumulating peat comparatively faster, and the hollows decomposing faster. As the hollows add more peat to the catotelm, so they slowly rise relative to the hummocks. The thickness of the hummock acrotelm therefore decreases, and it begins to add peat to the catotelm more rapidly (Clymo, 1991). Hollows and hummocks should therefore tend to persist, with their relative importance being affected by climate wetness (Walker and Walker, 1961; Barber, 1981).

Sphagnum species also lose their branches and leaves at different rates. Johnson *et al.* (1990) noted how *S. fuscum* plants remain comparatively well preserved compared to other hummock species which lose their branches, then leaves faster. *Sphagnum* section *Acutifolia* species also lose their leaves more easily than *Sphagnum* section *Sphagnum* species, suggesting a possible over representation in the fossil record (Barber, 1981).

It has been suggested that some of the differences in decay rates of *Sphagnum* species and the mire vascular plants may be related to chemical composition. An age-related increase in polymerized lipids, the components of suberin and cutin, in the cells walls of *S. fuscum* has been described by Karunen and Ekman (1982) and Kälviänen and Karunen (1984) which may be related to a decrease in decay rates. There could therefore be the possibility of predicting decay rates of each species, if the chemical breakdown methods are known.

Clearly then fossil mire bryophyte assemblages, like most other palaeoecological techniques, must be interpreted cautiously. The decomposition pattern of mire non-bryophytes is also complicated. The differing physical and chemical composition of the plant remains differ markedly among species, which would account for first year weight losses ranging from only 5-8% for *Calluna* wood up to 35-45% for *Narthecium* leaves (Heal *et al.*, 1978). This may explain why *Narthecium* remains are not detected in macrofossil remains from raised bogs. The most resistant mire plant remains as they are deposited in the peat profile are *Calluna* rootlets and *Eriophorum* roots (Heal *et al.*, 1978). This could possibly lead to an over-representation of these macrofossil remains when reconstructing past floral communities (Clymo, 1965), and this must be considered when interpreting macrofossil diagrams.

However, with these known caveats and careful interpretation it is still possible to successfully use plant macrofossil analysis to reconstruct past vegetative communities and hence changes in mire water table levels.

3.4.2 Laboratory procedures

The first form of quantitative macrofossil analysis was an Abundance Estimate (AE) method, based on the work of Walker and Walker (1961), and Barber (1981). The remains were classified on a 1-5 scale (where 1=rare, 2=occasional, 3=frequent, 4=common, 5=abundant). Ben Gorm Moss was analysed using this method as plant macrofossil remains were sparse.

This technique has since been modified to allow a more precise quantification of the ecological data contained within peat cores. The Quadrat and Leaf Count (QLC) method of macrofossil analysis was devised in the Palaeoecology Laboratory at the University of Southampton (PLUS) in the early 1980's by Dr Keith Barber and Dr Mike Clarke. A new technique was necessary in response to the lack of quantifiable methods for bryophyte fragments as opposed to other macrofossil data (Birks and Birks, 1980). Semi-quantitative methods have been developed by van Geel (1978), Janssens (1983) and Dupont (1986), but all are labour and time intensive. Haslam (1987) suggested that the lack of techniques reflected the view of the time that macrofossil data were only used as an accessory to palynological data for interpreting climatic changes. Plant macrofossil data, notably from ombrogenous bogs, now appear more firmly established (e.g. Barber *et al.*, 1994b; van Geel *et al.*, 1996; Chiverrell and Atherden, 1999; McGlone and Wilmshurst, 1999; Mauquoy and Barber, 1999a,b; Hughes *et al.*, in press).

Each site was sampled at an interval of 4cm, with the exception of Ben Gorm Moss which was sampled at 2cm intervals and Temple Hill Moss which was sampled at 8cm intervals between 3-5 metres. In order to avoid sample contamination with material which may have been smeared along the corer surface in the field, the top layer of peat from the core/monolith being sampled was removed with a scalpel. A volume of $4cm^3$ was then cut from the core using either a scalpel or sharp scissors if fibrous roots were present. The sample was then washed on a 125µm sieve with 5 litres of water in order to remove any fine organic matter and disaggregate the macrofossils. The material remaining on the sieve was then washed into a glass trough and 70cl distilled water added to create an even film of macrofossils. Analyses were undertaken using a Nikon SMZ-10 at x10 magnification. A 10x10 square grid eyepiece was used to calculate the abundance of identifiable *Sphagnum*, ericaceous rootlets and wood, monocotyledons, unidentified organic matter (UOM), non-*Sphagnum* bryophytes, and empty space. Where a component covered over half an individual square it was given a score of 1%.

These quadrat counts were carried out 15 times per sample, the trough being gently shaken each time so that the same assemblage of macrofossils could not knowingly be counted twice. The counts were then averaged on the assumptions of Haslam (1987) and Clarke (1988), who suggested these analyses represented an accurate estimation of the sample population, based on a Gaussian distribution.

Monocotyledon remains were identified to species level where possible. Material was mounted in Aquamount and identified at x 400 under the microscope whenever present. Identifications were based on epidermal cell structures from the drawings in Katz *et al.*, (1977), Clarke (1988), as well as the photographs in Grosse-Brauckmann (1972) and Mauquoy (1997), and a reference collection of type material. Epidermal material is not always preserved, and when the Cyperaceous remains could not be identified to species level they were recorded as monocotyledons undifferentiated (monocots. undiff.) and expressed as a percentage value.

Sphagnum leaves were quantified by counting between 100-150 leaves which had been mounted on a slide in Aquamount. In levels where *Sphagnum* was sparse, as many leaves as possible were counted up to a maximum of 100. Leaves were identified at x400 magnification using the drawings and keys in Smith (1980) and Daniels and Eddy (1990), as well as an extensive bryophyte reference collection. The leaf counts are expressed as a percentage of the total identifiable *Sphagnum* for that level. Non-*Sphagnum* bryophytes were also found at each site and recorded as percentages of the peat components.

In order to record the most diverse amount of material possible, some other remains were classified on the abundance estimate 1-5 scale of Walker and Walker (1961). This was generally done for ericaceous remains (which were classified Ericaceae on the quadrat counts) and included *Vaccinium oxycoccus* leaves, *Calluna vulgaris* and *Erica tetralix* leaves, as well as their flowers and seed boxes which were identified from an extensive reference collection. *Cenococcum*, *Sphagnum* spores, *Eriophorum vaginatum* spindles, and charcoal were also recorded on a 1-5 scale.

Seeds were recorded on a 1-5 scale when found within the peat matrix. Identification was undertaken using the type collections at Southampton and helped by the texts of Berggren (1969), Berggren (1982) and by R. Scaife (pers. comm.).

3.5 Humification

3.5.1 Introduction

The decomposition of peat involves many physical and chemical changes. The combined activities of faunal and microbial populations in mire ecosystems result in the decomposition (degree of humification) of the plant remains. Clymo (1984) defines humification as being used to describe the chemical changes affecting peat, although physical processes also affect the degree of decomposition of peat, and these are therefore also reviewed here. The influence of environmental factors, notably changes in climate, result in variations in the rate of decomposition in time and space (Heal *et al.*, 1978). Hence, the degree of peat humification can be linked to surface wetness at the time of peat accumulation (Aaby and Tauber, 1975), and therefore changes in humification can be used for interpreting past changes in climate.

Humification was first defined by von Post as "*the degree of disintegration of the organic substance, regardless of the way this disintegration has taken place, and of what substances resulted from it*" (von Post and Granlund, 1926). A ten point humification scale based on physical deterioration and colour was derived by von Post (1924) but this was not used in this project for determining changes in humification (with the exception of field stratigraphy) as it is by nature a classification. As quantifiable changes occur within the peat matrix as plant matter decomposes, measurements on a continuous numeric scale are more accurate for the detection of small scale, more subtle palaeobotanical changes (Blackford and Chambers, 1993).

Comprehensive studies on the properties of peat materials have been carried out by Levesque and Mathur (1979). They found that the fibre content of the peat was significant in correlations with physical properties of the peat, and concluded that measuring fibre content was a suitable method of determining peat decomposition. However, reconstructing mire surface wetness solely from plant fibres requires a great deal of caution as plants decay at different rates, imposing a species-related signal within the data. Some roots are especially resistant to decay, for example *Eriophorum vaginatum* fibres, and hence interpretation of any signal would be complicated (Mauquoy, 1997).

Clymo (1965) suggests that direct attack on fresh *Sphagnum* by animals may be rather unimportant as an explanation of weight loss. The combined direct and indirect effects of a population of *C. sphagnetorum* (an enchytraeid worm) on *Sphagnum* decomposition over a year was estimated at 10% (Standen, 1978). However, Coulson and Butterfield (1978) suggest that the amount of decomposition caused by animals varies markedly according to the plant species, as litter bag experiments show losses from *Calluna* to be 43%, whereas only 1% of *Eriophorum vaginatum* leaves were lost. The amount of decomposition caused by animals on *Sphagnum recurvum* and *Eriophorum vaginatum* is negligible however, which is important, as *Sphagnum* spp. and *Eriophorum* form a large amount of the preserved material found in many peats. Coulson and Butterfield (1978) surmise that the chemical composition of a plant species is of utmost importance in order to determine the rate of decomposition on a blanket bog.

Differential decay has also been observed between different *Sphagnum* species occurring in contrary microforms. Hogg (1993) measured CO_2 emission rates from hummock and hollow peats at different depths from a Swedish raised bog. It was found that the amount of exposure of *Sphagnum* peat to decay processes (i.e. residence time in the acrotelm) was more important than the species composition when determining rates of decay. This would suggest that there should be no species signal in humification data.

The method used in this project utilises the extraction of humic acids, which are produced by the decomposition of organic material. Humic substances are probably the most widely distributed natural products on the earth's surface (Schnitzer and Khan, 1972), although they have also been described as nature's least understood materials (Davies and Ghabbour, 1998). When plant matter decomposes a vast array of chemical transformations occur which are extremely complex and far from being fully understood (Blackford and Chambers, 1993). However, Hayes and Swift (1978) have distinguished between recognisable plant/animal debris and the highly transformed material which contain no recognisable plant, animal or microbial structures. All of the recognisable debris constitutes non-humic substances, whereas the highly transformed amorphous dark coloured materials are classified as humic substances. A classic and generalised definition of humic substances is given by Aiken *et al.*, (1985, p.1) stating that *humic substances* are *"a general category of naturally occurring heterogeneous organic substances that can be generally characterised as being yellow to black in colour, of high molecular weight and refractory"*. However, Hayes (1998) suggests that although broadly relevant, humic substances do not always have high molecular weight values, and they are only refractory when contained in protected environments. Humic substances can be broken down into the generalised terms of humic acids, fulvic acids, and humins which cover the main components, although the boundary between these fractions is not clear (Hayes, 1998).

The proportion of humic acids increases as peat decomposes. Measuring the degree of peat decomposition (humification) can therefore be ascertained if these acids can be extracted. A dark colour would be indicative of well humified peat, and suggest a drier mire surface, whereas a lighter colour would indicate peat which has been less humified, suggesting a wetter mire surface.

The most effective agent of extracting humic substances from soils or sediments most commonly involves dilute aqueous NaOH solution (Schnitzer and Khan, 1972). However, it has also been noted that other compounds are created by the extraction procedure (Hayes, 1985). Different solvents have been tested for extracting humic acids as efficiently as possible, and NaOH was shown to be the most reliable single reagent (Hayes *et al.*, 1975). Blackford and Chambers (1993) used alkali-extraction and fibre content methods to test the degree of peat humification and it was the colorimetric measurements which provided a robust and replicable record. The NaOH method appeared to be more suitable for reconstructing past climates from peat stratigraphy due to the differential response of changing species assemblages in the fibre content method. This NaOH approach has been used successfully in palaeoclimate reconstructions by a host of workers, notably Aaby (1976), Blackford (1990), Blackford and Chambers (1991, 1995), Nilssen and Vorren (1991), Chambers *et al.* (1997a), and Mauquoy and Barber (1999a,b). It is the alkali extraction method which has therefore been utilised in this project.

3.5.2 Colorimetric analysis of peat humification

The methods used followed Blackford (1990) and Blackford and Chambers (1993). Changes in humification are recorded as the absorption of light from an alkaline extract of peat is proportional to the amount of humic matter dissolved, with greater transmission of light through less humified material (Aaby and Tauber, 1975).

Subsamples of $4cm^3$ were taken from the peat cores and dried under an infra-red lamp overnight. The dried peat samples were then ground down with a pestle and mortar, and 200mg of peat weighted out. This sample was then brought to boiling point in 100ml of freshly mixed 9% NaOH and simmered gently for one hour. Blackford and Chambers (1993) suggest using 8% NaOH, but personal communication with F. Chambers revealed that 9% NaOH is now used. Blackford (1990) suggests that absorbance is only reduced, and hence extractant of humic acid lost when using strengths in excess of 10% NaOH. Haslam (1987) reported that boiling the samples leads to differential evaporation which was also noted in the present study. However, Blackford and Chambers (1993) quantified this effect by covering some samples whilst leaving others uncovered, and found no statistical differences between either method. After cooling the samples were transferred into a 200ml flask, topped up to the mark with distilled water, and shaken well. Samples were filtered through Whatman Qualitative 1 paper into a beaker. 50ml of the solution was then diluted 1:1 with distilled water into a 100ml flask and shaken well. Four hours after the initial mixing the samples were measured 3 times on a Spectrophotometer. The absorbance was recorded at 540 nm, and the average reading calculated. Percentage transmission was also recorded, but due to some errors in recording the data, these values have been eliminated from the study.

It has been suggested that inorganic particles such as clay could gather around humic molecules in such a way as to prevent their solubility in alkali (Schnitzer, 1967). This would reduce the amount of humic acids extracted by the alkali method, and distort the humification data (Chambers, 1984). Blackford (1990) tested this theory by deliberately contaminating some samples with clay, and was able to demonstrate a linear relationship between increasing mineral matter content and absorbance. This can then be corrected by determining loss on ignition values for each sample. These studies were conducted on blanket mires which can receive colluvial inwash from steeper slopes above. In this study however, only raised mires and a water shedding blanket mire were investigated. Since these mires receive inputs only from the atmosphere it was assumed that there would be negligible amounts of mineral content within the peat. This was also tested when large (5cm^3) samples were ashed for tephra content and only negligible amounts of inorganic material were retained.

3.5.3 Near infrared reflectance spectroscopy (NIRS)

NIRS has only recently been developed in palaeoecological studies of peat (McTiernan *et al.*, 1998) and so was not used in this project. However, it is a rapid, non-destructive technique, and therefore warrants inclusion in this section. McTiernan *et al.* (1998) obtained NIRS spectra for each sample used for palaeoecological analyses, and calibrated between the conventional analyses and the NIRS spectra using modified partial least squares and stepwise regression. The conventional palaeoecological techniques have been shown to correlate very well with NIRS analyses, although further work is necessary to identify whether calibrations can be developed across a number of cores from the same and different sampling locations (McTiernan *et al.*, 1998). It is suggested that NIRS should be attemped in future palaeoecological investigations.

3.6 Testate amoebae analysis

3.6.1 Introduction

Testate amoebae (Rhizopoda, Protozoa), also know as Testacea or thecamoebae, are unicellular animals with a discrete shell enclosing the cytoplasm and are the most common group of single celled organisms living in surface peat (Warner, 1990). Heal (1962) estimates an abundance of 1.6 x 10^7 individuals per m^2 from an investigation into Testacea found in *Sphagnum recurvum* in wet flushes at Valley Bog, Moor House National Nature Reserve. Peatland testate amoebae are important indicators of hydrology in fossil studies (Charman and Warner, 1997) and therefore warrant their inclusion in this study. Testate amoebae have been studied in peatlands since early this century, with some of the pioneering work being undertaken by Steinecke (1927), which was carried on by Harnisch (1927, 1951) and Grospietsch (1958), and later in Britain by Heal (1961, 1962, 1963, 1964). Their work suggested that moisture is the main ecological variable governing variations in distribution and abundance of Testacea outside aquatic habitats (Tolonen, 1986).

Testaceans can only live in an aquatic environment, but are able to withstand periods of desiccation by way of encysting. The ability to withstand desiccation may therefore be a factor separating pool species from hummock species, but Fantham and Porter (1945) have shown that some pool species are able to withstand desiccation for more than 6 years.

The qualitative work on reconstructing past water table fluctuations from fossil testaceans has more recently become quantitative (Woodland, 1996; Charman, 1997; Woodland *et al.*, 1998). Transfer functions are now being utilised by palaeoecologists to provide more quantitative data rather than the qualitative and semi-quantitative data which has previously been available. Transfer functions can be described as regression techniques in which the dependent variable is the environmental factor of interest, which is used to calibrate fossil species data so as to derive the estimates of the unknown environmental variable (Charman, 1997).

In developing transfer functions modern training sets must first be derived. Warner and Charman (1994) sampled modern testate assemblages from peatlands in Ontario and Minnesota as a guide to interpreting fossil assemblages. The modern data sets were subjected to canonical correspondence analysis (CCA) which suggested that moisture variables, most significantly depth to the water table, were strongly associated with Axis 1. Axis 2 appears most strongly correlated with pH. These relationships were noted to be similar to those described from peatlands in northeast Ontario (Charman and Warner, 1992) and Finland (Tolonen *et al.*, 1994). Reconstructions were able to be made on the fossil data spanning the whole of the Holocene, although Testacea were only recovered from peat post-dating 6500 BP.

It is important to note that the main environmental variable to which the testate amoebae respond is depth to the water table, so reconstructions of past mire water table depths can be based on these modern training data sets (Warner and Charman, 1994; Charman, 1997; Charman and Warner 1997; Woodland *et al.*, 1998). Woodland (1996) developed a modern training set for British Testacea, which was tested extensively with British testate fossil data by Hendon (1998). The results are comparable with other fossil data from around the world, specifically northern America and Canada (Charman and Warner, 1997), New Zealand (Charman, 1997), and Finland (Tolonen *et al.*, 1992; 1994).

The New Zealand work has showed that although screening the data (removing samples where predicted values and observed values differ greatly, after Birks *et al.*, 1990) may reduce prediction errors, this may provide a less satisfactory analogue for fossil samples (Charman, 1997). However, the lack of fossil data available has not yet allowed this potential problem to be addressed. The fossil data used in this study will enable further testing of the transfer function developed by Charman, Woodland, and Hendon at Plymouth University, and be amongst some of the first quantitative fossil data to be compared with other proxy-climate qualitative data (DCA and humification) of changes in mire water levels from ombrogenous bogs.

3.6.2 Testate identification

There has been a long debate about the taxonomy and identification of testate amoebae since the turn of the century. Books describing testates in English are now out of print and include *Freshwater Rhizopods of North America* by J.Leidy (1879) *and British Freshwater Rhizopods and Heliozoa*, by J.Cash, G.H.Wailes and J.Hopkinson (1905-1921: 5 volumes). Useful keys have since been produced, notably by Corbett (1973), Ogden and Hedley (1980), and Ellison and Ogden (1987). However, confusion still exists in the literature between the identification of species. There are two main schools of thought: clumping species together or separating species apart. Charman *et al.* (in press) has recently provided a comprehensive taxonomic key with illustrations which has been used for this thesis. Charman *et al.* aggregate some of the species together into types, which proves useful for identification purposes, but conversely some environmental information may be lost with this approach.

Testates were sampled from six of the seven sites selected for study, although species diversity was very low at Ben Gorm Moss. The reasoning was to test the transfer function against previously tested models of proxy-climate reconstruction, that of plant macrofossil data (DCA analysis) and humification (Chiverrall and Atherden, 1999; Mauquoy and Barber, 1999b; Charman *et al.*, 1999).

3.6.3 Methodology

The methods of preparing the thecamoebae follows closely that of Hendon and Charman (1997). Samples were taken at every 8cm except for the lower peats of Temple Hill Moss which was sampled every 16cm from 3-5 metres. Time constraints precluded a sampling interval at higher resolution. This was very much seen as the third ranked proxy indicator in importance after plant macrofossil and humification analyses, based on the conclusions of Mauquoy (1997). However, the transfer function and taxonomy of Charman *et al.* (in press) was not available in 1997, and as these are seen as major advances in the applicability of Testacea in palaeoclimatic reconstructions, results corroborating the plant macrofossil and humification data would be viewed as a major breakthrough.

A volume of $2cm^3$ was sampled and broken up in a beaker. Three *Lycopodium* tablets were added to enable concentrations of Testacea to be calculated (Stockmarr, 1971). The peat was boiled for 10 minutes in 100 ml of distilled water and then left to cool. The samples were sieved and the fraction between 250µm and 24µm retained. Hendon and Charman (1997) suggest using 300µm and 15µm sieves, but small species less than 24µm were still found in the samples. However, it is noted that a few of the smaller species may have been lost, but this was considered negligible as some of the smaller species were unidentifiable as they were obscured on the slide by the remaining detritus.

The material left was then washed, centrifuged, and stained with safranin. Another wash was followed by storing the remains in glycerol in a glass vial. A small amount of material was mounted on a slide and two drops of water added, as this improves the definition of the Testacea under the microscope (Charman, pers. comm). The addition of water, however, means that the slides must be analysed within a few hours or they will dry out. The slides are then completely sealed with nail varnish. At least 100 tests were counted per slide where possible. Levels were rejected when over 1000 *Lycopodia* were counted before 100 tests had been counted (Charman, pers. comm.). *Pinus sylvestris* pollen was also counted as concentrations of pine could be calculated, suggesting a crude date for the start of the 'pine rise'. Clymo *et al.* (1990) date a rise in pine pollen to 1835 AD by wiggle matching from Ellergower Moss, Dumfriesshire, and Tipping (1995) attained a date of 175 ± 25 cal. BP for an increase in pine at Burnfoothill Moss, Solway. Barber (1978) and Dumayne (1992) conducted extensive literature searches for a 'pine rise' date in Cumbria (also discussed in Dumayne-Peaty and Barber, 1998), and found it to occur at *ca.*1800 AD.

3.7 Spheroidal carbonaceous particle (SCP) analysis

High temperature fossil fuel combustion produces fly-ash particles which can be divided into two groups: spheroidal carbonaceous particles (SCPs) which are composed mainly of elemental carbon (Goldberg, 1985) and are formed from the incomplete combustion of the fuel, and inorganic ash spheres formed by the fusing of mineral inclusions (Rose *et al.*, 1995). The UK commonly uses three fossil fuels, but only coal and oil produce

SCPs. Peat combustion particles have an amorphous appearance, many still retaining a cellular structure (Rose, 1990).

In order to use SCPs as a dating tool, consistent particle trends must be present in cores which have been dated using a reliable technique, such as ^{210}Pb or varve counting. This trend can then be related to undated cores to provide a chronology for the uppermost sediments. The methodology was refined using lake cores, the recent sediments of which are easier to date with ^{210}Pb than peat bogs (Oldfield *et al.*, 1995). The SCP record in the UK has been divided into 3 characteristic trends (Rose *et al.*, 1995): the start of the record (A), the rapid increase in SCP concentrations (B), and a sub-surface peak in concentration (C). More importantly, Rose *et al.* (1995) analysed SCPs in lake cores from different regions within the UK and Ireland, the dates of which were found to differ slightly, as shown in Table 3.2.

Table 3.2 ^{210}Pb dates (means) for three SCP dating features (after Rose *et al.*, 1995).

Site	**Start of record (A)**	**Rapid increase (B)**	**Sub-surface peak (C)**
North Scotland	1850s	1960s	1976 ± 2
S.Scotland/N.England	1850s	1950s	1977 ± 2
Ireland	1880s	1960s	1981 ± 2
Wales	1850s	1950s	1978 ± 2
Southern England	pre-1920s	1950s	1969 ± 2

SCP studies in peats have also shown the three characteristic trends A, B, and C to be recognisable (e.g. Punning and Alliksaar, 1997; Mauquoy, 1997). There is evidence to show that fly-ash particles do not migrate through *Sphagnum* layers. Laboratory experiments show that only 0.8% of particles which were placed on a surface layer of 6-10cm thick *Sphagnum* were washed out with water (700-750mm) during the 241 days of the experiment (Punning and Alliksaar, 1997).

3.7.1 Methodology

As comparatively few SCP studies have been carried out on peats compared to lakes, no set methodology for peats exists. The method used here follows that for lakes (Rose, 1990; Rose, 1994), with slight modifications (N. Rose, pers. comm.; D. Mauquoy, pers. comm.).

0.1g samples of dried peat were weighed out into a polypropylene tube. 3ml of concentrated HNO_3 were added and left overnight in a fume cupboard. A further 3ml of concentrated HNO_3 were then added and the tubes placed in a boiling water bath for 2 hours. The tubes were then left to cool, and centrifuged at 1500rpm for 5 minutes, followed by decanting, and repeating the HNO_3 digestion in the water bath as necessary. Once the organics were removed, the tubes were topped up with distilled water, centrifuged and decanted. This was then repeated. Finally, the residue at the bottom of the test tube was decanted into a glass vial of known weight. The vial was then reweighed and centrifuged. The remaining water was then pipetted off, and the vial reweighed once more. The weight of the residue is now known. A small proportion of the residue was then evaporated onto a cover slip, which was then mounted onto a slide in 'Naphrax' diatom mountant. The whole of each coverslip was counted at x400 and concentrations recorded in grams per dry weight (g DW^{-1}, after Rose, 1990).

As well as providing reliable dates for the most recent peats, SCP analysis can also give indications of whether a bog has been cut or damaged. The upper two radiocarbon dates from Shirgarton Moss suggested ambiguous accumulation rates for the upper metre of the profile (see section 5.3.4), and raised doubts as to whether the bog was intact or not. No unambigouous pine rise was detected at this site. SCPs were therefore counted at this site. Time precluded additional analyses at other sites.

3.8 Radiocarbon dating

A reliable geochronology is essential in order to establish the synchroneity of changes in climate as well as to make precise comparisons with palaeoclimatic data from other sites, whether at the local (inter site), regional, or global scale. Radiocarbon dating is the most frequently used method to establish the geochronology of peat sequences, and that is also the case in this study.

The physics of radiocarbon dating relies upon the fact that carbon 14 atoms (^{14}C) decay radioactively over time with a half life of 5730 years (Stuiver and Polach, 1977). The activity can be measured for a given sample, usually by gas or scintillation counting, or by mass spectrometry for AMS dates. Once the activity is known, the proportion of ^{14}C

per sample can be calculated, and the age of the material can be determined. This is the age in *radiocarbon years*, and is calculated from AD 1950 which equates to 0 BP, defining a conventional ^{14}C age. The proportion of ^{14}C in the atmosphere has changed significantly since the 1950's due to nuclear bomb tests, and hence this date is used as a baseline for radiocarbon years.

If reconstructions from past levels of ^{14}C concentrations are to be made, based on present levels of ^{14}C, then it is a fundamental assumption that natural concentrations of ^{14}C in a given carbon reservoir of zero ^{14}C age remain constant over the entire ^{14}C time scale (Taylor *et al.*, 1996). However, pioneering work in the late 1950's by Hessel de Vries (de Vries, 1958; 1959) showed that there have been changes in the production of exchange parameters governing atmospheric ^{14}C levels. These secular variations in atmospheric ^{14}C concentrations over time have become known as 'de Vries effects' or ^{14}C 'wiggles' and constitute a need to calibrate the radiocarbon date to either calibrated years BP or calibrated years BC or AD. This project will use cal. BC/AD. The wiggles are thought to correspond directly to variations in solar output, and therefore may be a proxy for changes in solar variation (van Geel *et al.*, 1996).

The calibration data have been based on dendrochronological records from Irish and German oaks, as well as Douglas fir, sequoia and bristlecone pine which can now document about 9800 years of dendrochronological time (Stuiver and Pearson, 1993; Pearson and Stuiver, 1993; Pearson *et al.*, 1993; Pearson and Qua, 1993). The record has now been provisionally extended further back in time, including a 'floating' tree ring record to the Lateglacial and calculated as far back as 21,950 cal. years BP from U-series and ^{14}C dating of corals (Taylor *et al.*, 1996).

The short term fluctuations in the radiocarbon calibration curve can, in some instances, form the basis for very accurate and/or precise dating (Pearson, 1986). The dating of a series of samples spanning a very short time interval, whose dendrochronological ages are known, can be utilised for wiggle-matching (van Geel and Mook, 1989). High-precision dates obtained from the samples form a time-depth plot which is matched against the dendrochronologically based calibration curve (Lowe and Walker, 1997). The dated section may, in some cases, be matched to within a few years on the calendar timescale (Pilcher, 1991). Wood is the ideal material to use for wiggle-matching, but Clymo *et al.* (1990) used seven dates on peat samples with good results. Wiggle-match dating in peats have also been used to show how the sharp rise in atmospheric ^{14}C between 850 and 760 cal. BC can be correlated with synchronous climatic change in northwest Europe (van Geel *et al.*, 1998).

It would therefore appear that wiggle-matching is the best technique to use when dating peats, when high precision dates are required to underpin climate interpretations and regional or global correlations (Pilcher, 1991). However, this approach is expensive. To wiggle-match one climatic event in the peat-based palaeoclimatic records requires at least five dates. Therefore to date six major changes in a three metre core would require 30 dates, and such an approach employed at the seven sites used in the project would therefore require around 210 dates, a number which was judged to be beyond the resources that could reasonably be expected from the NERC Radiocarbon Facility Steering Committee.

Peat, however, is not necessarily an easy material to radiocarbon date. Shore *et al.* (1995) have demonstrated how discrete chemically defined fractions within the peat can produce significant variations in radiocarbon content. The humic, fulvic, and humin acids can all display differing radiocarbon activities, even within *Sphagnum*-dominated peats. Pre-treatments for radiocarbon content which retain the acid insoluble humic and humin acids will therefore potentially record erroneous dates. Possible causes for the age differences are suggested by Shore *et al.* (1995) and include the upward movement of humic acid or the intrusion of younger rootlets from above.

Bulk radiocarbon dates have therefore been used in this project, and although Pilcher (1991) stresses that care must be taken when using and interpreting such dates, the accumulation rate of peat bogs has been found to be relatively constant over time (Clymo, 1991; Barber *et al.*, 1994b). Clymo (1991, p.100) comments on the peat accumulation rate as being a "*stately march through the millennia consistent with constant influx and constant proportional decay*" during the Holocene, suggesting that for the majority of mire systems a linear plot can be drawn through the accumulation rate, aiding the interpretation and error margins of the radiocarbon dates (*cf.* Mauquoy and Barber, 1999a).

Problems with radiocarbon dating have also been discovered by Kilian *et al.* (1995). They used AMS wiggle match dating on some previously dated peats from Enkhuizen-dijk, and found them to be 100-250 years younger than the conventionally dated peats, suggesting that a reservoir effect influenced the conventional age estimates. The original bulk samples contained 2-4% very fine Ericaceae rootlets, and it is possible that the problem could be related to the mycorrhizal fungi associated with *Calluna* roots. Methane is produced by bacteria in the permanently anaerobic peat, and slowly diffuses to the surface. Methane consuming bacteria nearer the surface oxidise the gas forming CO_2. This therefore has the effect of producing older CO_2 in the root zone of Ericaceae. It is suggested that this could be the exclusive carbon source for specialised fungi living in the rootlets, which are still present during ^{14}C analysis, thereby making the sample anomalously older. Further AMS ^{14}C dating of fungal remains will help tests this hypothesis.

These *caveats* notwithstanding, radiocarbon dating has been undertaken on 37 samples from the seven sites at the NERC Radiocarbon Laboratory at East Kilbride. An advantage within this project is the utilisation of tephra 'pinning points' (see Section 5) to aid chronological control as well as being able to correlate precisely the proxy climate records, something which is difficult with radiocarbon dating. However, as the dates of some of the tephra isochrones are well known (Hekla-4 has been wiggle-match dated (Pilcher *et al.*, 1995)) they can be used to test the accuracy of the ^{14}C dates (e.g. Barber *et al.*, 1999).

Methods for the calibration of ^{14}C dates used OXCAL (Bronk Ramsey, 1995) which utilises the calibration curves of Stuiver and Pearson (1986) and Pearson and Stuiver (1986) which are published in Stuiver and Kra (1986). Following the recommendations of Van der Plicht and McCormac (1995) the 1986 calibration curve was used in preference to the 1993 calibration curve (Stuiver *et al.*, 1993) as dates from the latter may be out by *ca.* 15 calendar years.

3.9 Conclusions

The methods closely follow and add to those used by Mauquoy (1997), and encompass his suggestion of utilising tephra isochrones (*cf.* Blackford, 1997; Chambers *et al.*, 1997a). The use of a transfer function with the testate amoebae data allows quantitative data for the fluctuating mire water tables to be constructed for the first time, which was unavailable to Mauquoy (1997). Other microfossils and macrofossils have been used for investigating mire water table reconstructions, including detailed studies such as those of Casparie (1972) and van Geel (1978). However, as Mauquoy (1997) comments, these investigations are very time consuming, and their uses not yet fully understood. They are therefore not included in this study. The techniques which have been used are considered capable tools with which to reconstruct climatic changes.

Chapter 4: Site descriptions

4.1 Introduction

It is estimated that from an original area of 27,000 hectares of raised bog within Scotland only 2,300 hectares (9%) of raised bog in a 'near-natural' state remains (Stoneman, 1997). These statistics illustrate how rare this resource has become, and lead to inherent difficulties with locating suitable sites for palaeoenvironmental analysis. All the sites studied in this thesis are not known to have been studied in detail previously, providing the opportunity for original palaeoecological investigations.

The study sites are all located in Scotland and are spatially well dispersed, the aim of which was to provide sites representative of specific climatic zones within the study area (Figure 4.1). The twelve sites shown in Figure 4.1 include all the sites initially investigated for tephra analysis (sections 3.3 and 5.2.1). The mires are dispersed over climatological gradients from the generally wetter west to the comparatively drier east of present day Scottish climates, as well as including a north-south distribution. The range and variation in site locations should reflect regional variability of climatic changes, as well as site sensitivity.

Generalised variations in the present climate of Scotland is summarised in Figure 4.1 via the bioclimatic sub-regions identified by Birse (1971). The three main zones are differentiated with respect to degrees of oceanicity, as this was thought to be the most useful factor in determining the potential forcing mechanisms with respect to oceanic-atmospheric interactions. Birse considered Scotland to fall wholly within the oceanic sector of Eurasia, and defined sub-sectors on thermal criteria alone, the basis for these being accumulated day degrees of frost at lower altitudes. Three categories were selected empirically, and drawn up by Birse as shown in Figure 4.1. Birse identified the problems of zoning bioclimatic features, and accepted that some sacrifice in precision was inevitable when equating the many climatic characteristics, but the zonations provide a first order impression of the bioclimatic variability within Scotland. The three sub-sectors distinguished by Birse are as follows:

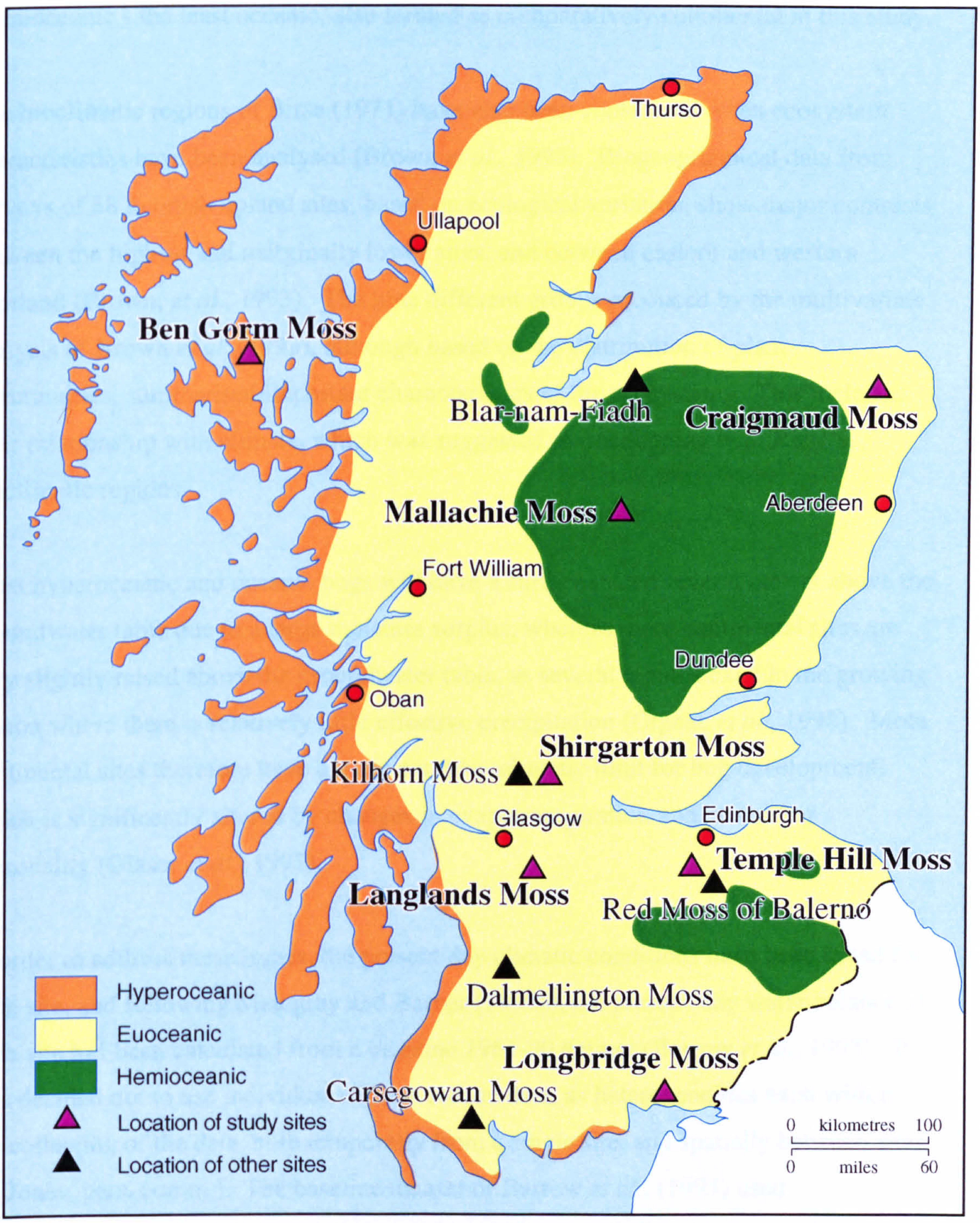


Figure 4.1: Location of sites and climatic sub-regions (adapted from Birse, 1971).

Hyperoceanic - extremely oceanic
Euoceanic - an intermediate term
Hemioceanic - the least oceanic, also termed as comparatively continental in this study.

The bioclimatic regions of Birse (1971) have also been considered when ecosystem characteristics have been analysed (Brown *et al.*, 1993). Biogeographical data from surveys of 88 Scottish upland sites, based on ecological variation, show major contrasts between the highest and marginally lower sites, and between eastern and western Scotland (Brown, *et al.*, 1993). The nine different groups produced by the multivariate analysis of Brown *et al.* (1993), although based on the distribution of plant communities, summarised important characteristics of the ecosystems. This included their relationship with climate, which was suggested as overlapping with Birse's bioclimatic regions.

Most hyperoceanic and oceanic bogs will have a surface raised several meters above the groundwater table due to a large moisture surplus, whereas more continental sites are only slightly raised above the groundwater table, as several months exist in the growing season where there is relatively little effective precipitation (Gignac *et al.*, 1998). More continental sites therefore have a more complex climatic limit for bog development, which is significantly altered by changes in evapotranspiration and increasing seasonality (Glaser *et al.*, 1997).

In order to address these issues, the present day climatic conditions have been noted for each site, and following Mauquoy and Barber (1999a), the present-day water balance of each site has been calculated from a baseline 1961-90 dataset (Barrow *et al.*, 1993). It was decided not to use individual climate station data, as heterogeneities exist within the collecting of the data, both temporally from a single site, and spatially between sites (P. Jones, pers. comm.). The baseline dataset of Barrow *et al.*, (1993) used homogenised data, thereby overcoming these problems, and allowing comparisons between sites. The dataset is comprised of a gridded 10 x 10km coverage of average climate measurements (minimum, maximum and mean temperatures, precipitation, rain days, frost days, relative humidity, wind speeds, and vapour pressure) for low and high relief within the individual grid squares. These measurements have been used to calculate the potential evaporation with the help of a program written by N. Arnell

Figure 4.2 Annual precipitation for all sites

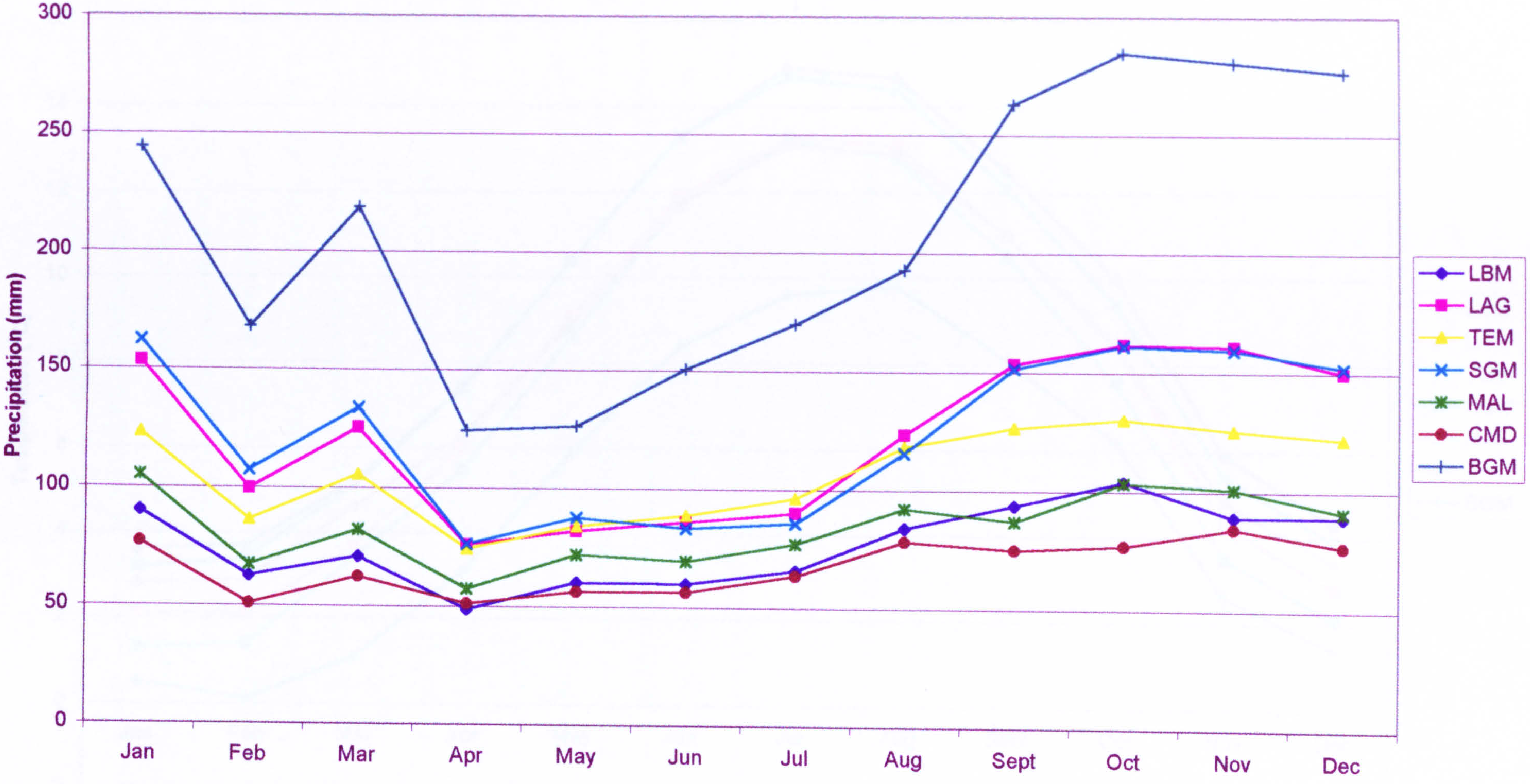
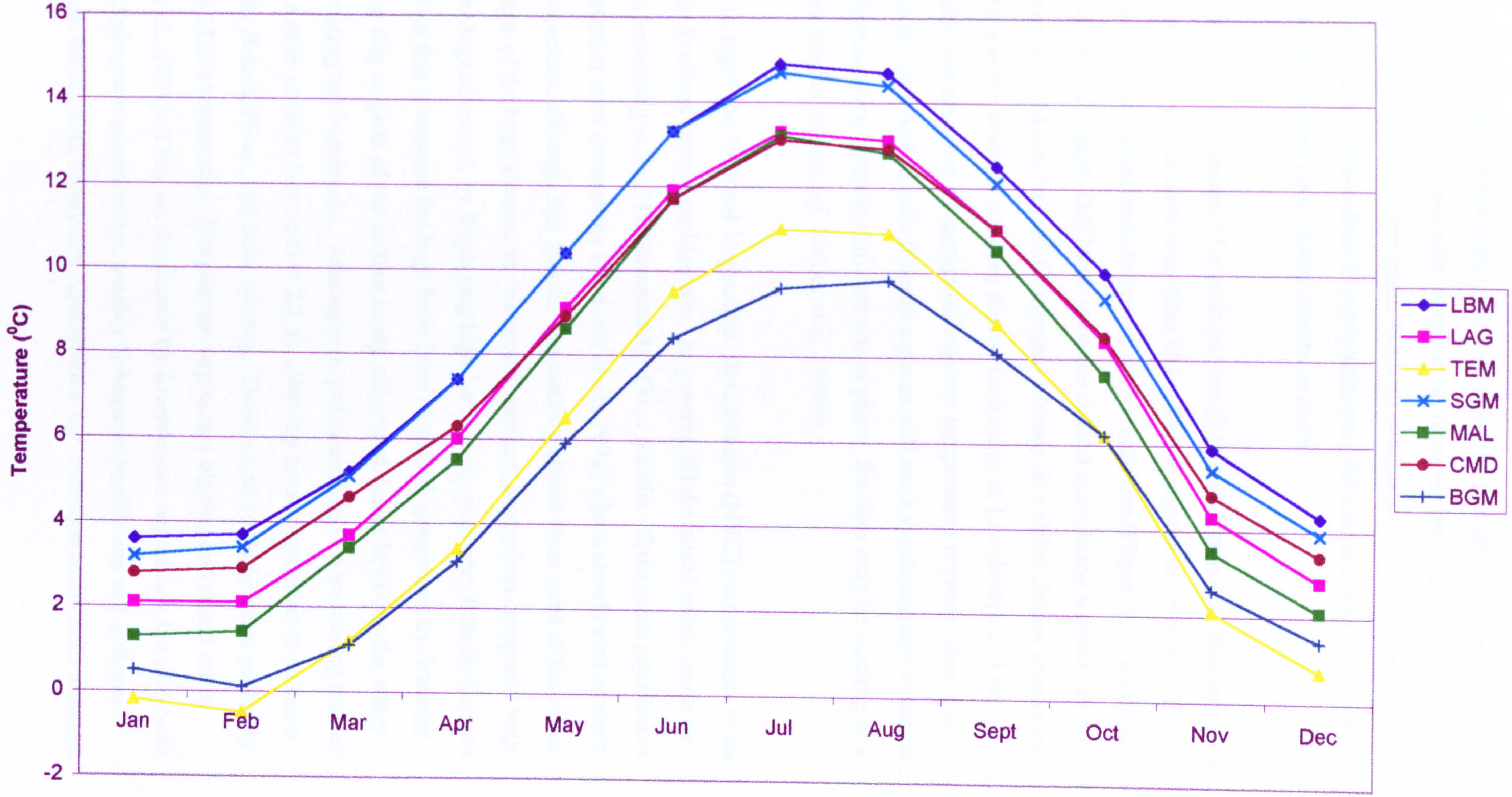


Figure 4.3 Mean annual temperatures for all sites



(Geography, Southampton University). Annual effective precipitation was then calculated by subtracting the potential evaporation from the precipitation values. These present-day climatic conditions are crude, but help comparisons between sites and regions, thus aiding the palaeoclimatic interpretations. All climatic data used in this thesis relate to the gridded data, unless otherwise stated.

The annual precipitation amounts for each site are shown in Figure 4.2, illustrating the difference between hyperoceanic sites (Ben Gorm Moss) and the other mires. This can be compared with the annual mean temperatures for each site (Figure 4.3), which shows Ben Gorm Moss and Temple Hill Moss to be the coldest sites, being a direct function of altitude. Langlands, Mallachie, and Craigmaud Mosses all exhibit similar temperature regimes, whereas the warmest sites are the lowland mires at Longbridge and Shirgarton. If climatic changes are more a function of coherent temperature regimes, then it is possible that the sites with similar thermal regimes will react synchronously to climatic changes, whereas when climatic shifts are out of phase, the sites may be reacting to a non-coherent rainfall regime (*cf.* Barber *et al.*, 1999).

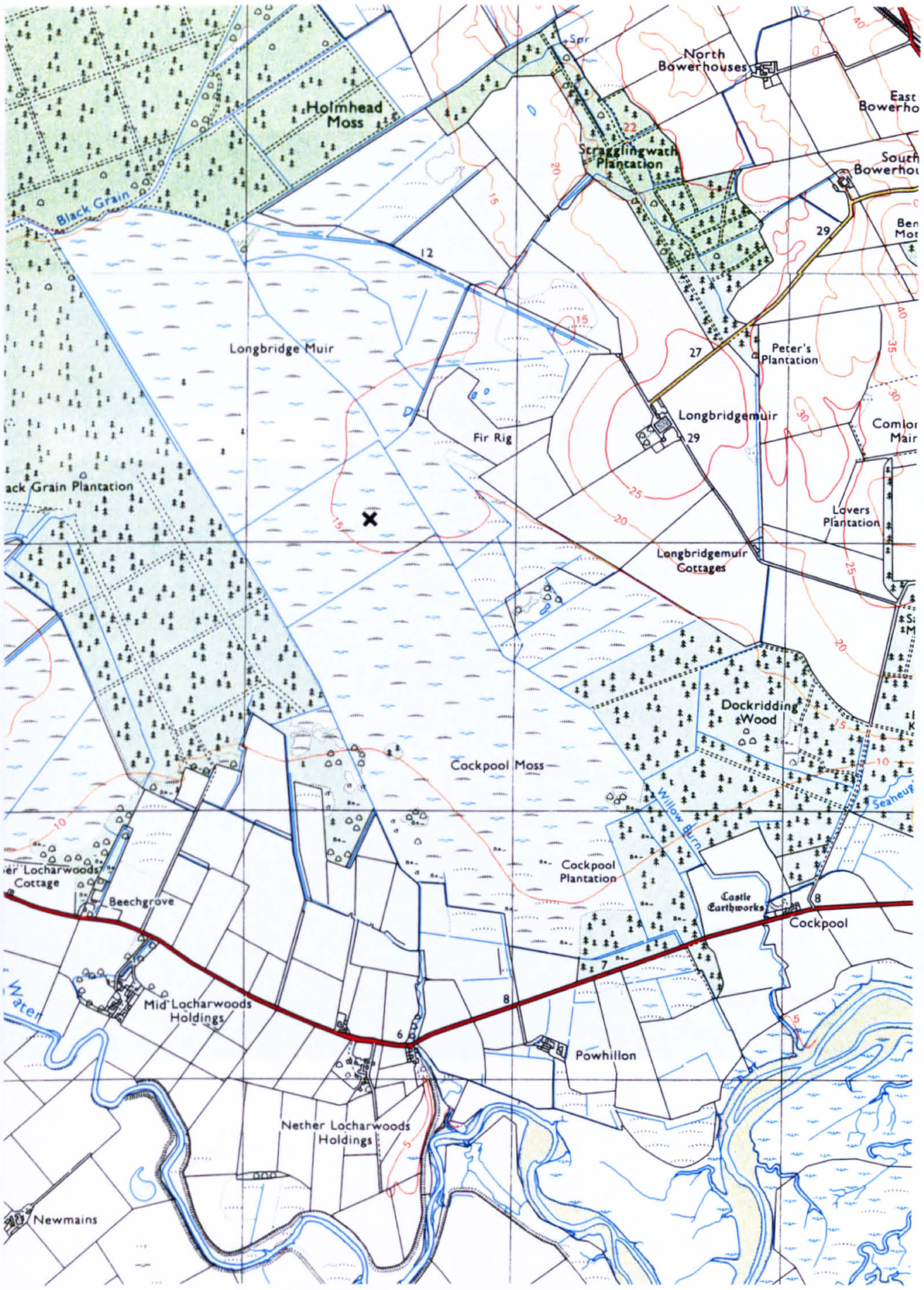
An attempt to apply the National Vegetation Classification (NVC) communities to the study sites in Scotland proved problematic. In general, all the raised mires *could* be classified as belonging to the M18 community: *Erica tetralix - Sphagnum papillosum* raised and blanket mire community (Rodwell *et al.*, 1991) which should also contain *Sphagnum tenellum*, although this species was notably absent from most of the mires studied. Some of the bogs showed an 'unnatural' surface, anomalous compared with the palaeoecological record, by displaying high hummocks with very little *Sphagnum* present. Often this is because the bogs have been disturbed, sometimes by drainage ditches being dug or parts of the surface being afforested, hence lowering the water table and creating tall hummocks. Atmospheric pollutants are also increasing nutrient status on the mire surfaces (see section 2.2.3), allowing scrub vegetation to colonise bogs, notably *Betula*, *Pinus*, and *Salix* species. These mires could be more generally classified as M20 communities: *Eriophorum vaginatum* blanket and raised mires (Rodwell *et al.*, 1991). They are dominated by *Eriophorum vaginatum*, but also have pockets of *Sphagnum* communities, notably *Sphagnum papillosum* and *Sphagnum magellanicum* lawns, with *Sphagnum capillifolium* var. *rubellum* hummocks amongst the *Eriophorum* and *Calluna*. Thus, for the purposes of this research NVC classifications have not been specifically assigned to the sites.

4.2 Longbridge Moss (LBM) - NY 3053 5694

Longbridge Moss is a SSSI, situated about 6 miles southeast of Dumfries within Forestry Commission land. It is one of only two remaining sites in the Nithsdale District and is the remnant of the once extensive Lochar Moss which embodied a complex of raised bogs. Afforestation with Lodgepole pine and Sitka spruce has now covered most of the original moss complex, but Longbridge remains unforested, and displays an extensive pattern of hummocks and hollows. Its location can be seen in Figure 4.4 and aerial photograph (Plate 4.1 – poor quality is due to a poor original supplied by Solway Heritage). The Lochar Moss complex was originally studied (albeit briefly due to time constraints) by Lewis (1905), who investigated boreholes on Racks Moss. The peat was found to be 15 feet deep, with a well marked basal layer of birch, overlain by *Scirpus*, *Sphagnum* and *Calluna* peat. Lewis noted that *Calluna* dominance occurred towards the top of the stratigraphy, indicating the possibility of lower water tables at these times. Lewis described the surface of the 'wetter' mosses in the region to be dominated by *Sphagnum*, with *Erica tetralix*, *Myrica gale* and *Eriophorum vaginatum*, with only little *Calluna*.

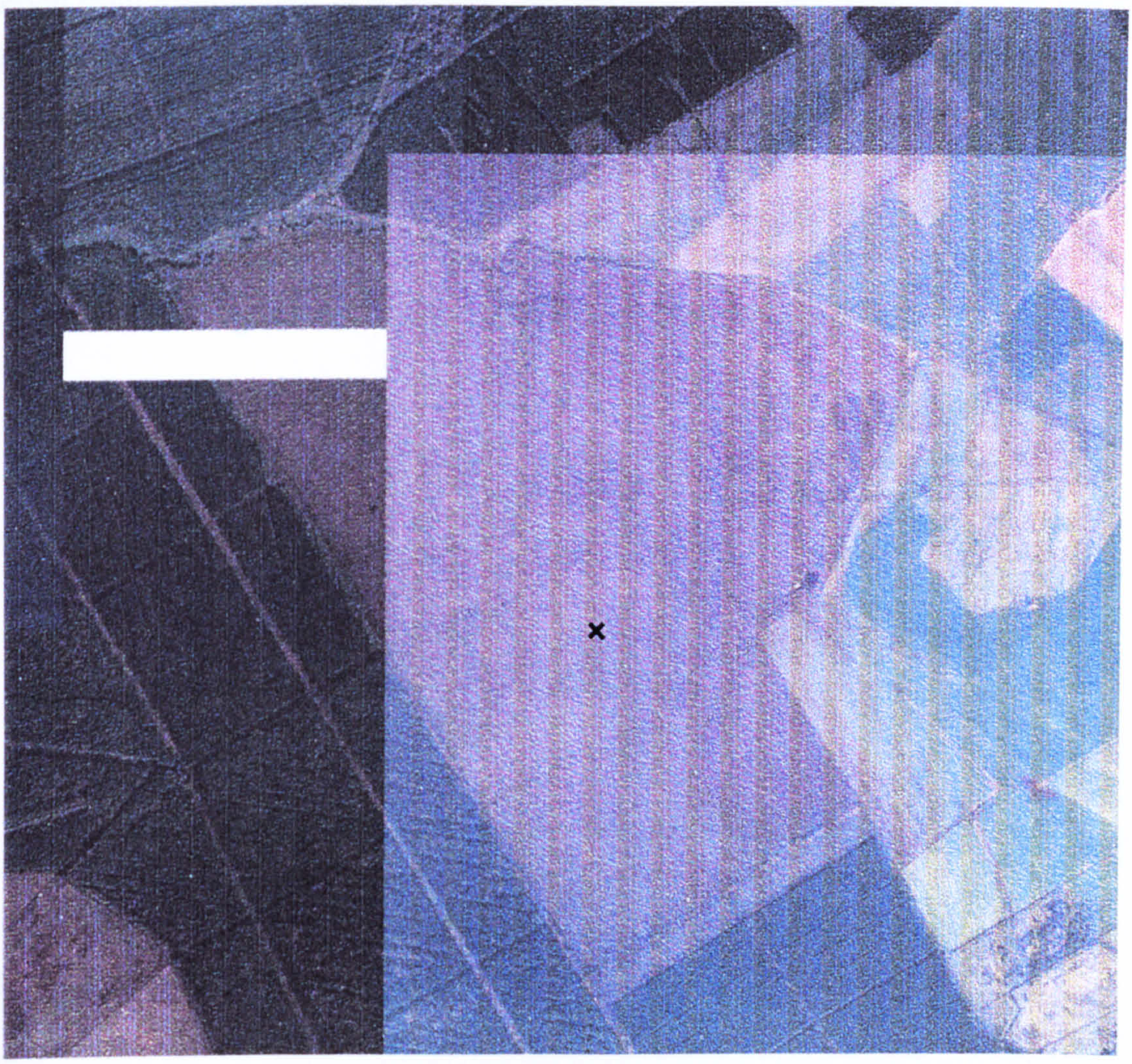
At present, Longbridge Moss covers 513 hectares, and much of the surface is dominated by *Calluna vulgaris*, *Erica tetralix*, and *Eriophorum vaginatum*, with reasonably large areas of *Sphagnum magellanicum* and *Sphagnum papillosum* lawn. *Andromeda polifolia, Narthecium ossifragum, Sphagnum cuspidatum, Sphagnum capillifolium* var. *rubellum* and *Sphagnum capillifolium* var. *capillifolium* are also locally common, and there is a large amount of *Cladonia impexa* present on the surface. Some small hummocks of *Leucobryum glaucum* can also be found. A few scattered small pine trees can be seen on the surface, usually near the margins. A three metre borehole was taken from this site for palaeoecological analysis.

Figure 4.4 Map of Longbridge Moss.



× Core location. Source: Ordnance Survey (1:25,000).

Plate 4.1 Aerial photograph of Longbridge Moss.



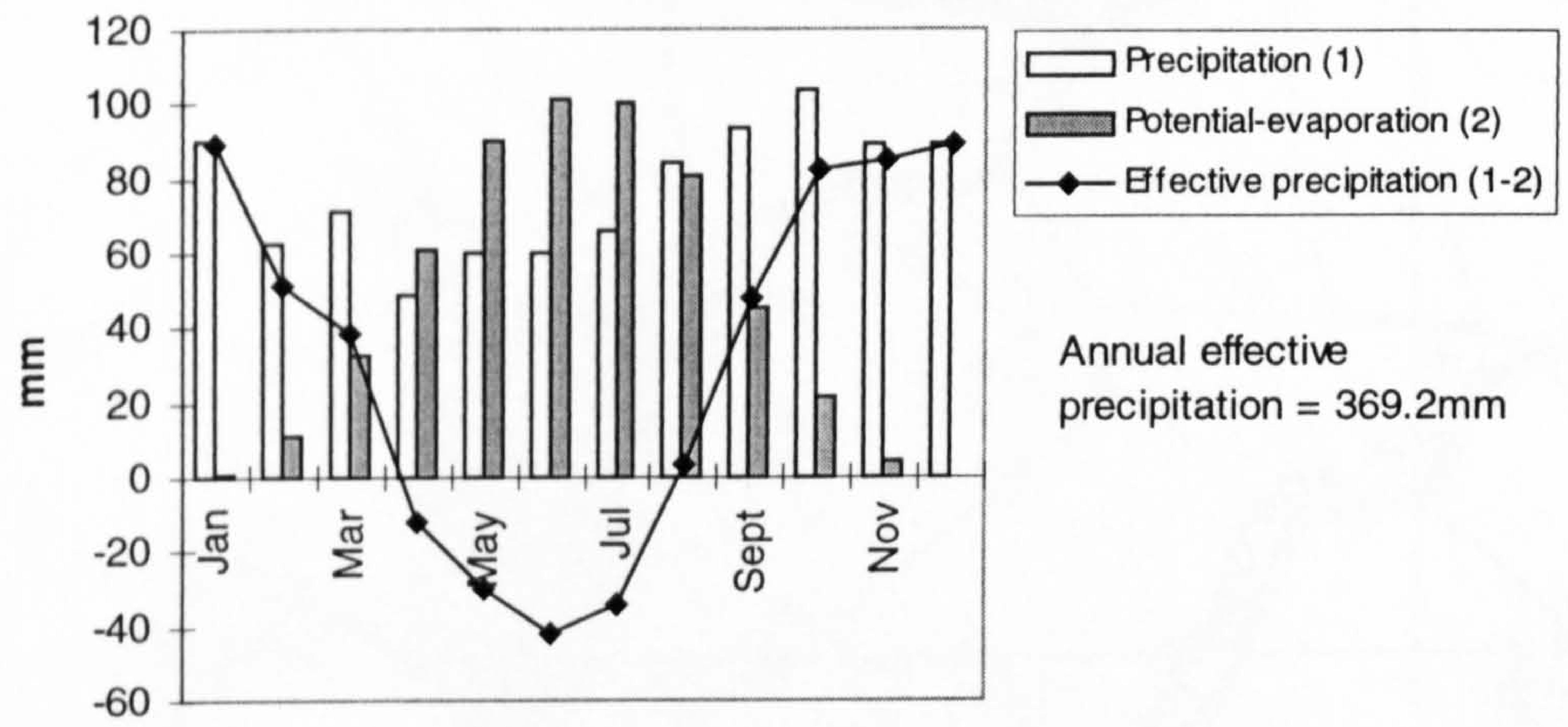
⥣ N. 0.________________500m.

× Core location.

4.2.1 Climate data

Longbridge Moss is a coastal lowland raised mire, with an altitude of 15m O.D., and therefore experiences relatively low amounts of precipitation compared with the other sites studied (Figure 4.2). It experiences the highest annual temperatures of any of the sites (Figure 4.3), which influences the potential evaporation, and hence annual effective precipitation (Figure 4.5) which is only 369.2mm, resulting in a large water deficit from April to July. It is possible that this may reduce the sensitivity of the site to climatic changes (Mauquoy, 1997), although it may also be the case that where a potentially large seasonal water deficit occurs the species assemblages respond to threshold changes, as was first suggested by Conway (1948).

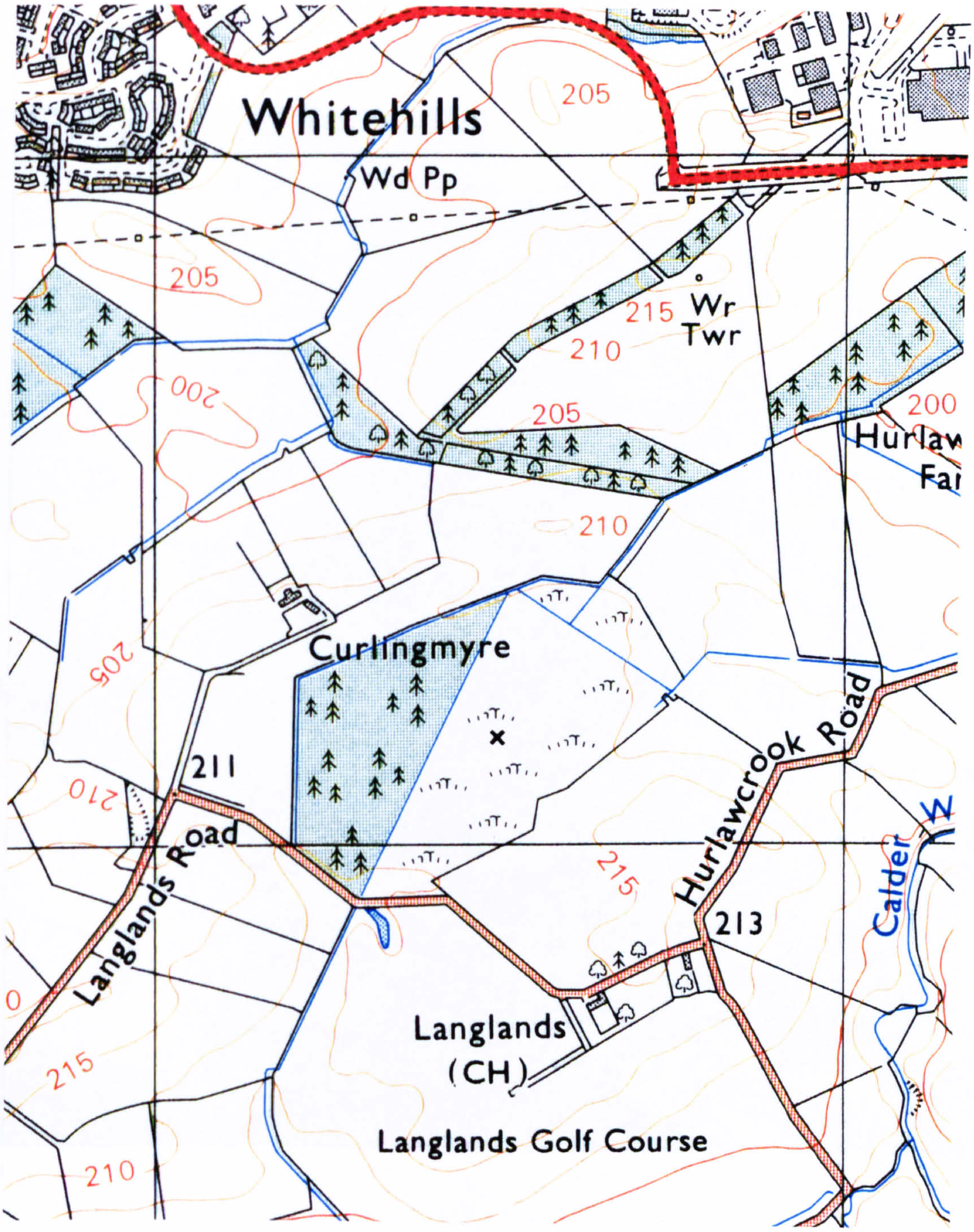
Figure 4.5 Water balance for Longbridge Moss



4.3 Langlands Moss (LAG) - NS 2634 6513

Langlands Moss is a nature reserve of the Scottish Wildlife Trust situated to the south of Glasgow near East Kilbride, only a mile or so from the NERC Radiocarbon Laboratory (Figure 4.6 and Plate 4.2). The site is *ca.* 25 hectares in size with a relatively large area of primary (not cut over) raised bog, which exhibits a marked domed appearance. The western half has been afforested, although the unafforested area still contains species typical of a raised bog. The trees were planted in the 1960's and coupled with the drains which had been emplaced, damaged the acrotelm and led to a reduced water table.

Figure 4.6 Map of Langlands Moss.



× Core location. Source: Ordnance Survey (1:25,000).

Plate 4.2 Aerial photograph of Langlands Moss.

⇑N. 0.___________________________________500m.

× Core location.

About 25,000 trees were removed from the surface in 1994 from the 10 hectare wooded area, and thus apart from the surface of the bog having been slightly damaged, a generally intact palaeoenvironmental archive remains. However, some peripheral peat extraction, drainage, and afforestation has led to a lowering of ambient water-levels in the bog, resulting in a loss of pool and lawn bog-vegetation communities (Brooks and Stoneman, 1997). The decreased levels of water input into the system, due to increased interception and evapotranspiration by the former trees has resulted in a drying of the surface and a damaged acrotelm. Therefore, presently, hummock vegetation species tend to dominate, notably *Calluna vulgaris, Erica tetralix, Eriophorum vaginatum*, as well as non-*Sphagnum* bryophytes, including *Hypnum cupressiforme* and *Polytrichum juniperum/alpestre* type. However, small areas of *Sphagnum* dominated lawns and hollows do exist.

A five metre profile was extracted for tephra analysis, but as high concentrations of tephra was only found in the top metre only three metres of peat were recovered for palaeoecological analysis from this site.

4.3.1 Climate data

Langlands Moss is situated at an altitude of 214m O.D., and has a mean annual temperature of 7.3^{0}C, being over 2^{0}C warmer than the coldest sites (Figure 4.3). It has relatively high levels of precipitation (Figure 4.2), with mean annual levels being 1461.3mm, although mean annual effective precipitation is 971.4mm. A slight summer moisture deficit exists over the months of May and June (Figure 4.7), which is possibly influenced by the relatively high temperatures at the site. However, the moisture deficit is only slight, and it is suggested that it would not affect the growth of *Sphagnum* to any great degree as the main increase in length of *Sphagnum* is determined by precipitation throughout the year, but specifically the preceding August, and between June and August of the current year (Backeus, 1988). These data would therefore suggest that conditions are probably near optimal for *Sphagnum* growth. However, changes in effective precipitation initiated by climatic changes would alter this relationship and influence *Sphagnum* growth, suggesting that this site is probably sensitive to such changes.

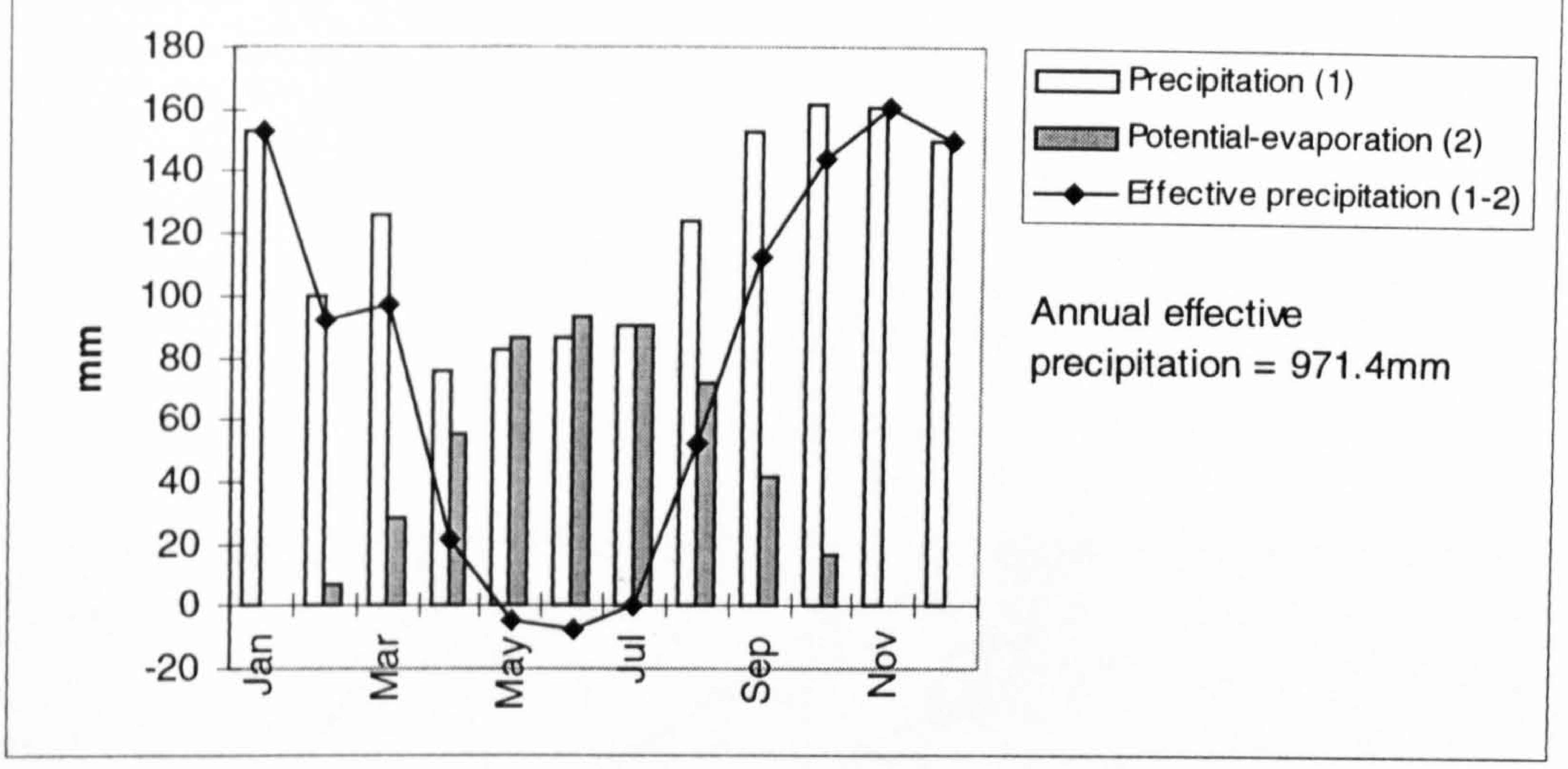


Figure 4.7 Water balance for Langlands Moss

Annual effective precipitation = 971.4mm

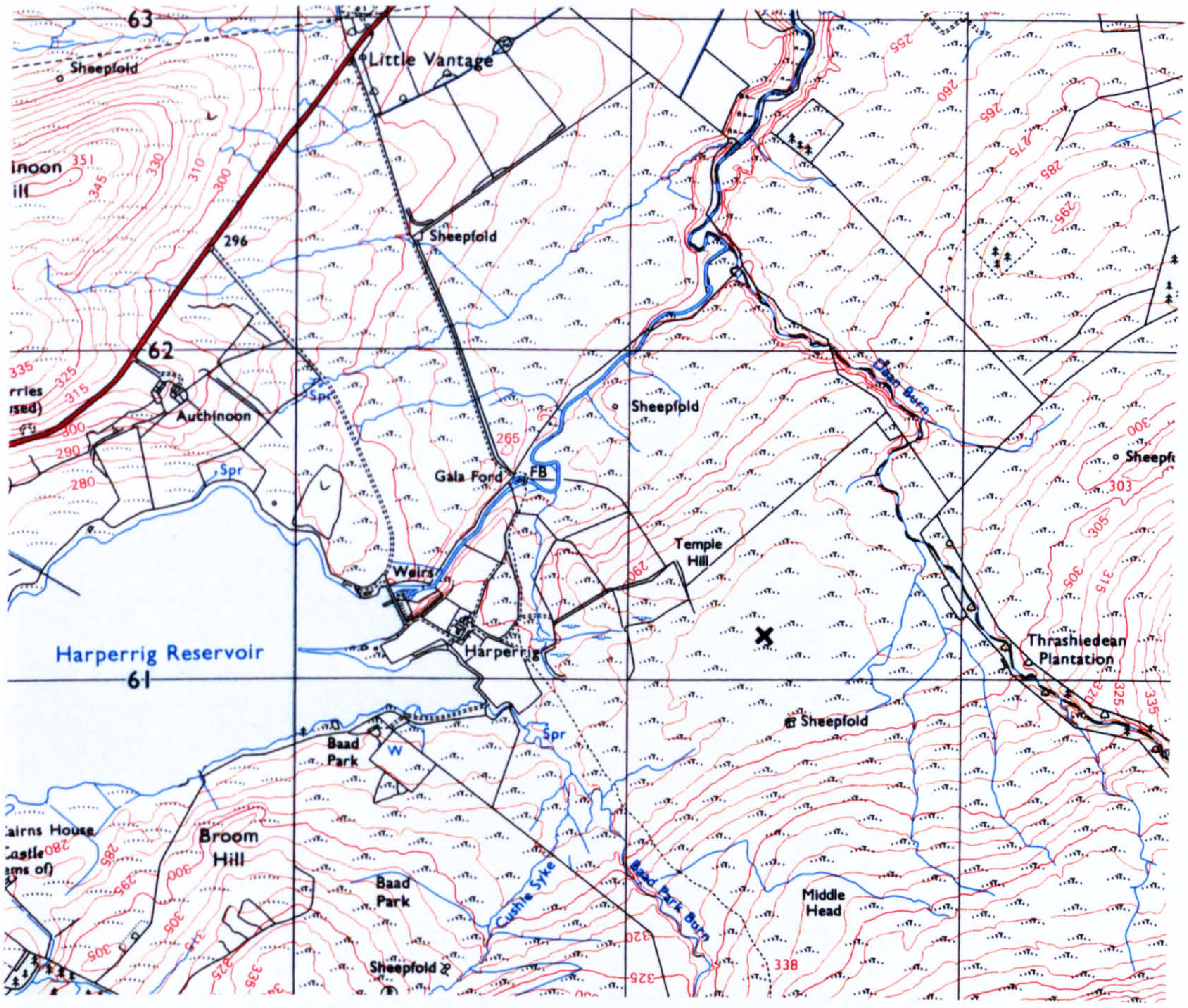
4.4 Temple Hill Moss (TEM) - NT 3113 6613

Temple Hill Moss is located just northeast of Harperrig Reservoir on the northern flank of the Pentland Hills, near the village of Balerno, about *ca.* 10 miles southwest of Edinburgh (shown in Figure 4.8 and Plate 4.3). This moss has an overall area of 43 hectares, and although there is a hint of some old drains from the aerial photograph, most of the site is in good condition. The southern edge of the mire on the approach from Harperrig consists of relatively bare peat, with hagging in places, and the possible remnants of a rand. The site, however, is structurally intact (as demonstrated in the stratigraphy, see Section 6.8.1), although in the past some burning and drainage has taken place. This is reflected in the present day flora, which is dominated by low hummocks with *Calluna vulgaris, Erica tetralix*, and *Eriophorum vaginatum* being locally abundant. However, *Eriophorum angustifolium, Sphagnum capillifolium,* and *Sphagnum fuscum* are also found on the surface, with *Sphagnum cuspidatum* occupying the hollows. A five metre core was taken at this site, and two tephra horizons located, hence the five metres of peat were retained for palaeoecological analysis.

4.4.1 Climate data

The relatively high altitude of Temple Hill Moss (294m O.D.), and the proximity of the

Figure 4.8 Map of Temple Hill Moss.



× Core location. Source: Ordnance Survey (1:25,000).

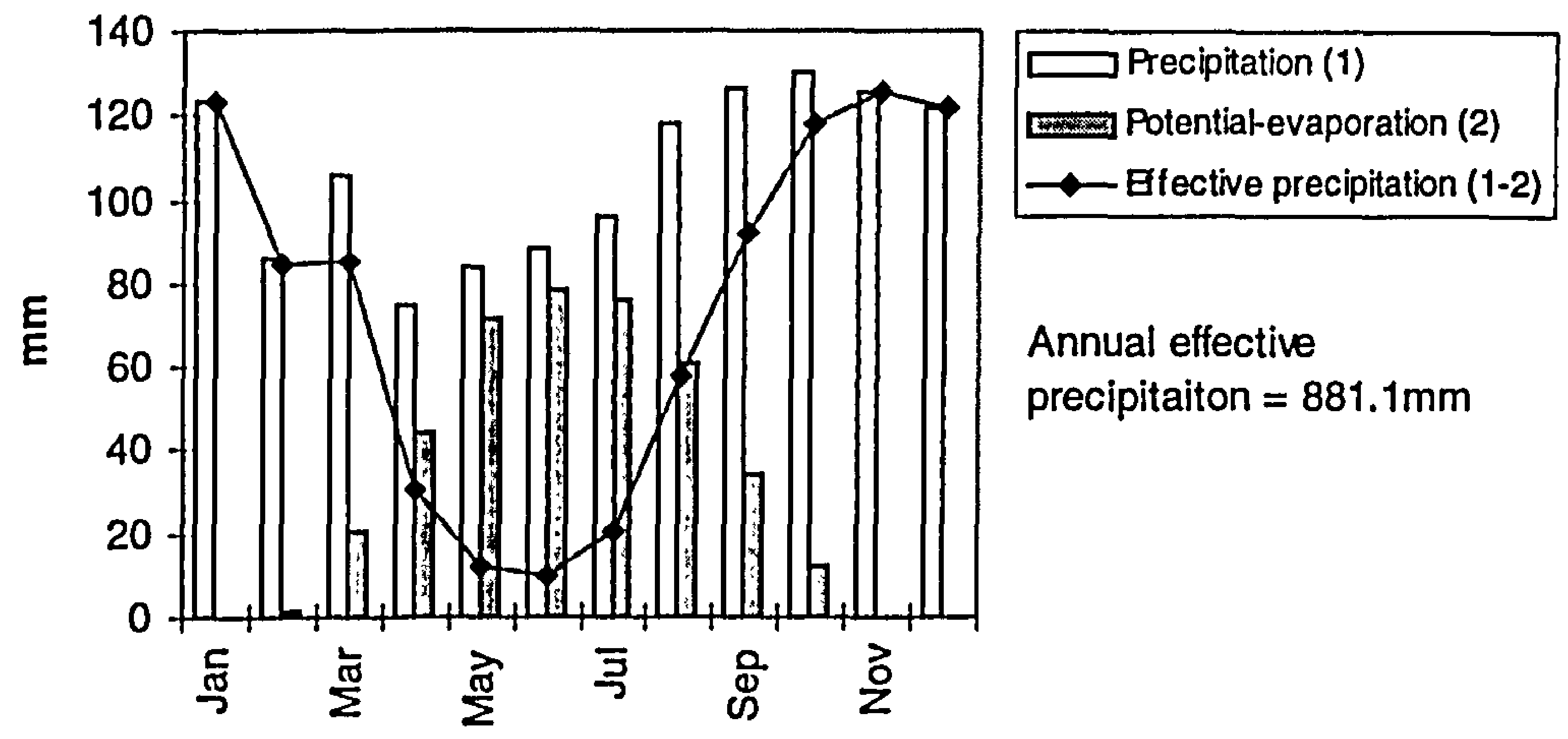
Plate 4.3 Aerial photograph of Temple Hill Moss.

↖ N. 0.________________500m.

× Core location.

Pentland Hills ensure that this site receives a relatively large amount of rainfall (Figure 4.2), with an annual precipitation of 1280.7mm. This enabled the region to be classified as euoceanic by Birse (1971), and not the more continental hemioceanic, considering it is situated towards the east coast of Scotland. The cool annual temperatures (Figure 4.3) lower the potential evaporation of the site, ensuring a relatively high annual effective precipitation of 881.1mm, and no mean summer water deficit (Figure 4.9).

Figure 4.9 Water balance for Temple Hill Moss

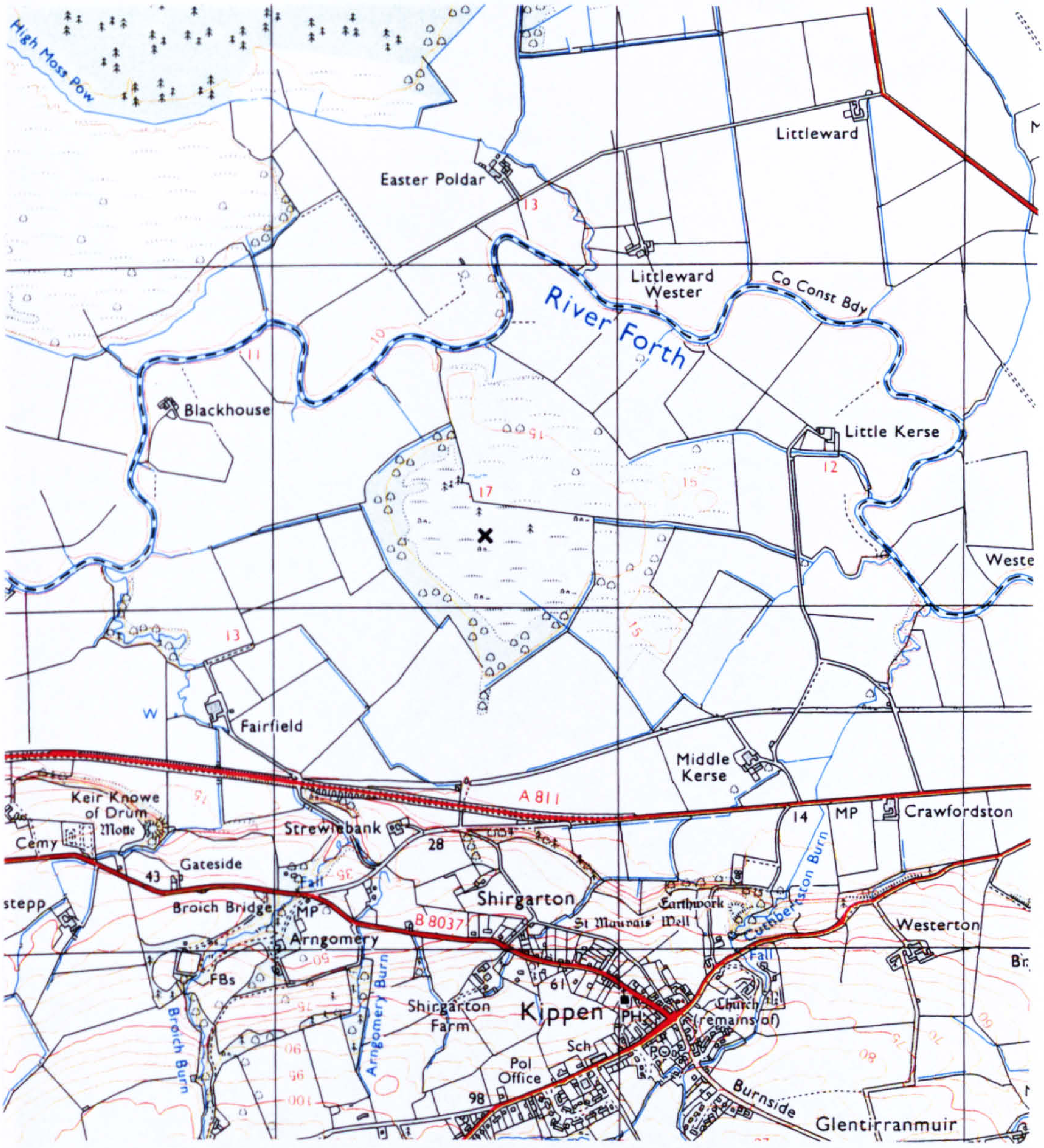


4.5 Shirgarton Moss (SGM) - NS 2645 6963

Shirgarton is located just to the south of the River Forth in the Carse of Stirling (Figure 4.10), and has a paired site within *ca.* 2 miles to the west, Kilhorn Moss. Most of the valley of the upper Forth was probably formerly covered with raised peat deposits of this nature, but now only scattered remnants can be found. The large mire complex of Flanders Moss lies just to the north and has been studied extensively (see Haslam, 1987).

Shirgarton Moss covers an area of 58 hectares of which about 16 hectares are in good condition. The bog is structurally intact although threatened by scrub invasion, notably *Betula pubescens* and *Salix* spp. The aerial photograph (Plate 4.4) shows some evidence of small drainage ditches, and a small amount of peat cutting, although the bog seems to have suffered little. A well developed cover of *Sphagnum* exists with

Figure 4.10 Map of Shirgarton Moss.



× Core location. Source: Ordnance Survey (1:25,000).

Plate 4.4 Aerial photograph of Shirgarton Moss.

⇑ N. 0.______________________________500m.

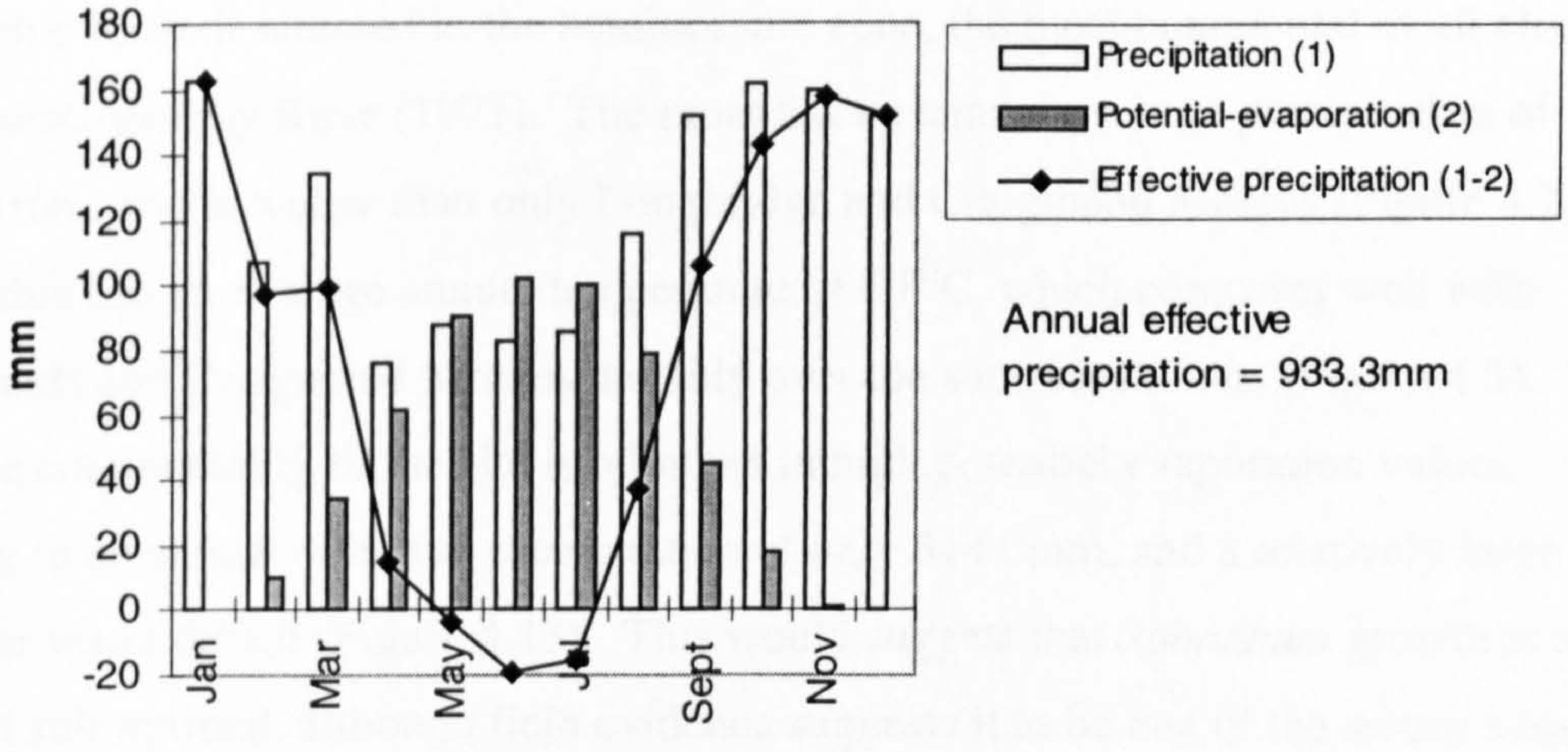
× Core location.

Sphagnum magellanicum and *Sphagnum papillosum* lawns plentiful. *Eriophorum vaginatum* and *Sphagnum capillifolium* var. *rubellum* tend to dominate hummocks along with *Calluna vulgaris* and *Erica tetralix*. *Vaccinium oxycoccus* and *Andromeda polifolia* are locally common. A three metre core was taken from this site for palaeoecological analysis.

4.5.1 Climate data

Shirgarton Moss receives a relatively large amount of rainfall (Figure 4.2), with an annual precipitation of 1476.2mm. This lowland site has an altitude of 18m O.D., but is surrounded by the Grampian Mountains to the north and west, and the Campsie Fells to the south, ensuring high levels of precipitation. However, the low altitude of the site is characterised by relatively high temperatures, with an annual average of 8.6^{0}C, being the warmest site along with Longbridge Moss (Figure 4.3). This leads to relatively high potential evaporation levels, and a subsequent slight water deficit during summer months (Figure 4.11). However, annual effective precipitation levels are comparably high, at 933.3mm, and it is assumed that under present conditions the moss is optimal for *Sphagnum* growth, similar to Langlands Moss.

Figure 4.11 Water balance for Shirgarton Moss



Annual effective precipitation = 933.3mm

4.6 Mallachie Moss (MAL) - NH 2963 8170

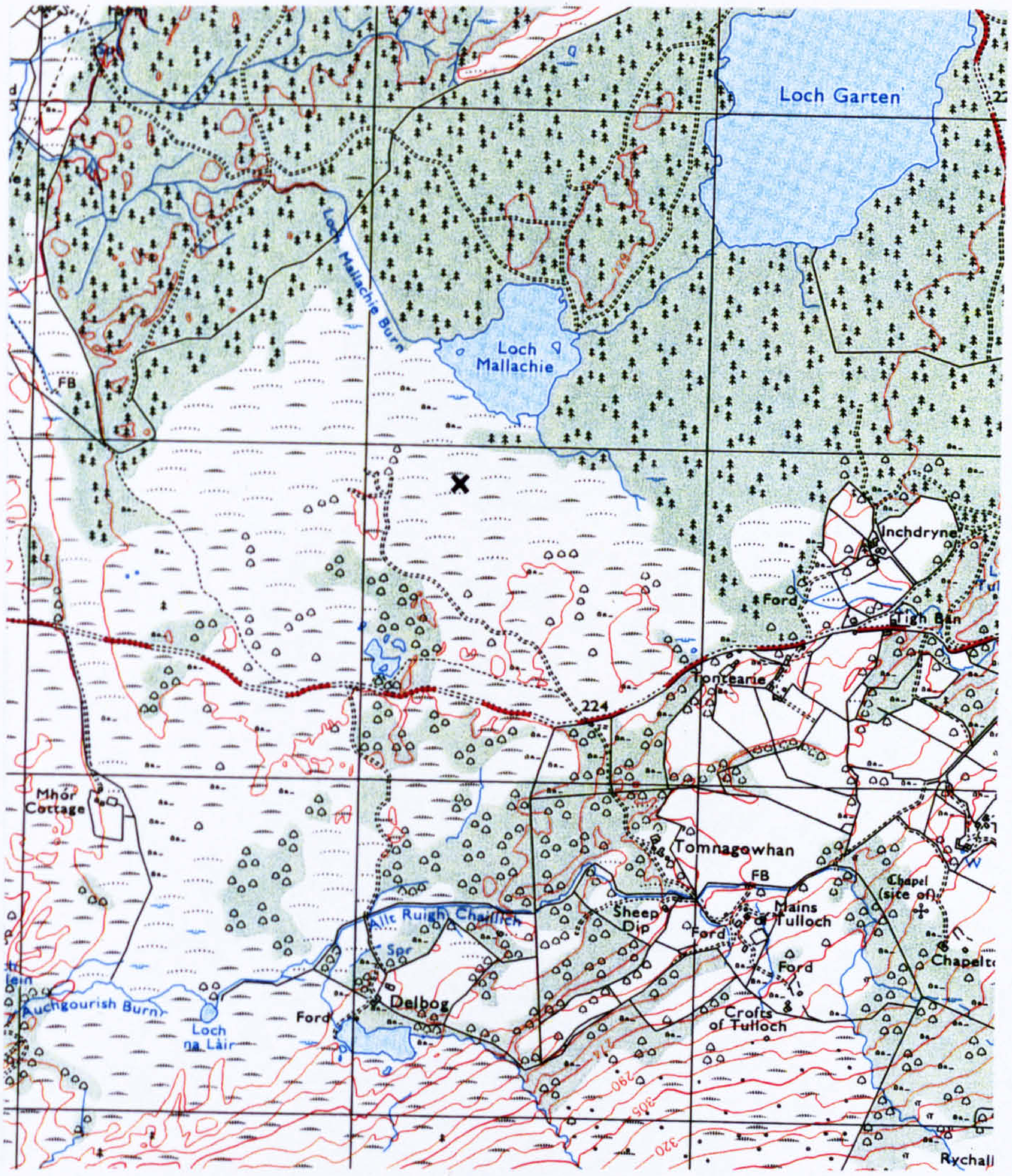
Mallachie Moss is an area of raised bog within Abernethy Forest, to the southwest of Loch Mallachie, comprising of approximately 80 hectares (Figure 4.12). Although a site at Abernethy Forest has previously been investigated by Birks (1970) and Birks and Mathewes (1978) their research was confined to Lateglacial and early Holocene sediments found in a glacial channel infill between Loch Mallachie and Loch Garten. The aerial photograph of the site (Plate 4.5) shows how the edges of the mire have been cut, and this was also noted in the field with pools situated around the mire edge having straight sides. A tough band of *Eriophorum vaginatum* at *ca.* 30cm depth can be traced across the mire surface which may have resulted from a period of disturbance (see Section 6.10.1 and Figure 6.39).

Large swathes of *Sphagnum magellanicum* and *Sphagnum papillosum* are abundant as is *Sphagnum cuspidatum* in the hollows. *Scirpus cespitosus* can be found on some of the drier parts of the mire, along with *Eriophorum vaginatum* tussocks. Tephra was searched for from a 5 metre peat core but only found in the top 3 metres, so a three metre core was taken from this site for palaeoecological analysis.

4.6.1 Climate data

Mallachie Moss is situated in the hemioceanic zone, the most continental of all climatic zones discussed by Birse (1971). The moss has an annual average precipitation of 1005.1mm, and is wetter than only Longbridge and Craigmaud Mosses (Figure 4.2). Mallachie has an average annual temperature of 6.8^{0}C, which compares well with Langlands and Craigmaud Mosses, notably over the summer months (Figure 4.3). The relative continentality of the site is reflected in high potential evaporation values, leading to an annual effective precipitation of only 514.0mm, and a relatively large summer water deficit (Figure 4.13). This would suggest that *Sphagnum* growth is at present sub-optimal, although field evidence suggests it to be one of the wetter sites. It is therefore possible that the *Eriophorum vaginatum* layer at 30cm, may be artificially raising the water table and promoting *Sphagnum* growth.

Figure 4.12 Map of Mallachie Moss.



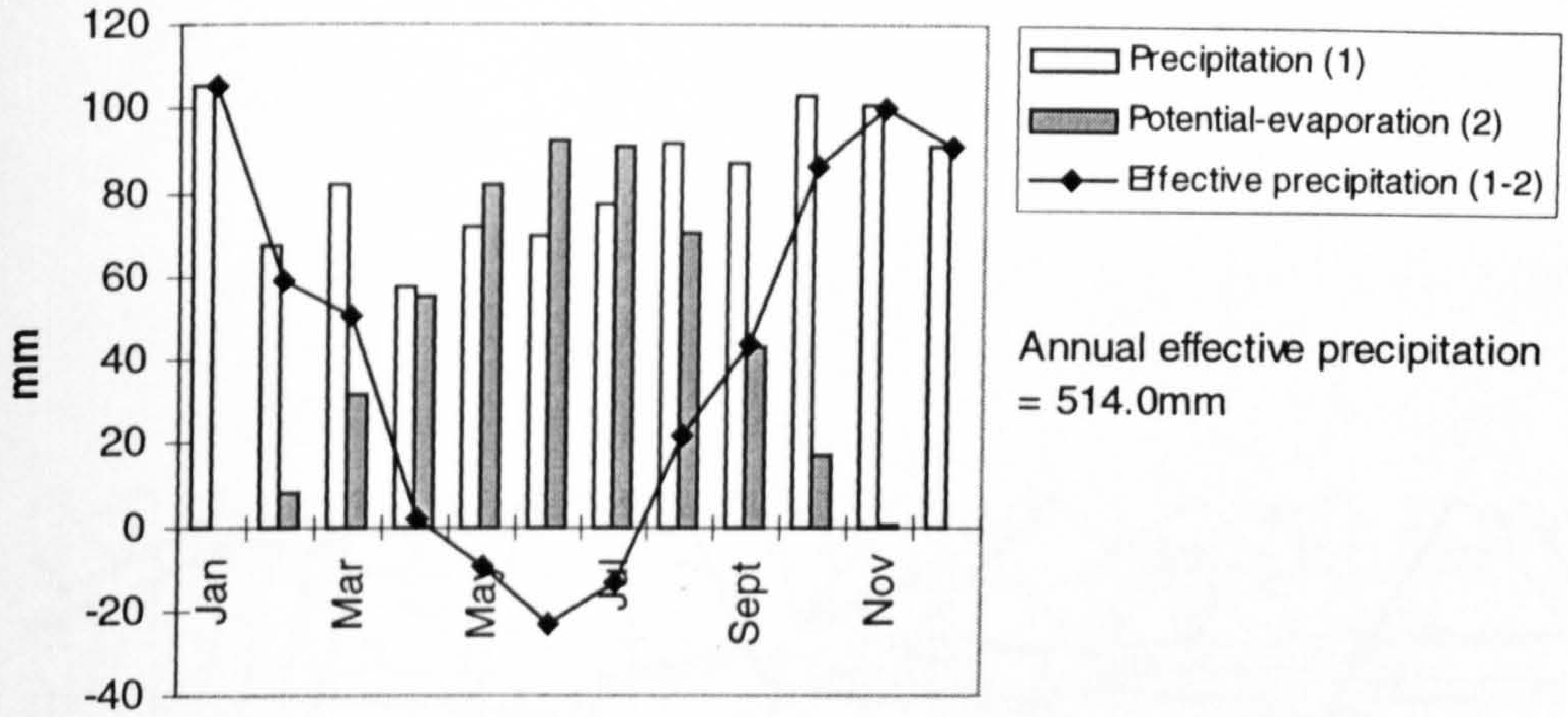
× Core location. Source: Ordnance Survey (1:25,000).

Plate 4.5 Aerial photograph of Mallachie Moss.

⥣N. 0.____________________500m.

× Core location.

Figure 4.13 Water balance for Mallachie Moss



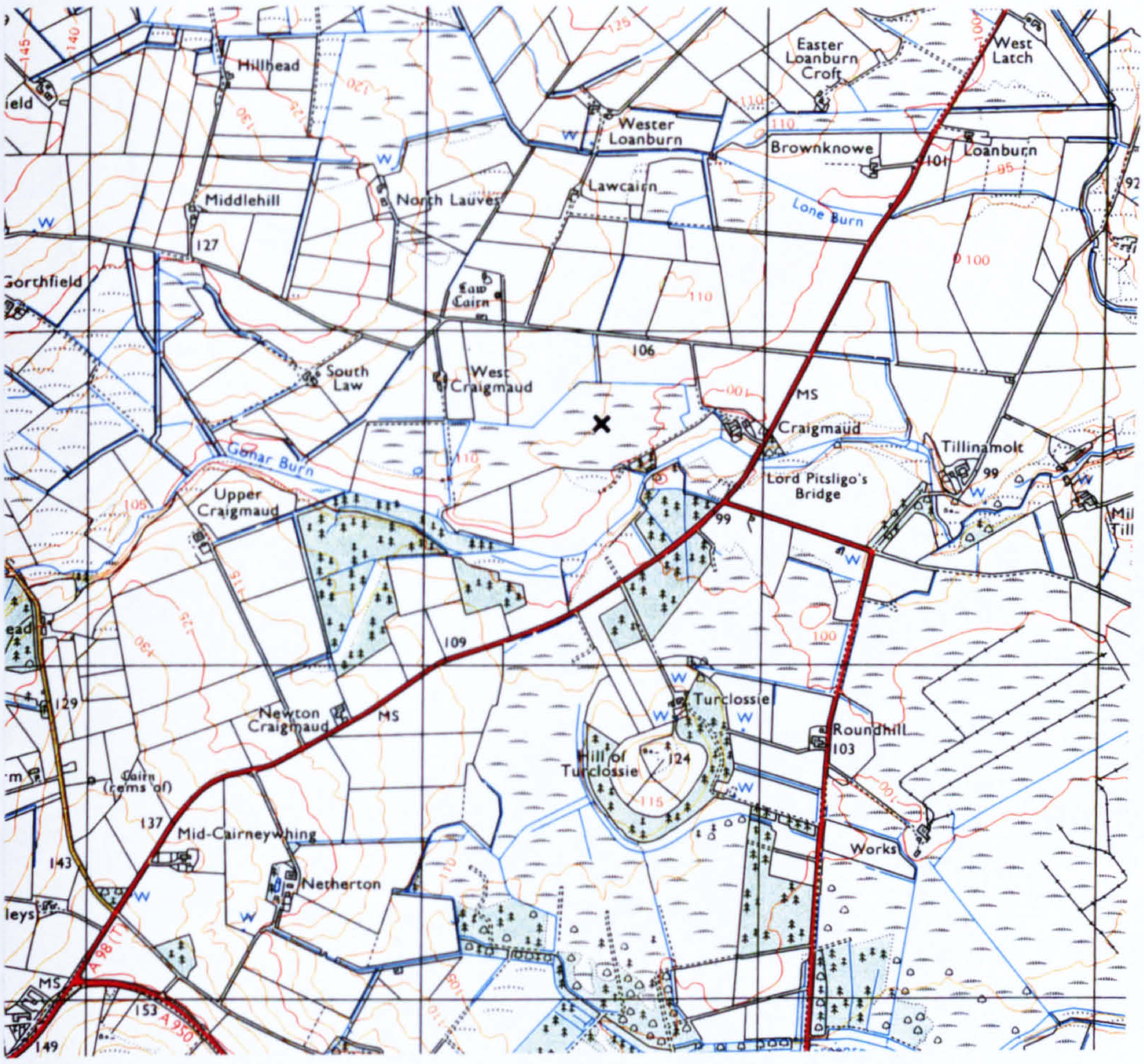
Annual effective precipitation = 514.0mm

4.7 Craigmaud Moss (CMD) - NJ 3885 8588

The Grampian lowlands were formerly covered in large peatlands which are now almost all destroyed, either having been drained for agriculture, or cut for domestic and commercial purposes. However, some sites do survive, and as this area typically exhibits a comparatively dry climate, a study site was sought. Many of the Scottish National Heritage (SNH) listed ombrotrophic peatlands of this area have been disturbed anthropogenically, and most exhibit predominantly *Calluna vulgaris - Eriophorum vaginatum* communities (NVC 19/20).

Craigmaud Moss is an example of one of these lowland raised mires (Figure 4.14, Plate 4.6), being one of the least disturbed, but due to the comparatively dry climate much of the peat has been well humified, rendering stratigraphic investigations ineffective. This is a small site of *ca.* 17 hectares on which *Calluna vulgaris*, *Erica tetralix* and *Eriophorum vaginatum* are abundant and the majority of *Sphagnum* cover comprises of *Sphagnum capillifolium*. *Sphagnum cuspidatum* and *Pleurozium schreberi* are also locally common. A core was taken to 275cm, due to the considerable amount of wood below this depth. Two tephras were identified in the top 140cm and so palaeoecological analysis was undertaken to this depth.

Figure 4.14 Map of Craigmaud Moss.



× Core location. Source: Ordnance Survey (1:25,000).

Plate 4.6 Aerial photograph of Craigmaud Moss.

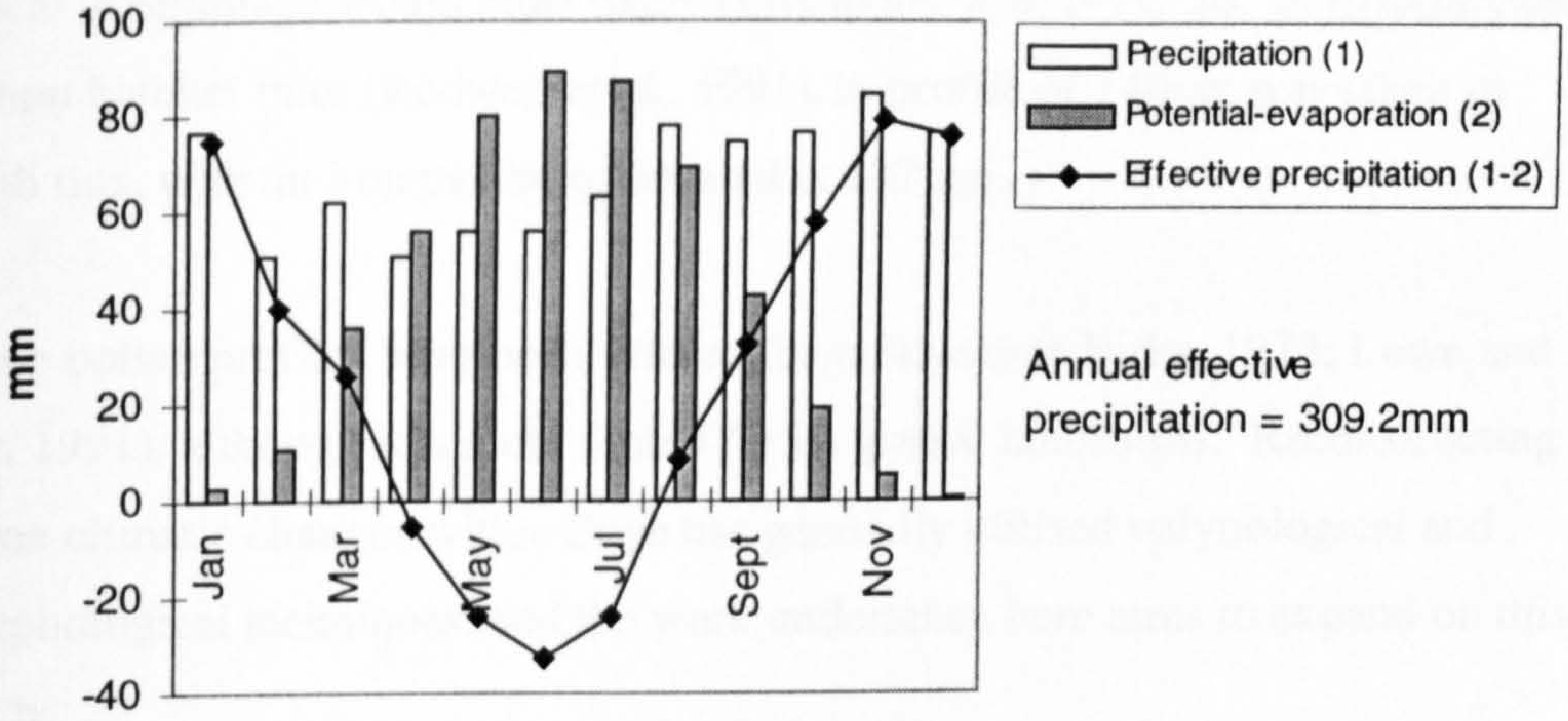
⥣N. 0.__________500m.

× Core location.

4.7.1 Climate data

Craigmaud Moss lies on the edge of Birse's (1971) hemioceanic/euoceanic boundary, as the far east coast sites can have a maritime influence from the east. However, Craigmaud has the lowest annual precipitation (808.2mm) of all the sites (Figure 4.2), although a maritime influence can be detected in its average annual temperature of 7.6^{0}C which is the third highest of all the sites. However, these characteristics result in a relatively high potential evaporation, leading to a low annual effective precipitation (309.2mm) and a large summer water deficit (Figure 4.15). This would suggest that growing conditions for *Sphagnum* are sub-optimal, and this is seen in the field as the mire is dominated mainly by Ericaceae and *Eriophorum vaginatum*.

Figure 4.15 Water balance for Craigmaud Moss



Annual effective precipitation = 309.2mm

4.8 Ben Gorm Moss (BGM) - NG 1432 8658

In order to complete a geographical and climatological transect of Scotland is was necessary to select a site on the west coast. It has been suggested that the ombrogenous mires of the Inner Hebrides represent a hyper-oceanic extreme, found only in Europe along Scotland's Atlantic seaboard and the west coast of Ireland (Goodwillie, 1980). Species able to tolerate the oceanic climate along the Atlantic coast of Scotland find there a greater variety of surface features in which they may obtain a suitable niche (Lindsay *et al.*, 1983). A site on the west coast would therefore present a location with differing climatological and ecological criteria as compared with the other localities in the study.

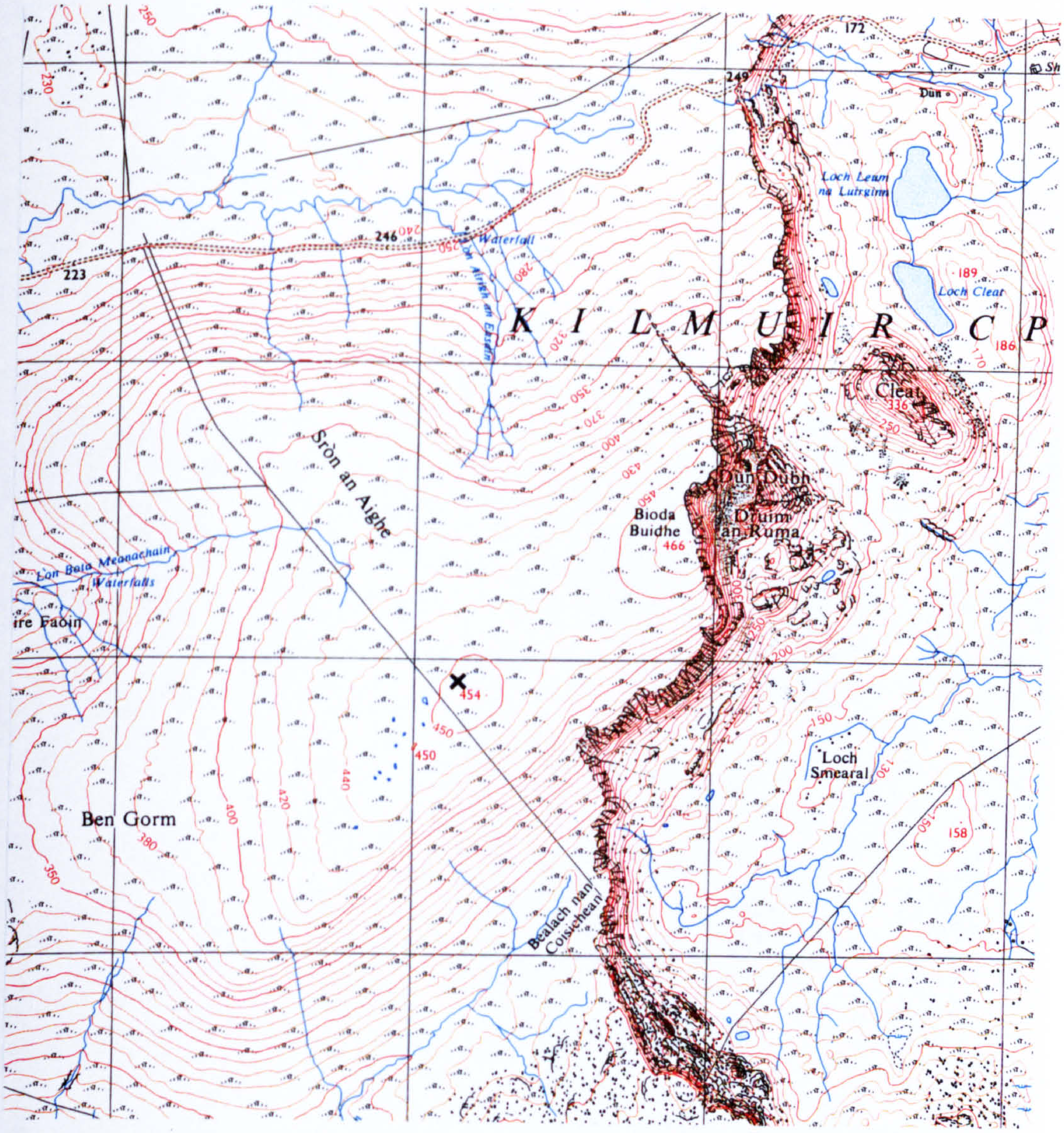
A reconnaissance of potential sites along the west coast was undertaken, and a water shedding blanket bog on the Isle of Skye selected. Finding intact raised bogs on the west coast proved difficult, and the Isle of Skye suggested an excellent location in terms of a geographical cross section across north central Scotland. Ben Gorm Moss is an upland blanket bog located within the Trotternish region of North Skye (Figure 4.16, Plate 4.7). Covering only *ca.* 10 hectares, it is situated on top of a water shedding ridge, spanning the peak of Bioda Buidhe. Locally dominant are *Eriophorum vaginatum, Racomitrium lanuginosum, Hypnum cupressiforme, Rhytiladelphus loreus,* and *Scirpus cespitosus.* Pockets of *Vaccinium oxycoccus* and *Empetrum nigrum* can be found, as well as areas of *Sphagnum capillifolium.* The lichen *Cladonia* is also present, and this ecological assemblage would most likely correspond with NVC 20, an *Eriophorum vaginatum* blanket mire (Rodwell *et al.*, 1991). A profile of 140cm was taken in monolith tins, with an iron pan being located at 137 cm.

Holocene pollen profiles have been studied from Skye (see Birks, 1973; Lowe and Walker, 1991), although it is more famed for its glacial landforms. Reconstructing Holocene climatic changes within Skye has generally utilised palynological and geomorphological techniques, and the work undertaken here aims to expand on this approach.

4.8.1 Climate data

Ben Gorm Moss possess the most extreme climate of any of the sites, having the highest average annual precipitation (2499.6mm) and the lowest average annual temperature (4.7^{0}C). Its geographical location in the hyperoceanic sector of Birse (1971), with an altitude of 454m O.D. ensures that potential evaporation is kept to a minimum and the average annual effective precipitation is high (2142.6mm), as shown in Figure 4.17. This sub-alpine climatic regime is comparable with other blanket mire sites which have been shown to be sensitive to climatic change, notably Migneint in Snowdonia (Blackford, 1990) and Moine Mhor in the Cairngorms (Barber *et al.*, 1999).

Figure 4.16 Map of Ben Gorm Moss.



× Core location. Source: Ordnance Survey (1:25,000).

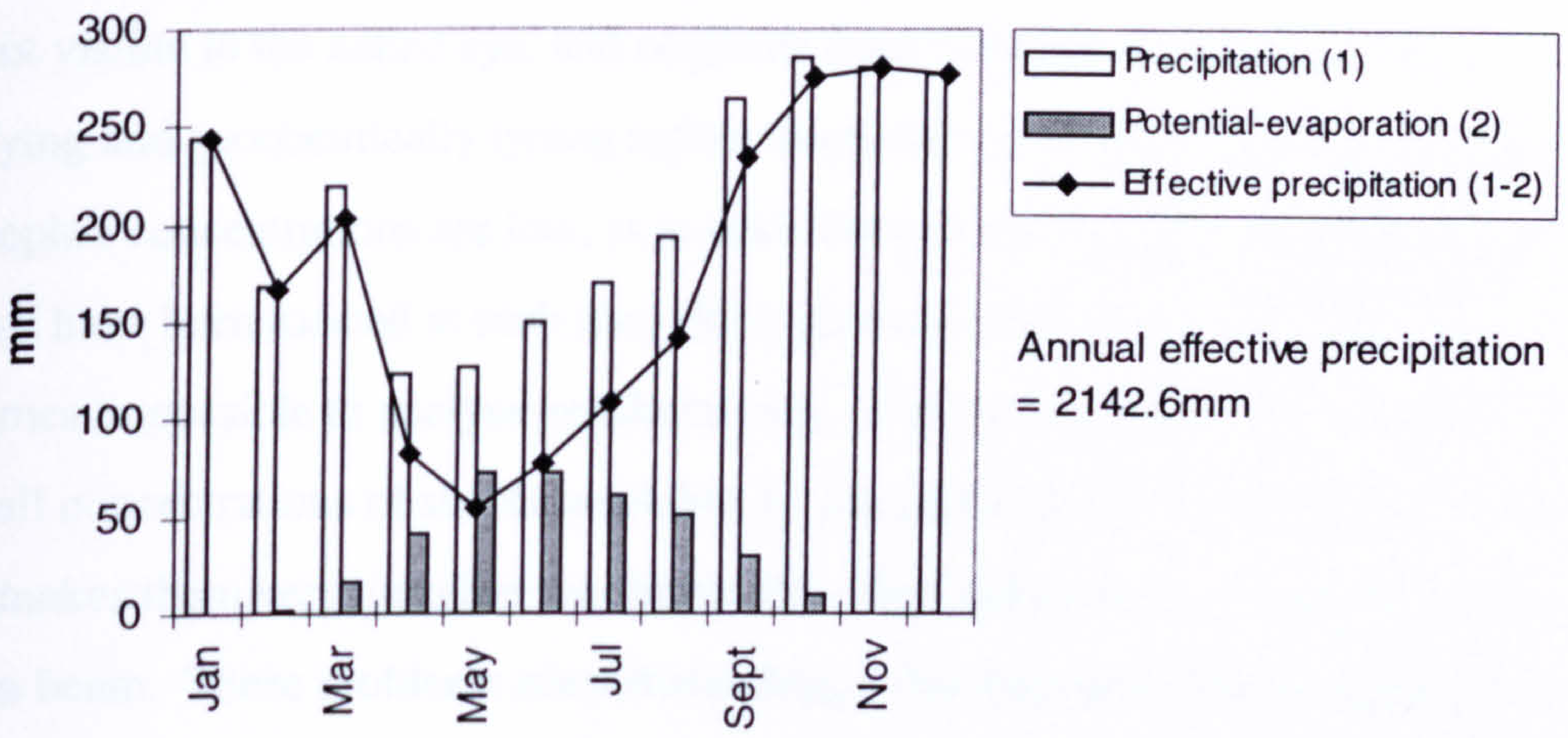
Plate 4.7 Aerial photograph of Ben Gorm Moss.

⥣N. 0.__________500m.

× Core location.

Mauquoy (1997) analysed an upland blanket mire at Shaft Hill National Nature Reserve in Cumbria, with similar climatic conditions to Ben Gorm. He found the mire to be comparatively insensitive to climatic change, but attributed this to a hiatus within the stratigraphy, and not local hydrological regimes.

Figure 4.17 Water balance for Ben Gorm Moss



Annual effective precipitation = 2142.6mm

Chapter 5.

Chronology: tephrochronology and radiocarbon dating

5.1 Introduction

The Holocene tephras found in Scotland are from stratigraphically discrete horizons in the peat, not visible to the naked eye, and originate from volcanic activity in Iceland. Identifying and geochemically typing tephra isochrones is not always easy, especially when tephra concentrations are low, as is often the case in Scotland. Potential tephra horizons have been located at each site, although some have proved difficult, and sometimes impossible to analyse geochemically. The main problems are usually due to the small concentrations of shards available, or the small size of the shards themselves, which makes them very unstable geochemically when subjected to analysis with an electron beam. These problems notwithstanding, it has been possible to construct a tephrochronology for the seven sites studied. This not only refines the geochronology of the sites, but allows precise correlations between the proxy palaeoecological records to be made, and also provides insights into the spatial and geochemical variability of tephra dispersal.

This chapter has two sections; the first on tephrochronology, and the second on radiocarbon dating. It culminates in the construction of age-depth models for each site, utilising the tephra isochrones as 'pinning points' (Barber *et al.*, 1999) within each site chronology.

5.2 Tephrochronology

5.2.1 Ashing results

Twelve sites were initially analysed for tephra content using the ashing method of Pilcher and Hall (1992), described in Section 3.2.2.1. The results are displayed in Figures 5.1(a-l) and clearly show where the main tephra isochrones occur. The data displayed in Figures 5.1 (a-l) relate to samples from the initial 50x5cm 'prospecting' core taken at each site, with the exceptions of Craigmaud Moss and Ben Gorm Moss, where the sampling was undertaken on the master cores. The advantage of initially using a prospecting core was to preserve material in the master core for further tephra and palaeoecological analyses, and more crucially, for radiocarbon dating.

The diagrams show the total number of shards against depth for contiguous 5x1cm samples, which are the result of counting each shard that has been identified under a 18x18mm cover slip. This process will not always represent the whole ashed sample, which occasionally was too large to be counted, and certain levels may therefore be underrepresented. However, the analysis was undertaken only to identify tephra isochrones, not as an accurate representation of the total amount of shards present, which in itself is variable within a peat core. Nonetheless, the technique is consistent and does give a first impression of tephra loading.

Distinct horizons can be seen in Figures 5.1(a-l) where there are concentrations of tephra shards, indicating a volcanic airfall event, almost certainly of Icelandic origin. Some of the diagrams show levels where only a few shards have been identified, but not necessarily enough to suggest that an isochrone is present. When these very low counts of shards occur relatively near to clearly identified tephra horizons, it is suggested that some of the shards could have been moved down the peat profile by plant roots or through dead root channels. Most *Eriophorum vaginatum* roots, for example, can penetrate to depths of 60cm below the surface (Gore and Urquhart, 1966). Where small quantities of tephra occur at isolated stratigraphic levels, Dugmore *et al.* (1995a) suggest three possible options: that the tephra may represent a primary airfall event, that the shards may have undergone reworking, or a combination of the two. Dugmore *et al.* (1995a) found that less than 1% of shards from all their analyses formed distinct chemical outliers from the main populations, suggesting that these deposits are of primary volcanic events. If the shards recovered from the sites located in Figures 5.1(a-l) are not of reworked grains, then the possibility must exist that they could be from a primary volcanic event, and further research into finding tephra isochrones at these sites/stratigraphic levels is to be encouraged. However, further research into these horizons was curtailed due to the lengthy time periods needed for such investigations. The absence of larger concentrations

Figure 5.1a Tephra shard abundance at Longbridge Moss

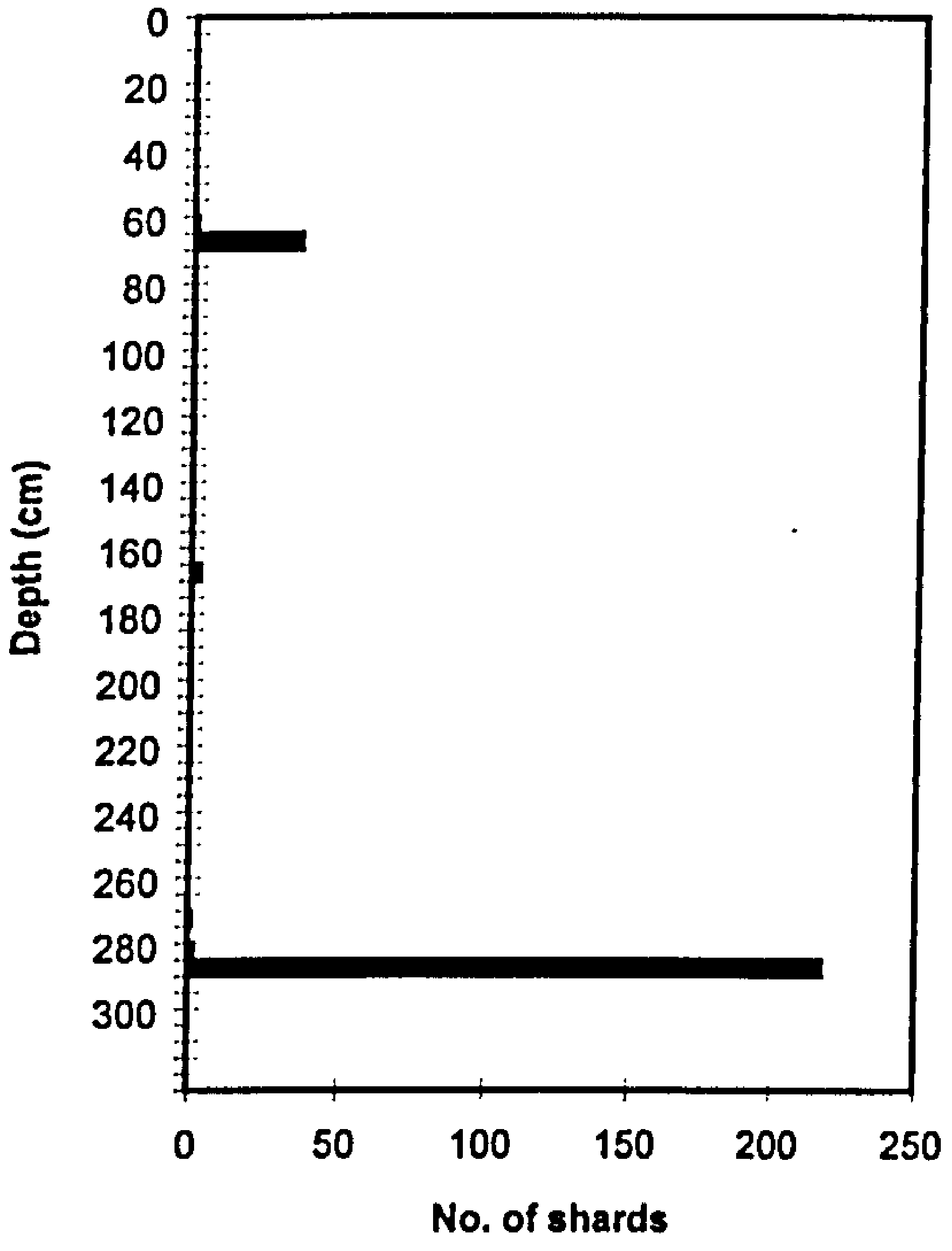


Figure 5.1b Tephra shard abundance at Carsegowan Moss

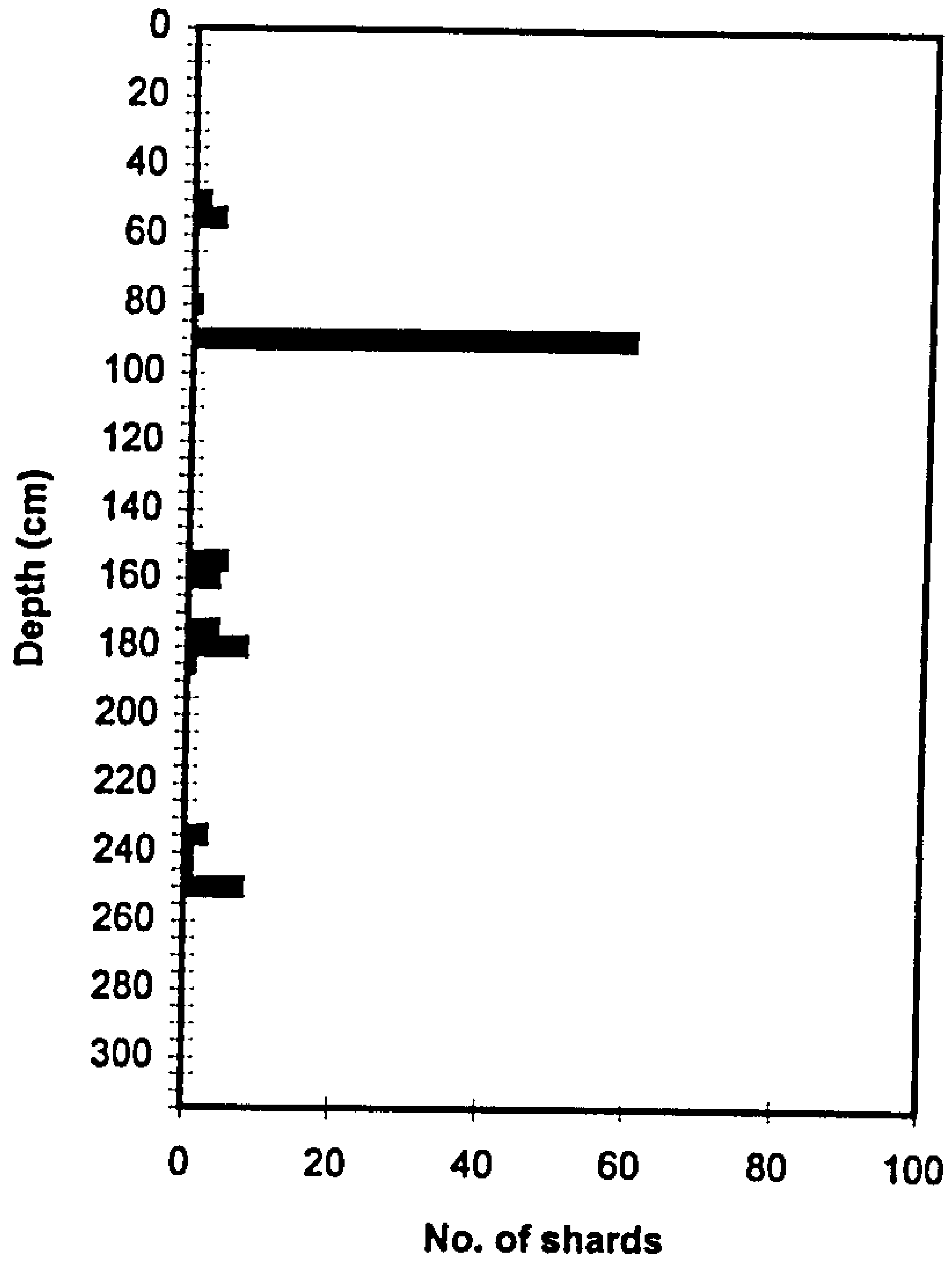


Figure 5.1c Tephra shard abundance at Shirgarton Moss

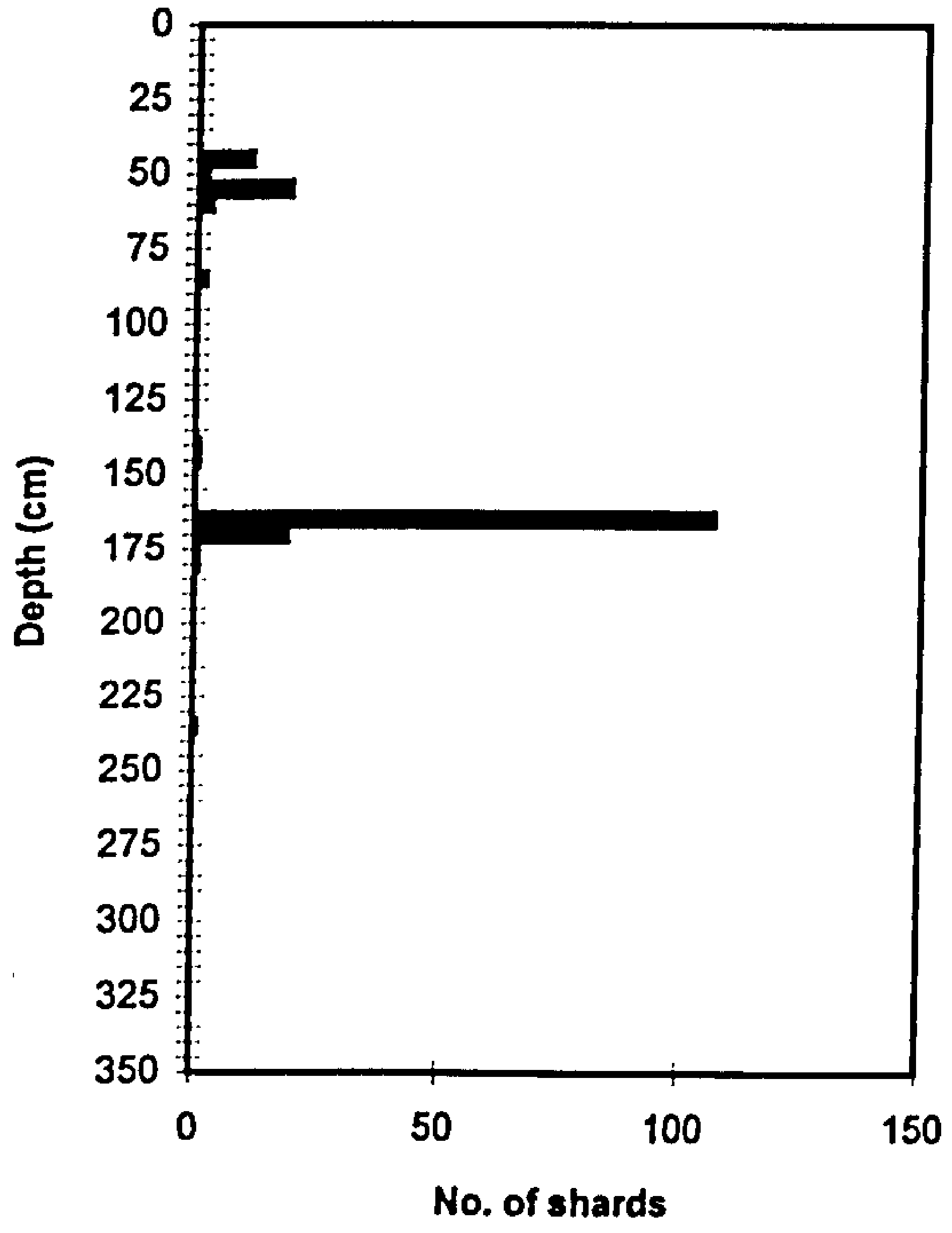


Figure 5.1d Tephra shard abundance at Kilhorn Moss

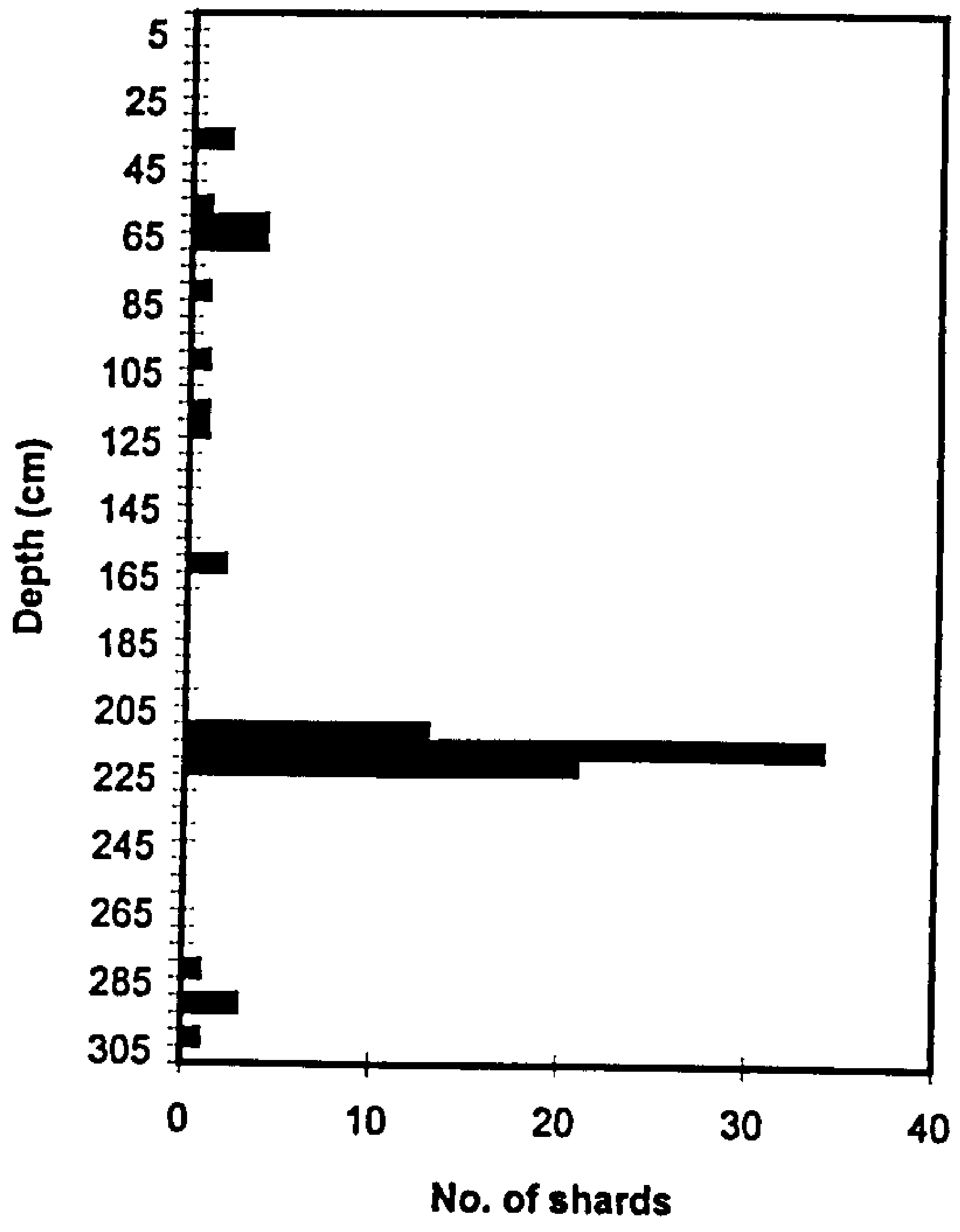


Figure 5.1e Tephra shard abundance at Temple Hill Moss

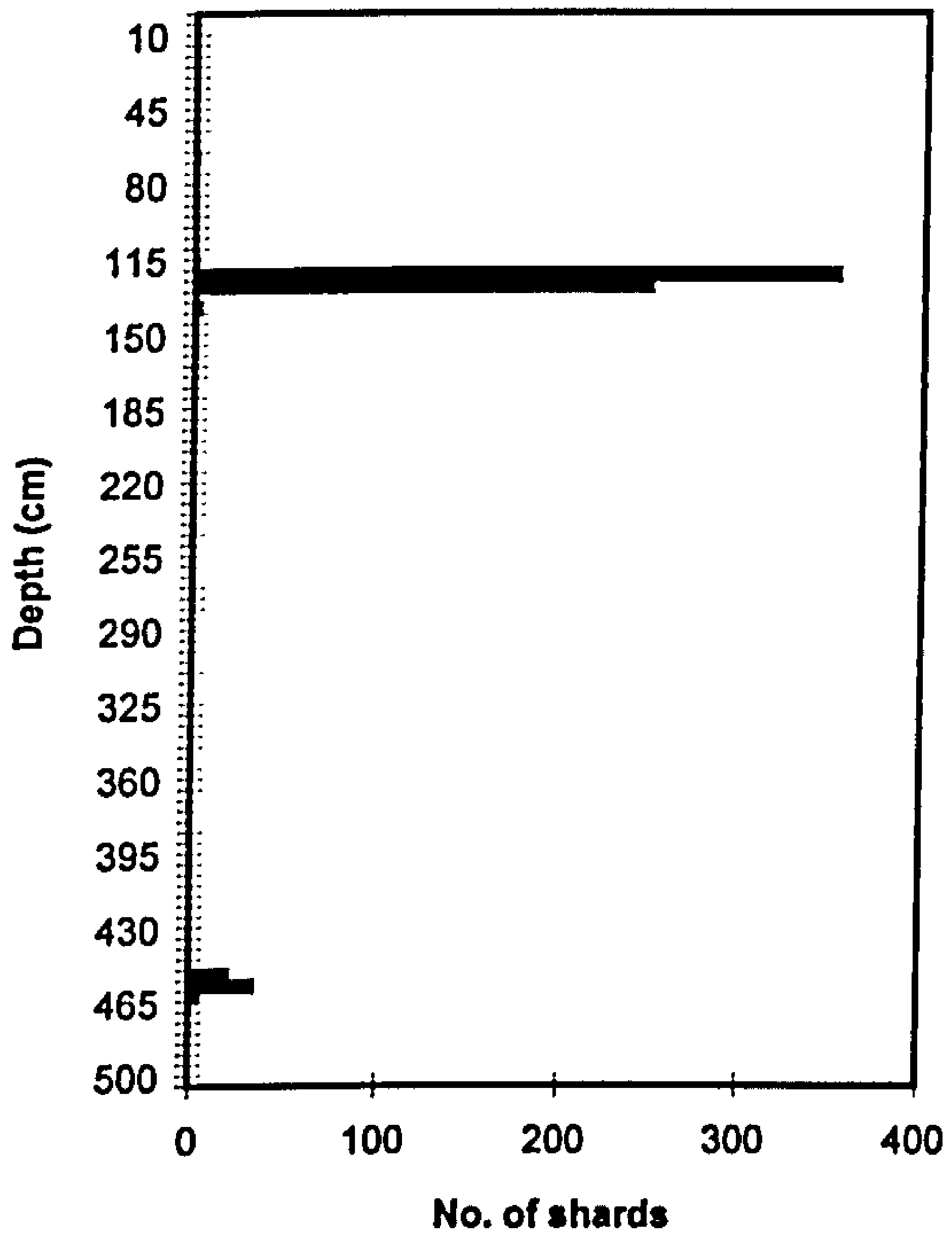


Figure 5.1f Tephra shard abundance at Langlands Moss

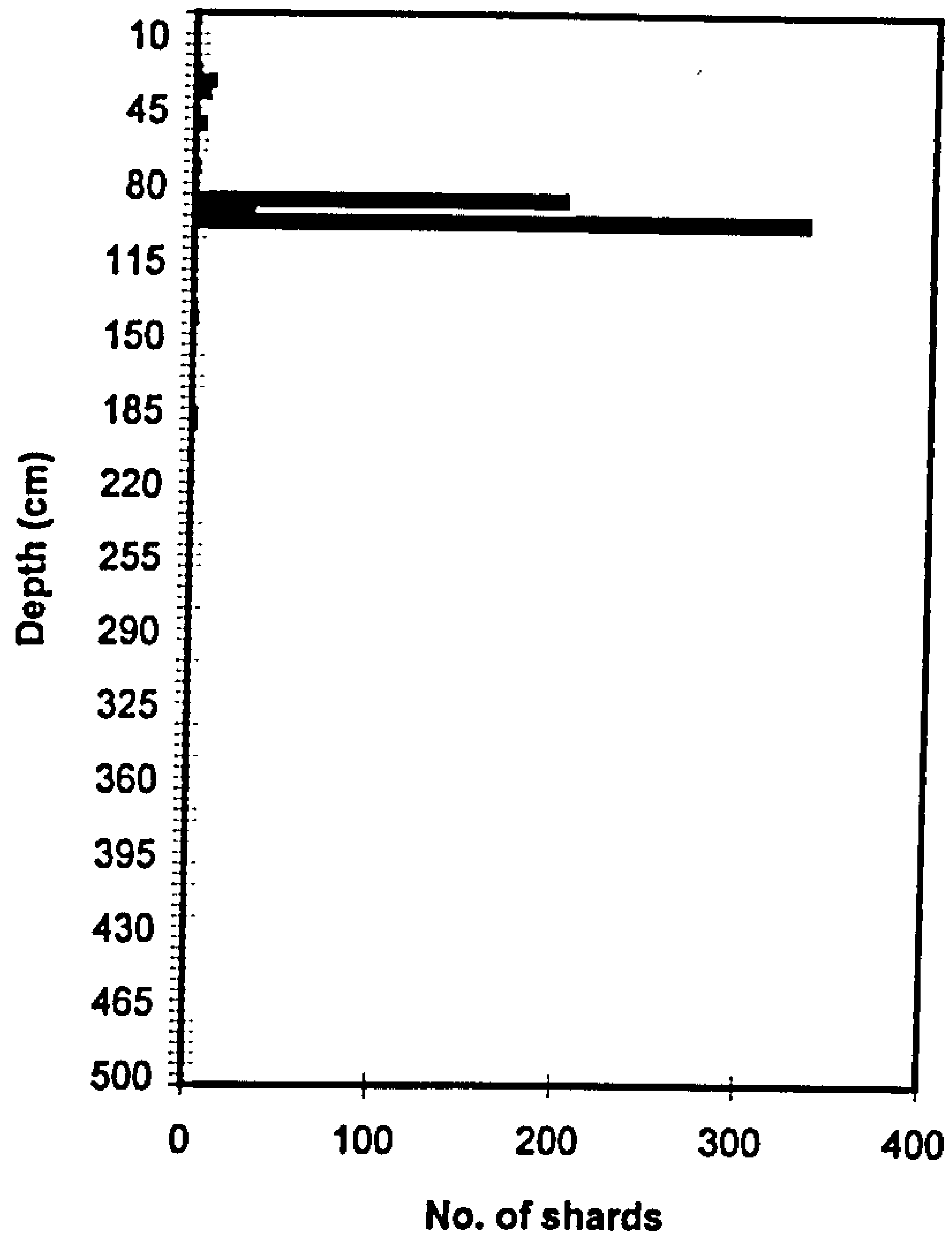


Figure 5.1g Tephra shard abundance at Dalmellington Moss

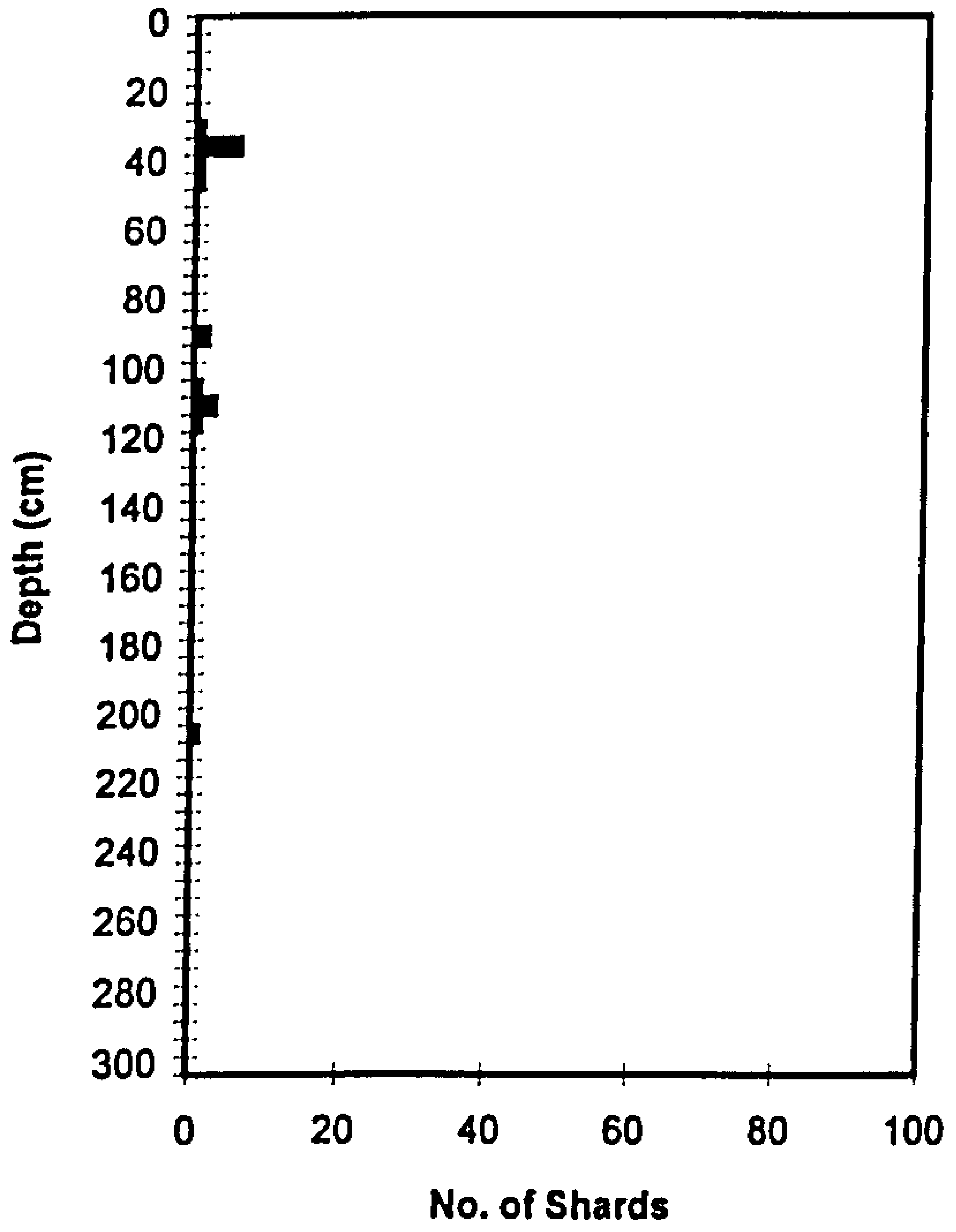


Figure 5.1h Tephra shard abundance at Red Moss of Balerno

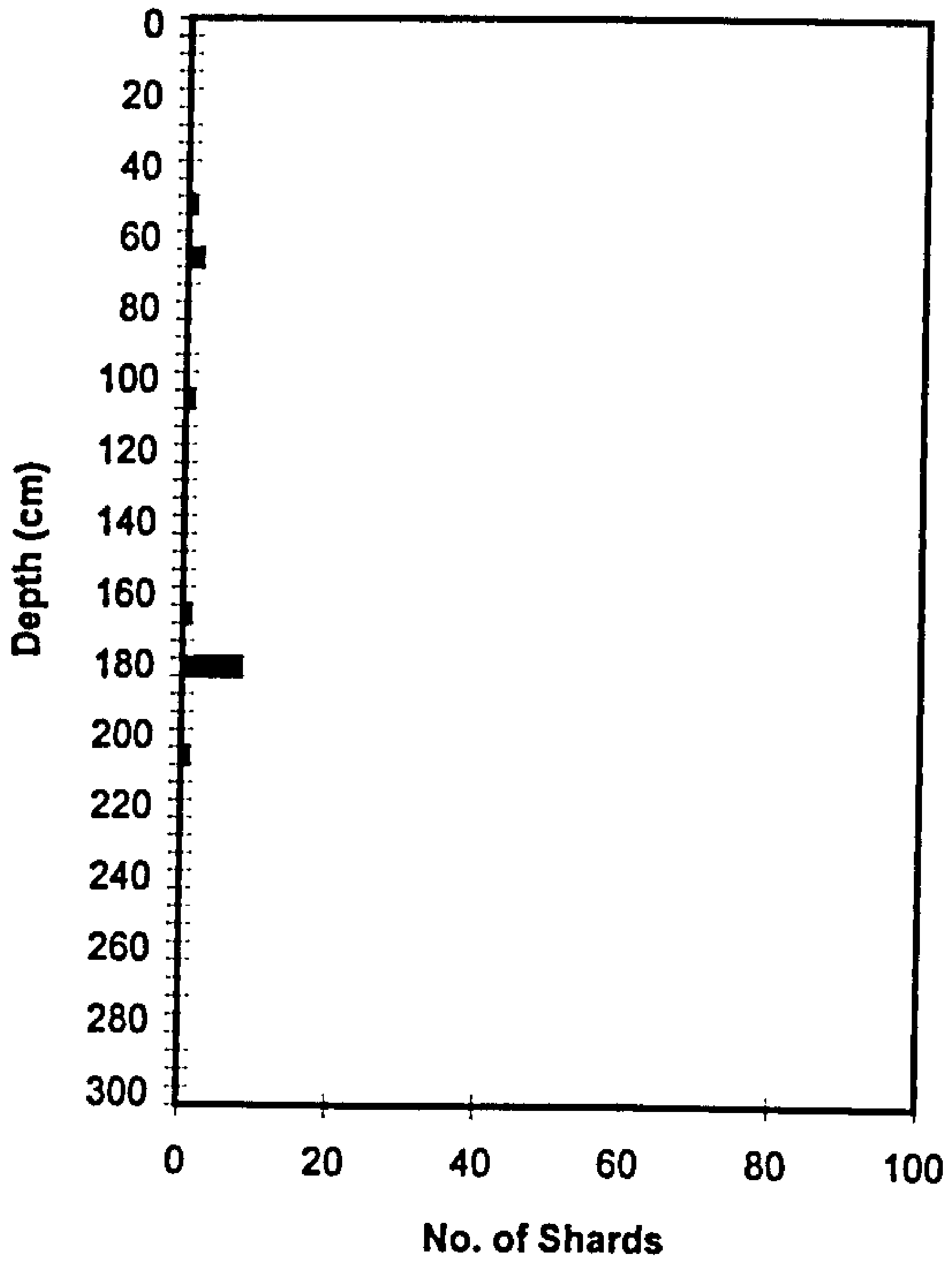


Figure 5.1i Tephra shard abundance at Mallachie Moss

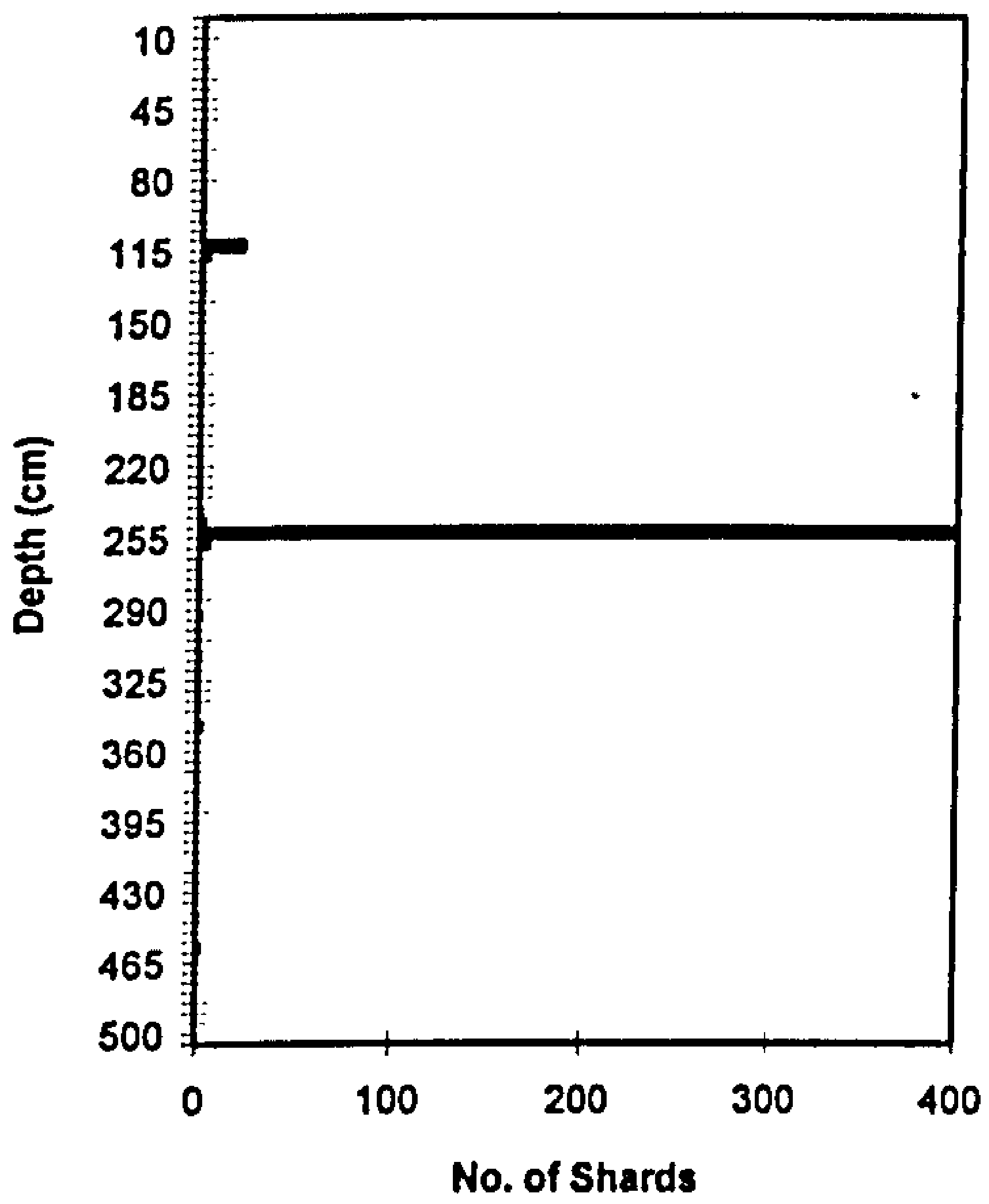


Figure 5.1j Tephra shard abundance at Blar-nam-Fiadh Moss

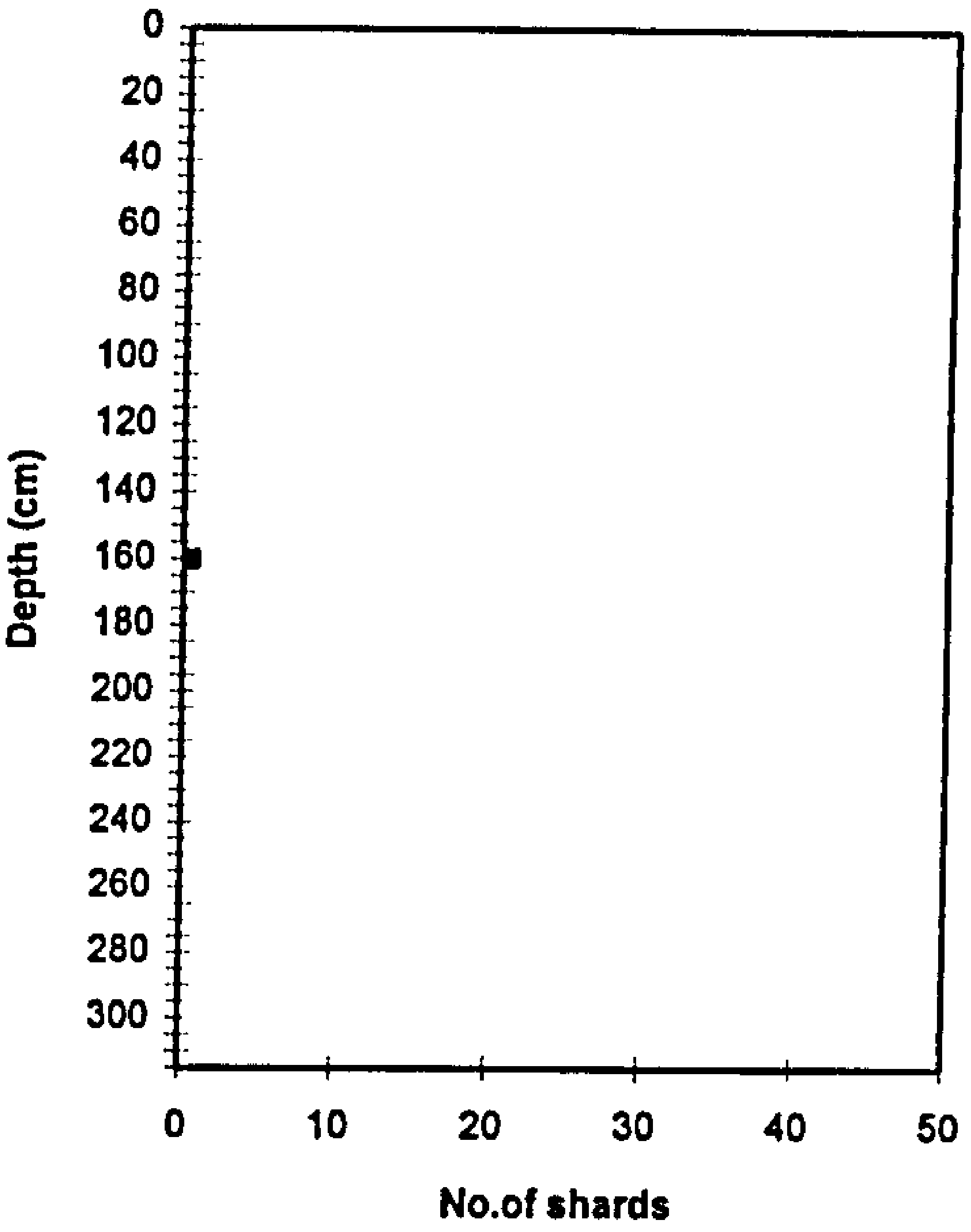


Figure 5.1k Tephra shard abundance at Ben Gorm Moss

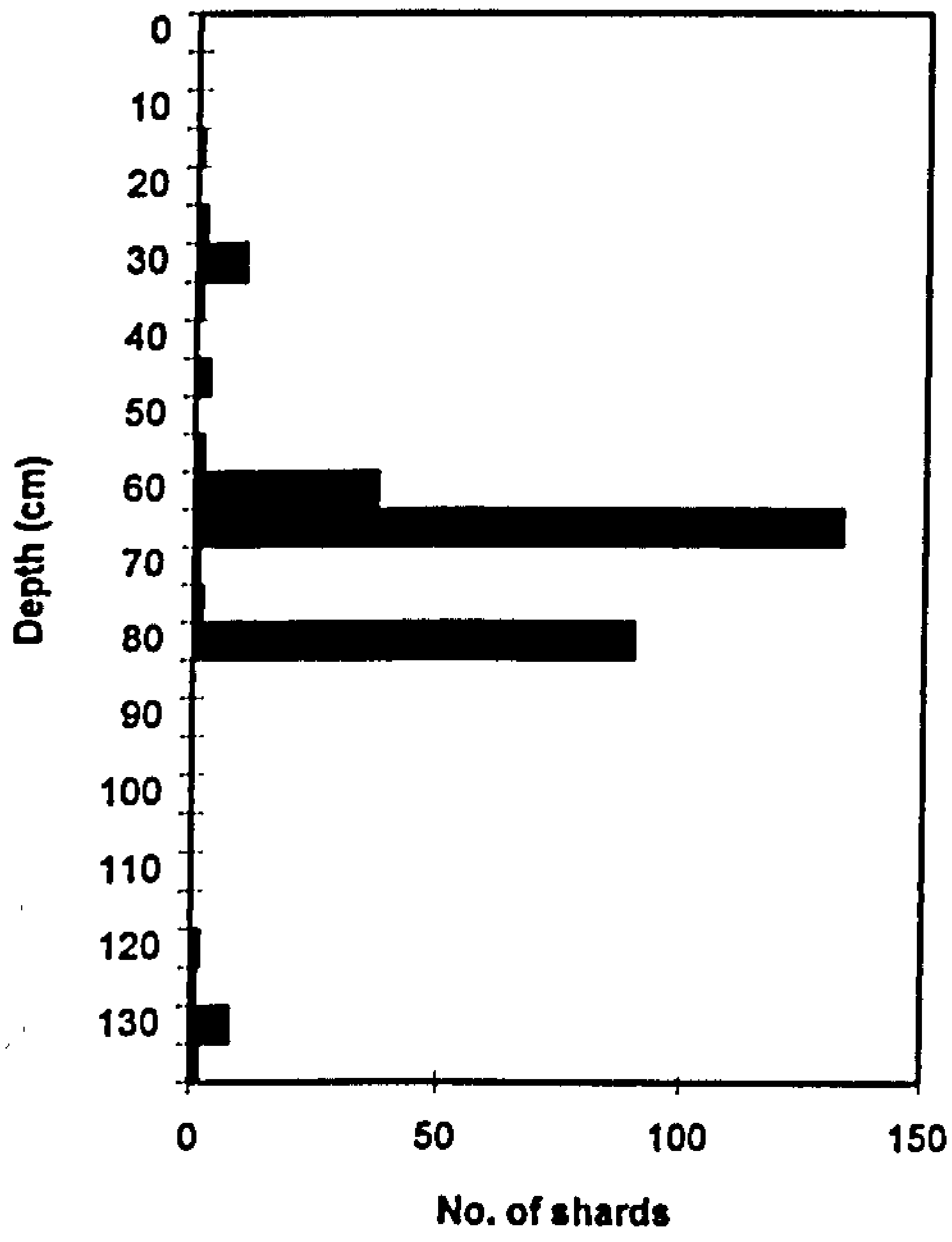
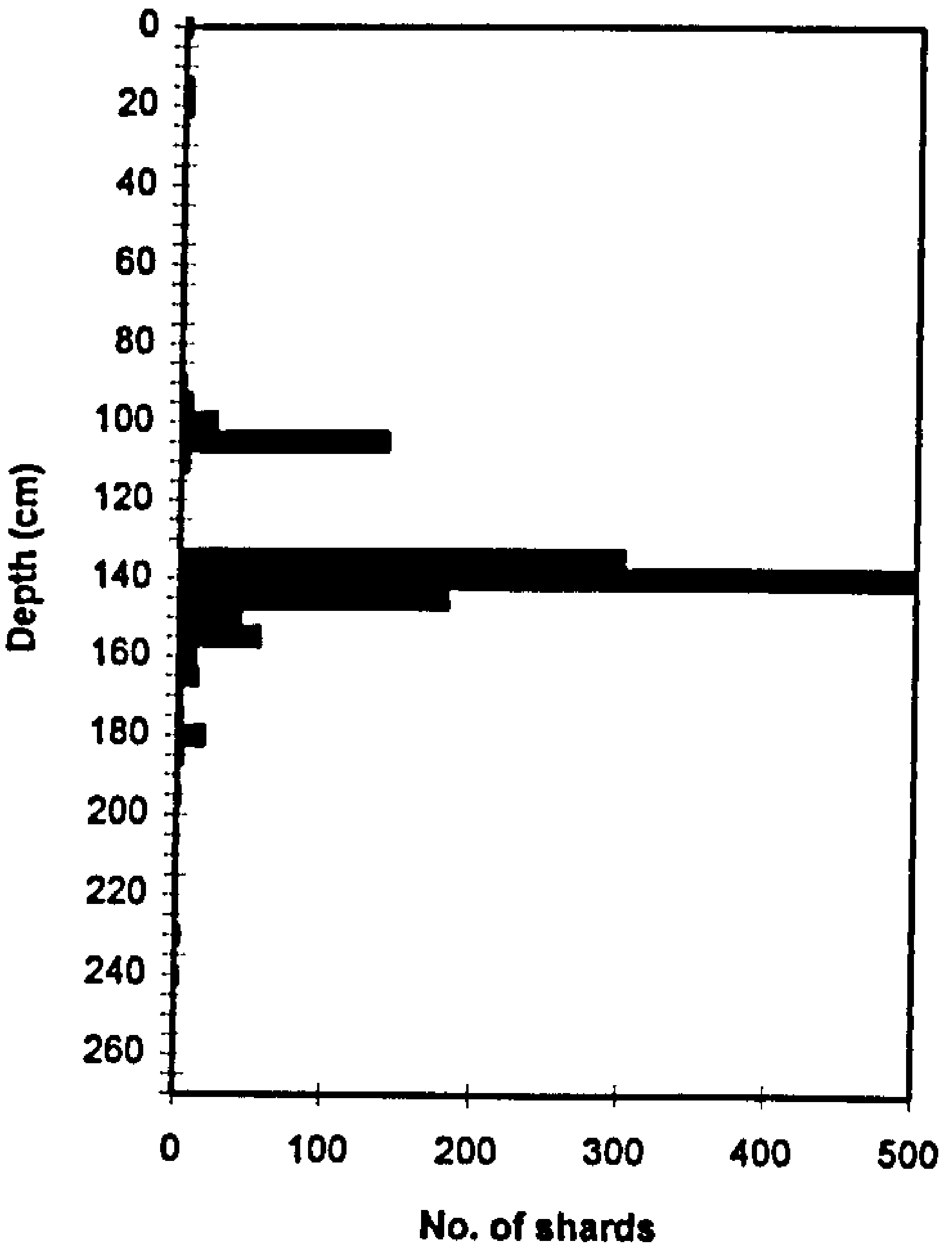


Figure 5.1l Tephra shard abundance at Craigmaud Moss



of shards at these levels does not imply that distinct tephra horizons are not present at the site in question, merely that they are not present in abundance from the profile taken. Recent work across a transect of profiles from Temple Hill Moss suggests that the spatial abundance of shards does vary considerably (Craggs, 1998; Langdon, unpublished data).

Figures 5.1(a-l) show at least two distinct tephra isochrones within the seven sites selected for palaeoecological analysis. Geochemical analysis was undertaken on each of these isochrones.

SEM micrographs of individual shards can be seen in Plates 5.1-5.6. The images are of shards which have been separated from the peat matrix by the acid digestion technique (Section 3.2.2.2). Plates 5.1-5.4 are of the Hekla-4 tephra at Mallachie Moss whereas Plates 5.5 and 5.6 are of the Glen Garry tephra at Ben Gorm Moss. The highly vesicular nature of the shards can be clearly seen, as well as the curved platelets of bubble walls (well defined in Plates 5.3 and 5.5). The shard depicted in Plate 5.6 appears highly weathered, a trend which was noted in the majority of shards from Ben Gorm Moss compared to other sites, and may thus be a consequence of the site location and prevailing climatic conditions.

5.2.2 Subsampling of tephra layers

In order to precisely correlate climatic changes between sites it is necessary to know if the maximum concentration of tephra shards within each layer is confined to peaks at centimetre resolution or less, or whether variations in peak concentrations occur. To test this, contiguous centimetre samples were taken from each tephra layer from the master core, and the results can be seen in Figures 5.2(a-n).

Figures 5.2(a-n) demonstrate that the peak tephra concentrations are almost entirely restricted to samples of 1cm thick, and at certain sites potentially less, as also found by Hall *et al.* (1994b). As the sampling intervals for plant macrofossil and humification analyses were every 4cm, there was no need to sample the tephra at the sub-centimetre scale. The maximum number of shards per gram is considered to represent the volcanic airfall event, as the other levels with relatively high concentrations of shards occur

Plate 5.1 SEM showing a tephra shard (Hekla-4) from Mallachie Moss.

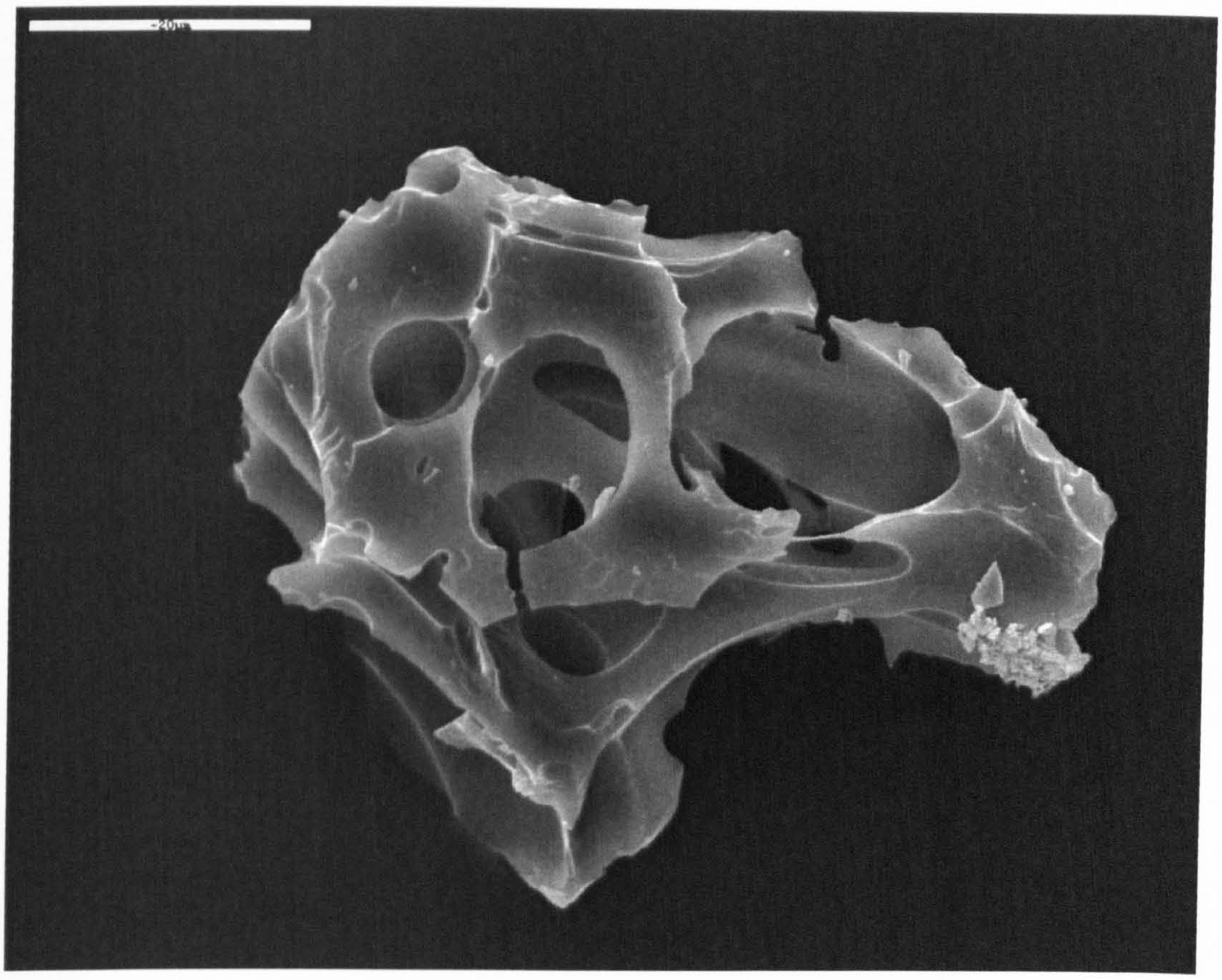


Plate 5.2 SEM showing a close up of the vesicles on a tephra shard (Hekla-4) from Mallachie Moss.

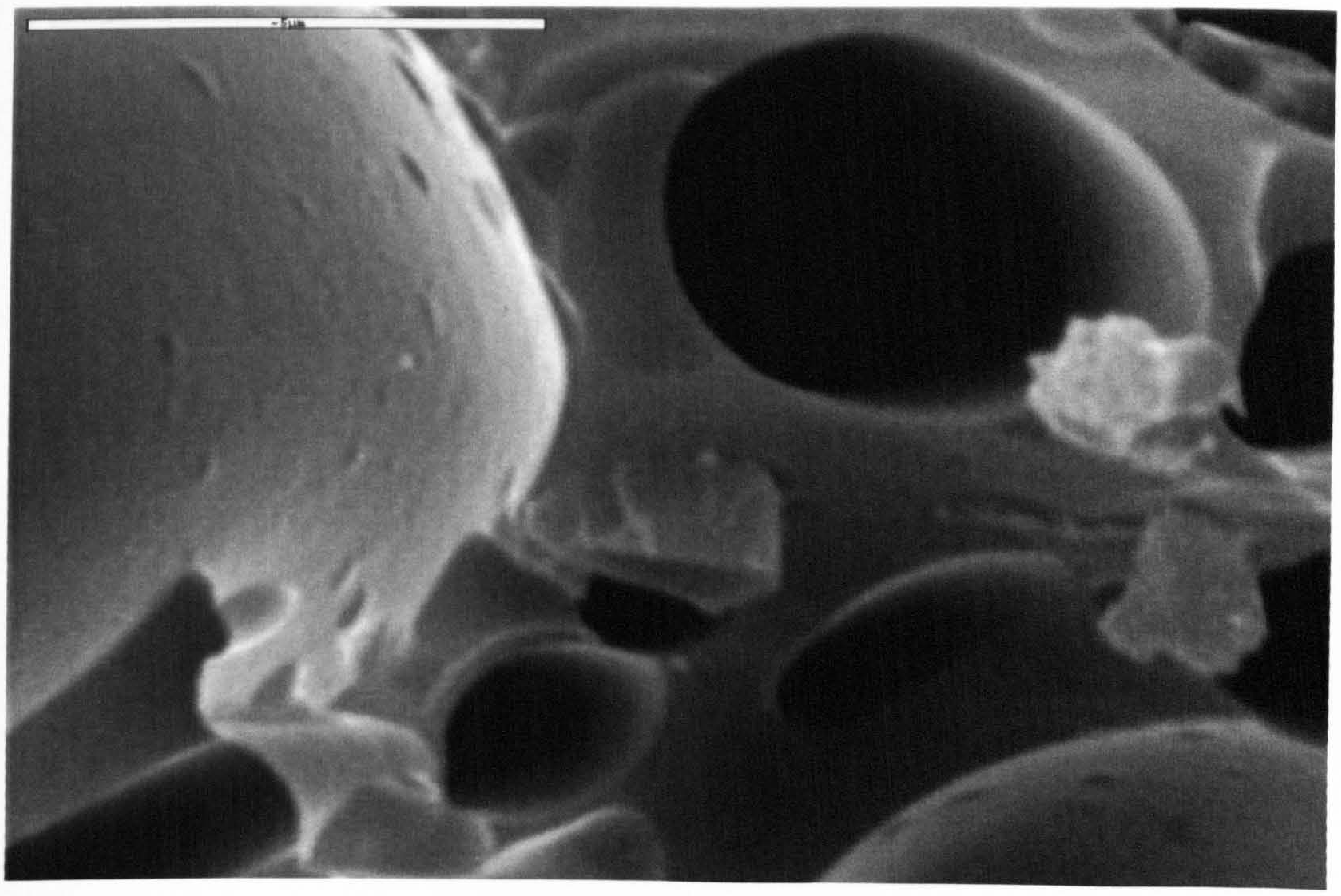


Plate 5.3 SEM showing a tephra shard (Hekla-4) from Mallachie Moss. The curved platelets of bubble walls can be clearly seen.

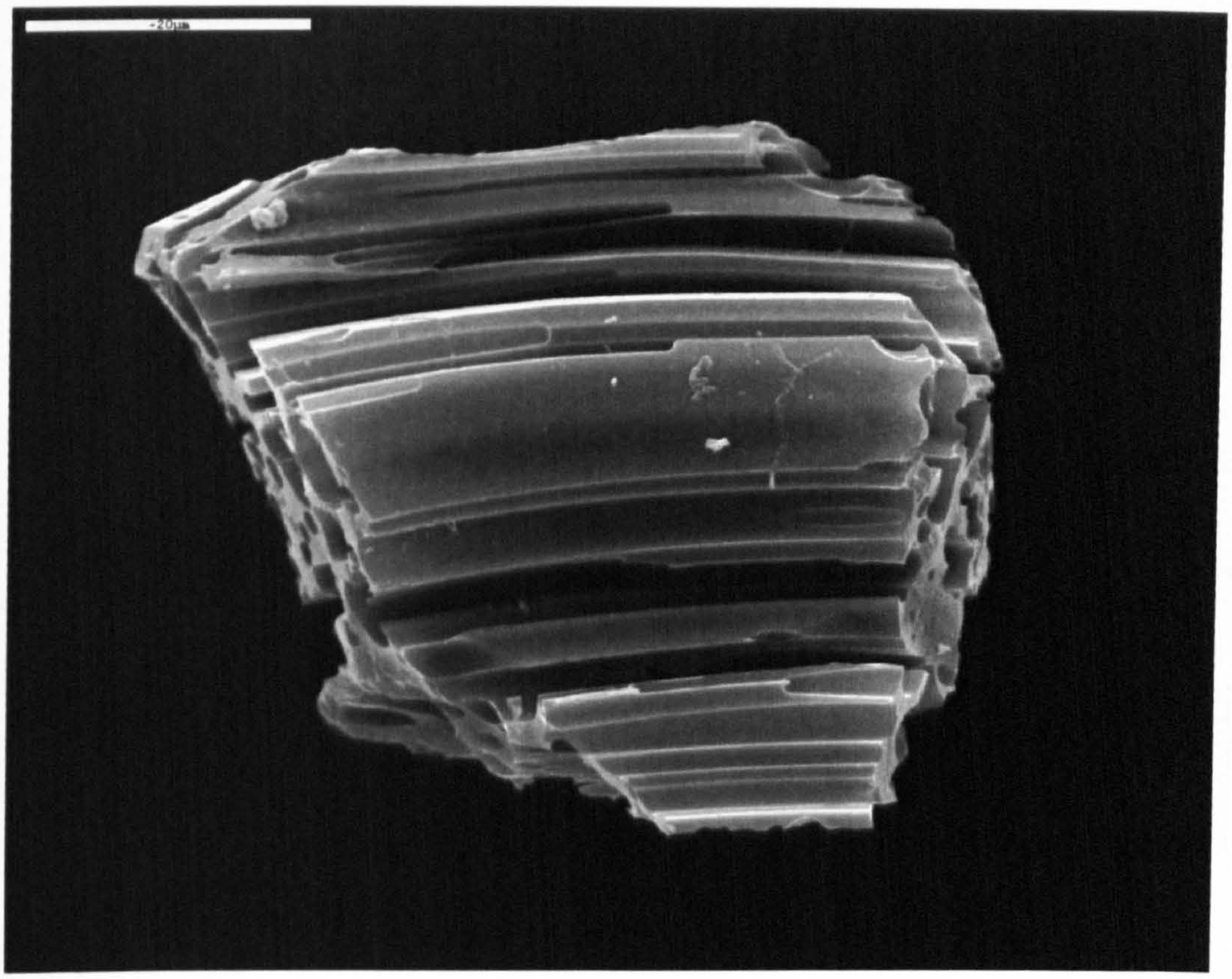


Plate 5.4 SEM showing a highly vesicular tephra shard (Hekla-4) from Mallachie Moss.

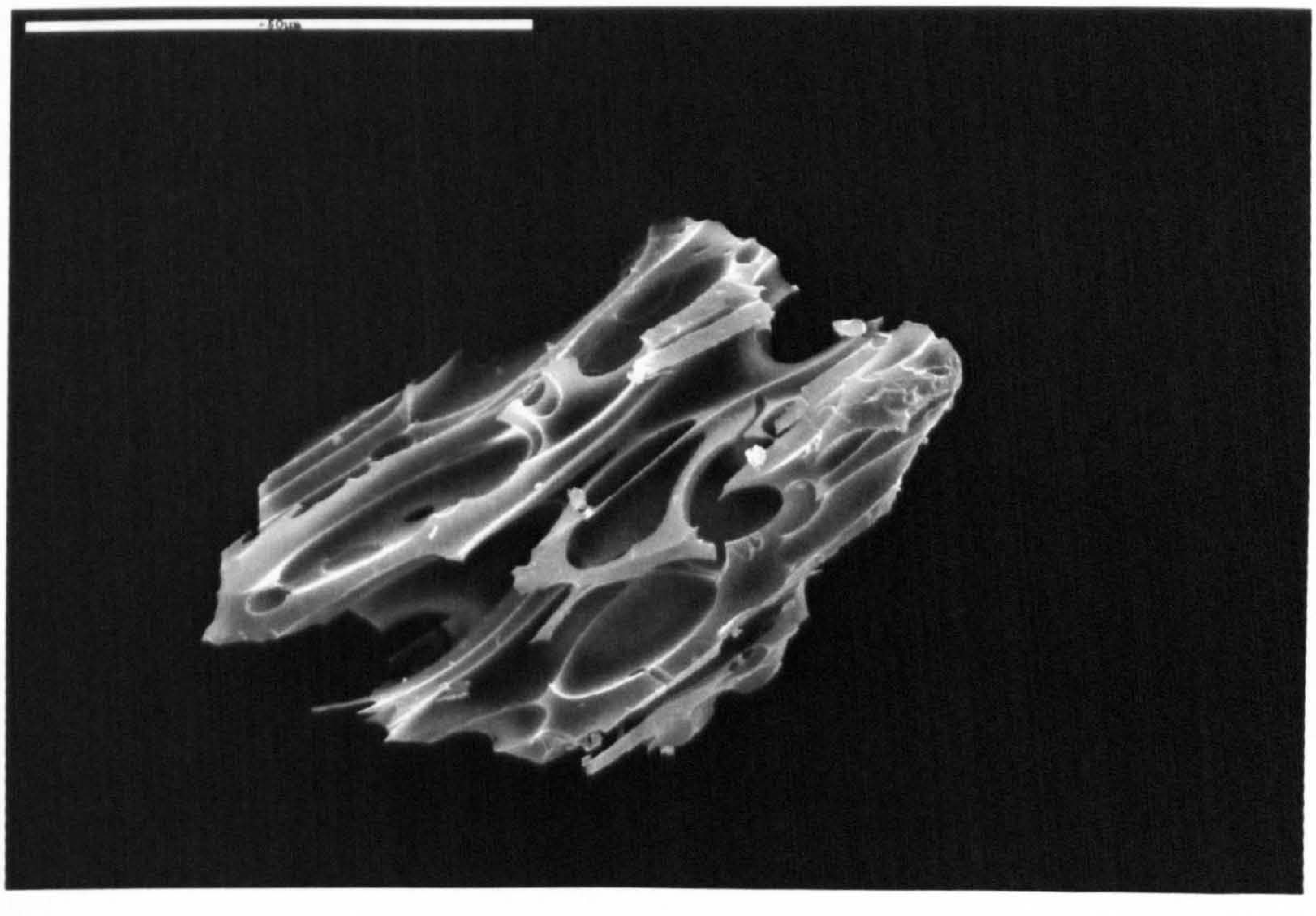


Plate 5.5 SEM showing curved platelets of bubble walls from a tephra shard (Glen Garry) from Ben Gorm Moss.

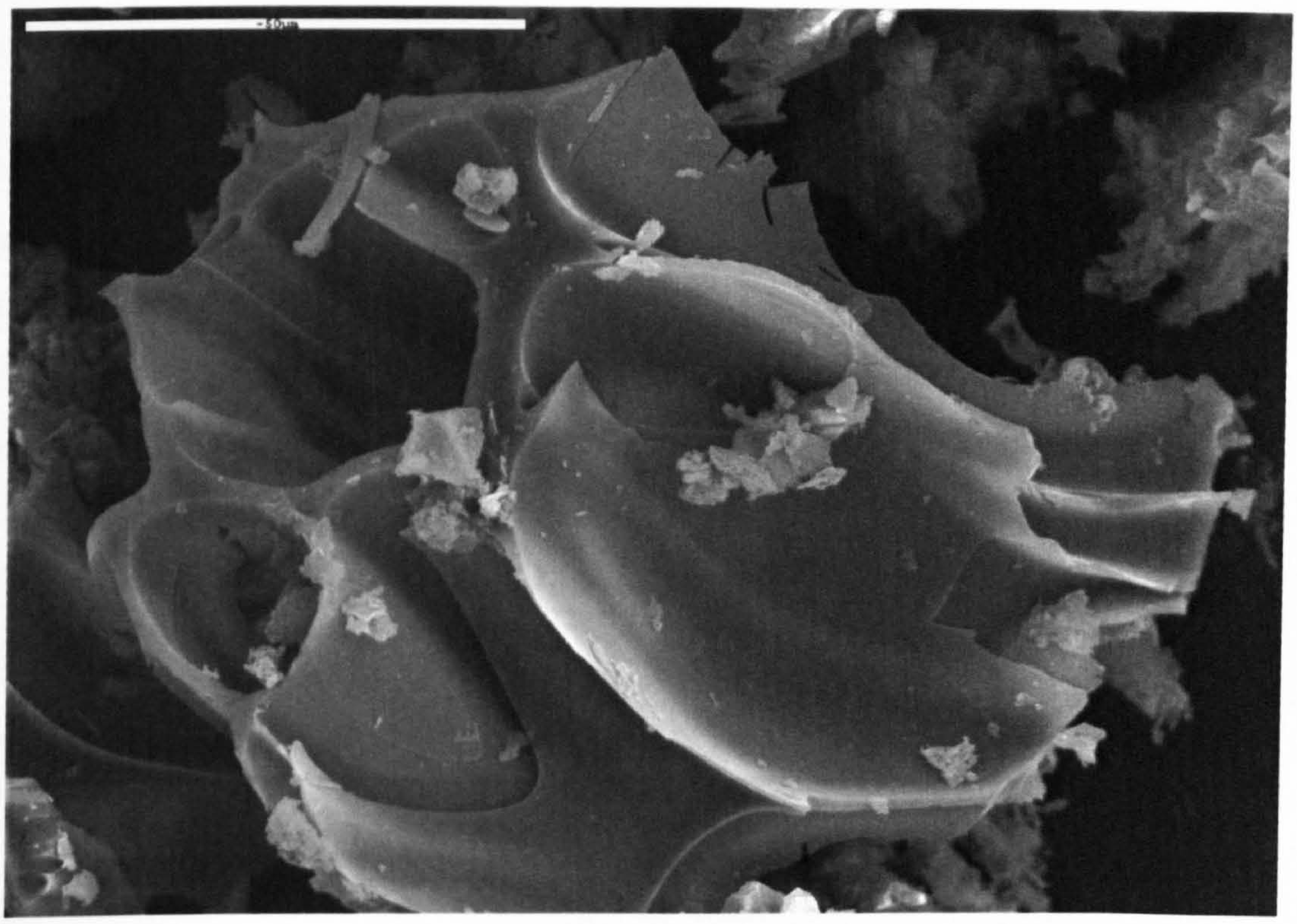
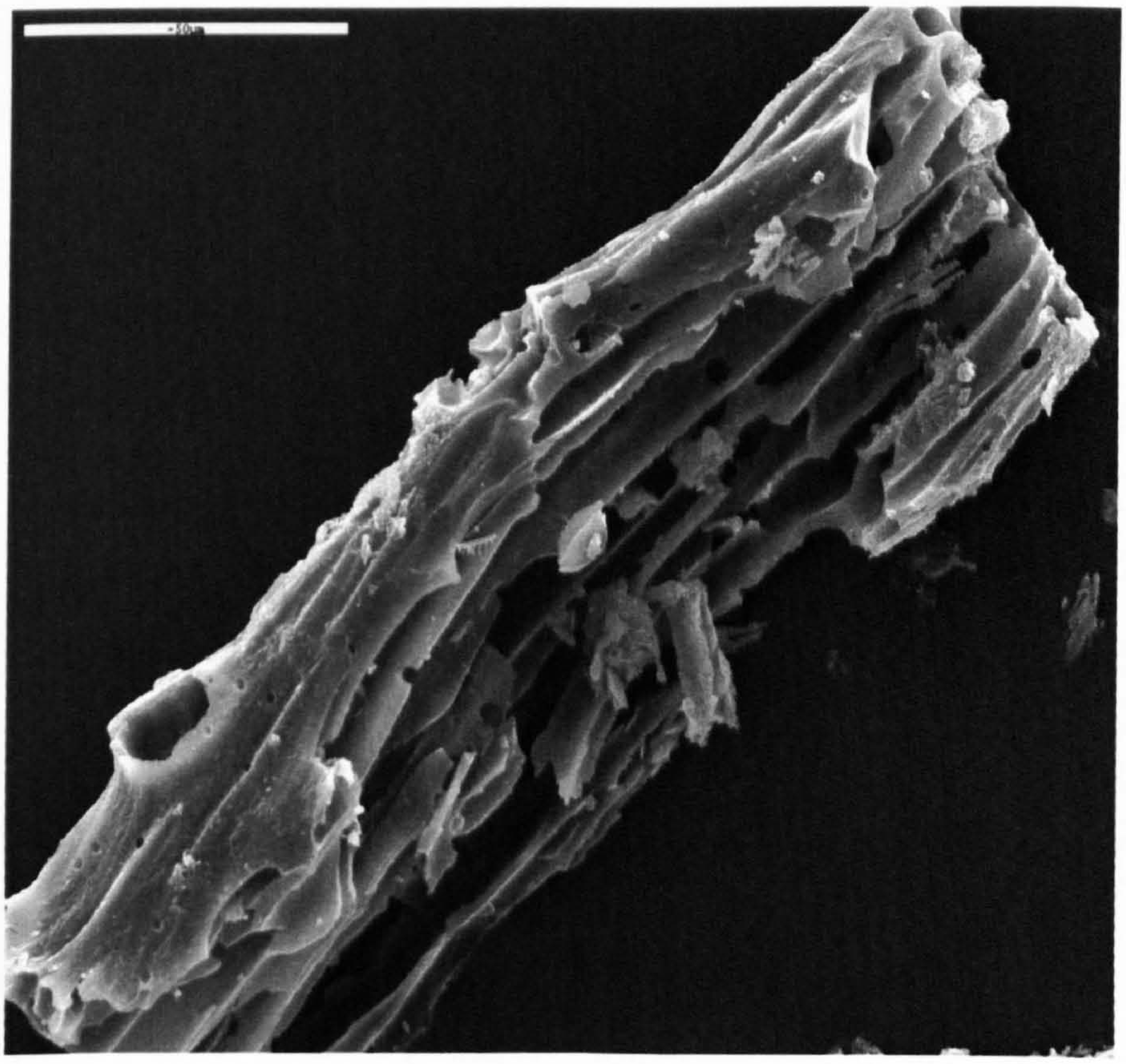


Plate 5.6 SEM showing a highly weathered tephra shard (Glen Garry) from Ben Gorm Moss.



stratigraphically beneath the peak concentration (Hall *et al.*, 1994b), probably due to redistribution by vascular plants, as documented above. The ability to locate the peak concentrations of tephra enables the palaeoclimatological records derived from these mires to be compared at precisely these depths.

By comparing the depths at which the tephras are located from Figures 5.1 and 5.2, differences can be noted of up to 15cm stratigraphically between the master and prospecting cores from most sites, with a difference of *ca.* 30cm for Shirgarton Moss. Care was taken to sample the two cores as near to each other as possible, so it must be noted that these stratigraphic differences can occur on small spatial scales. The stratigraphic differences are interpreted as being a function of mire surface topography at the time of aerial deposition.

5.2.3 Geochemical results

5.2.3.1 Introduction

Scottish palaeoenvironmental records covering the last 6000 years have to date yielded seven distinct tephra isochrones, which have been geochemically typed and dated (Dugmore *et al.*, 1995a and Dugmore *et al.*, 1995b). Other Icelandic tephras have also been located in Ireland, which may have also been deposited in Scotland, even if they have not yet been identified. The tephras found in Britain with published geochemistries are shown in Table 5.1. Other ash horizons which could conceivably be found in Scotland have been geochemically typed from Iceland (Larsen *et al.*, 1999), allowing potential correlations to be made with volcanic ash found in Scotland.

Figure 5.2a Distribution of Hekla-4 tephra in 1cm peat slices from Longbridge Moss

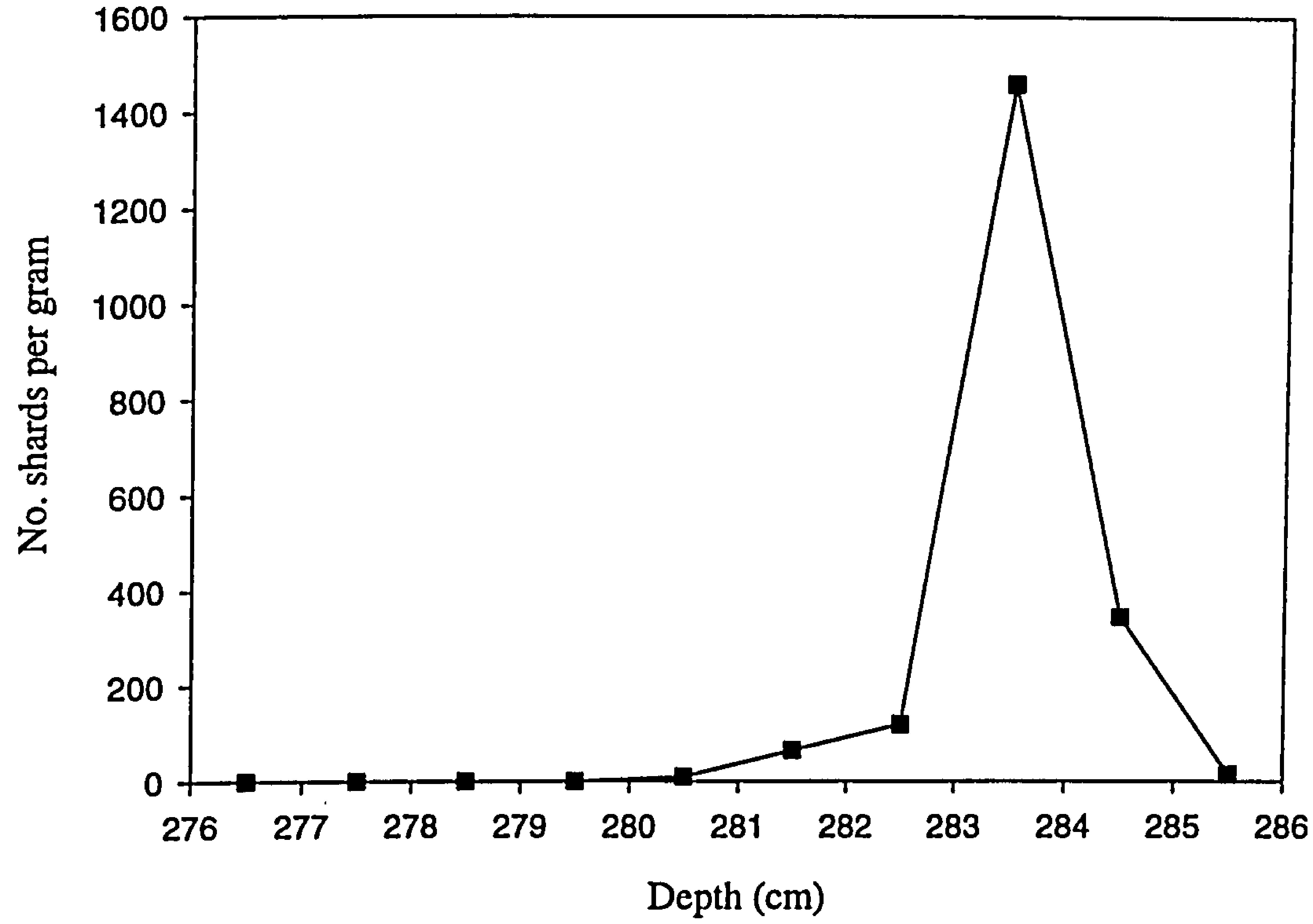


Figure 5.2b Distribution of Glen Garry tephra in 1cm peat slices from Shirgarton Moss

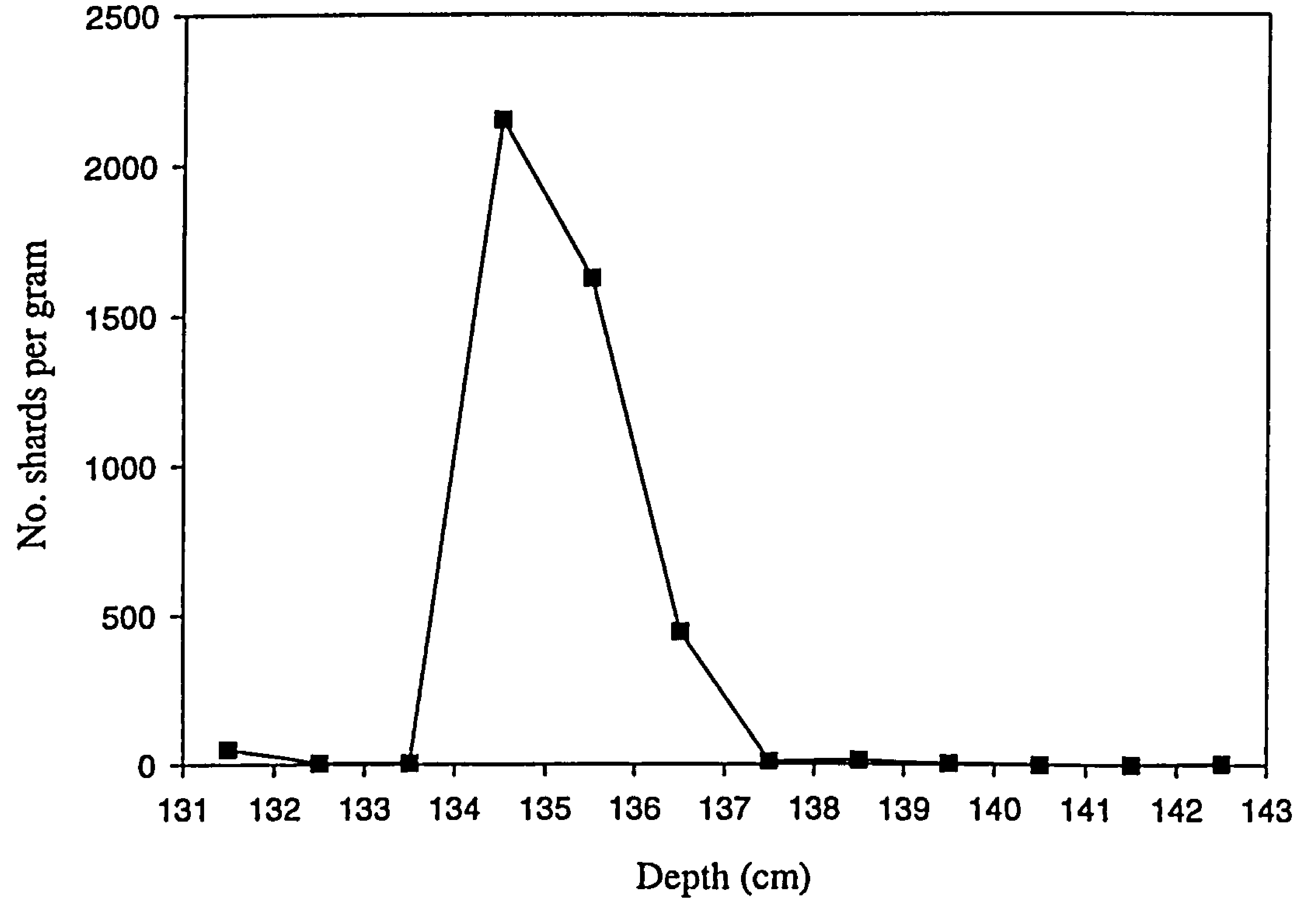


Figure 5.2c Distribution of AD 860 tephra in 1cm peat slices from Langlands Moss

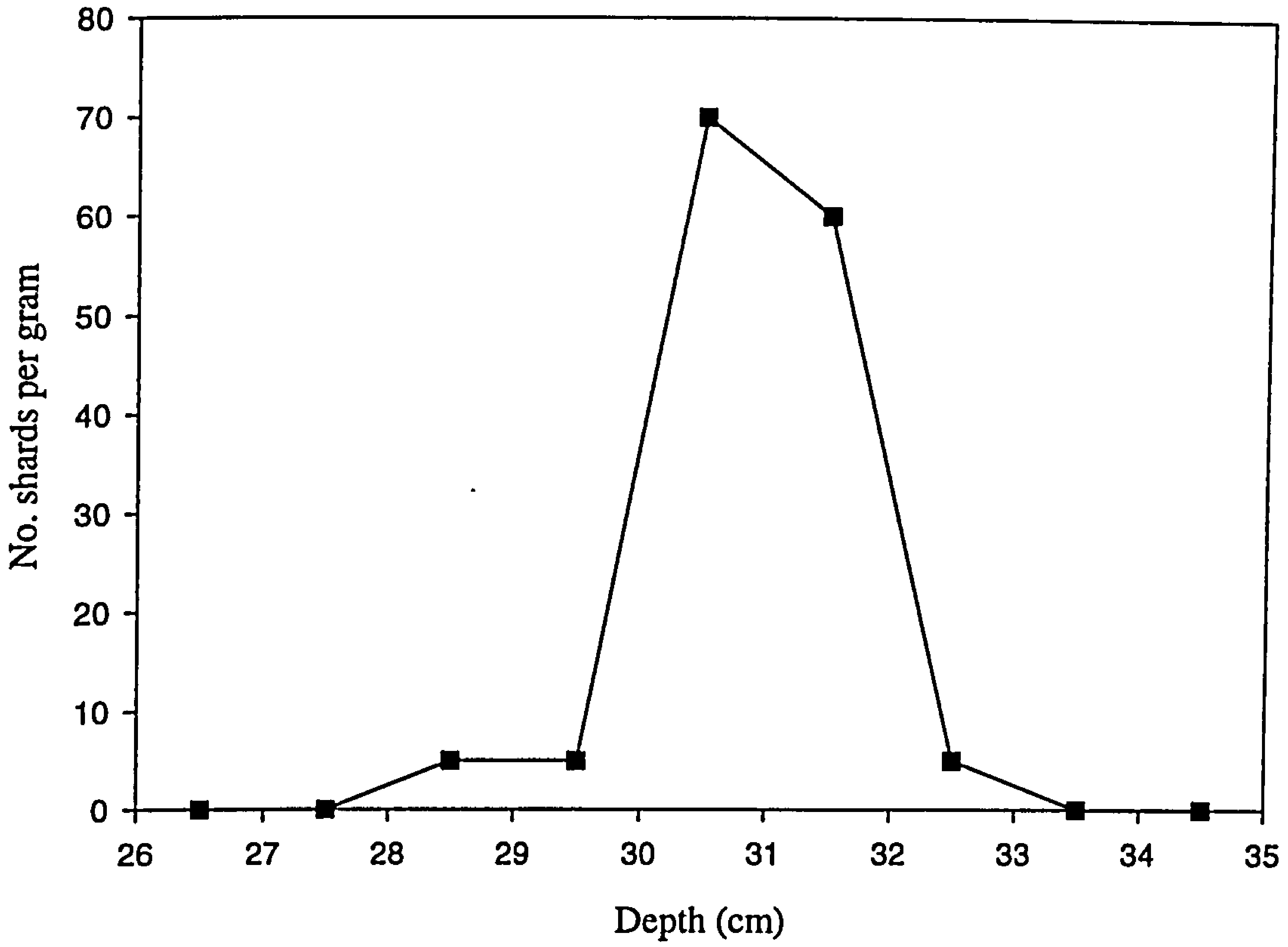


Figure 5.2d Distribution of Glen Garry tephra in 1cm peat slices from Langlands Moss

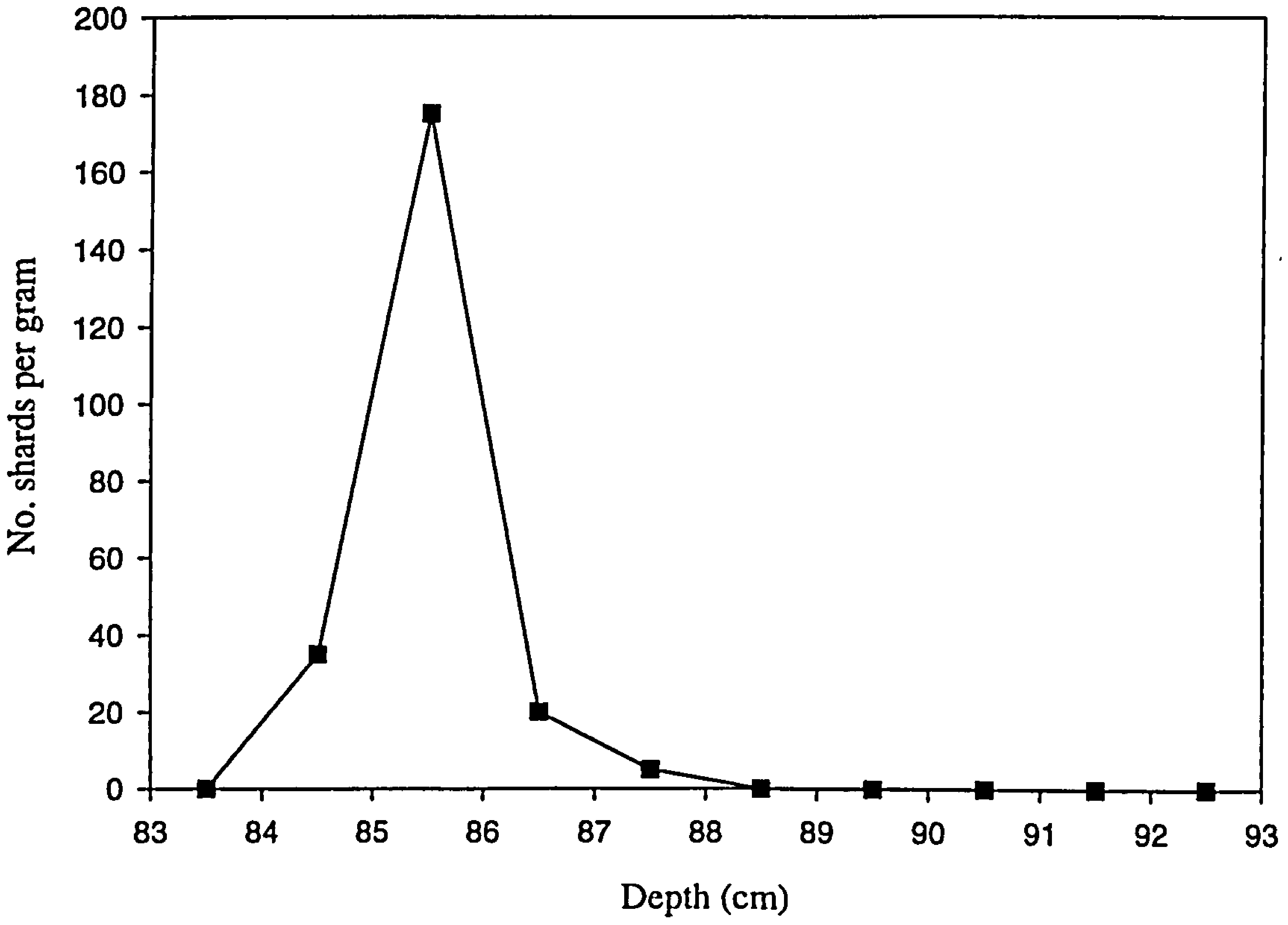


Figure 5.2e Distribution of Glen Garry tephra in 1cm peat slices from Temple Hill Moss

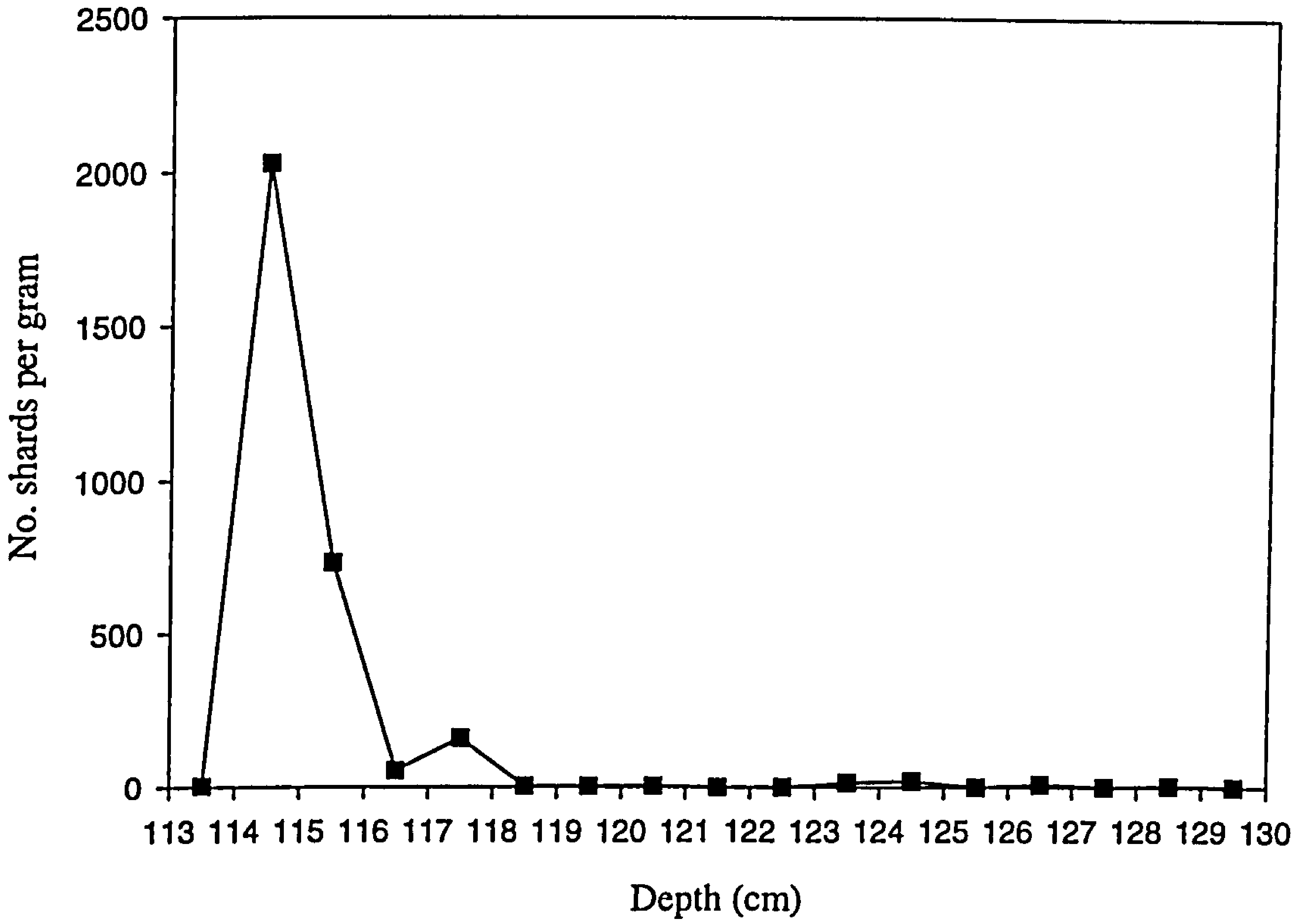


Figure 5.2f Distribution of Lairg layer tephra in 1cm peat slices from Temple Hill Moss

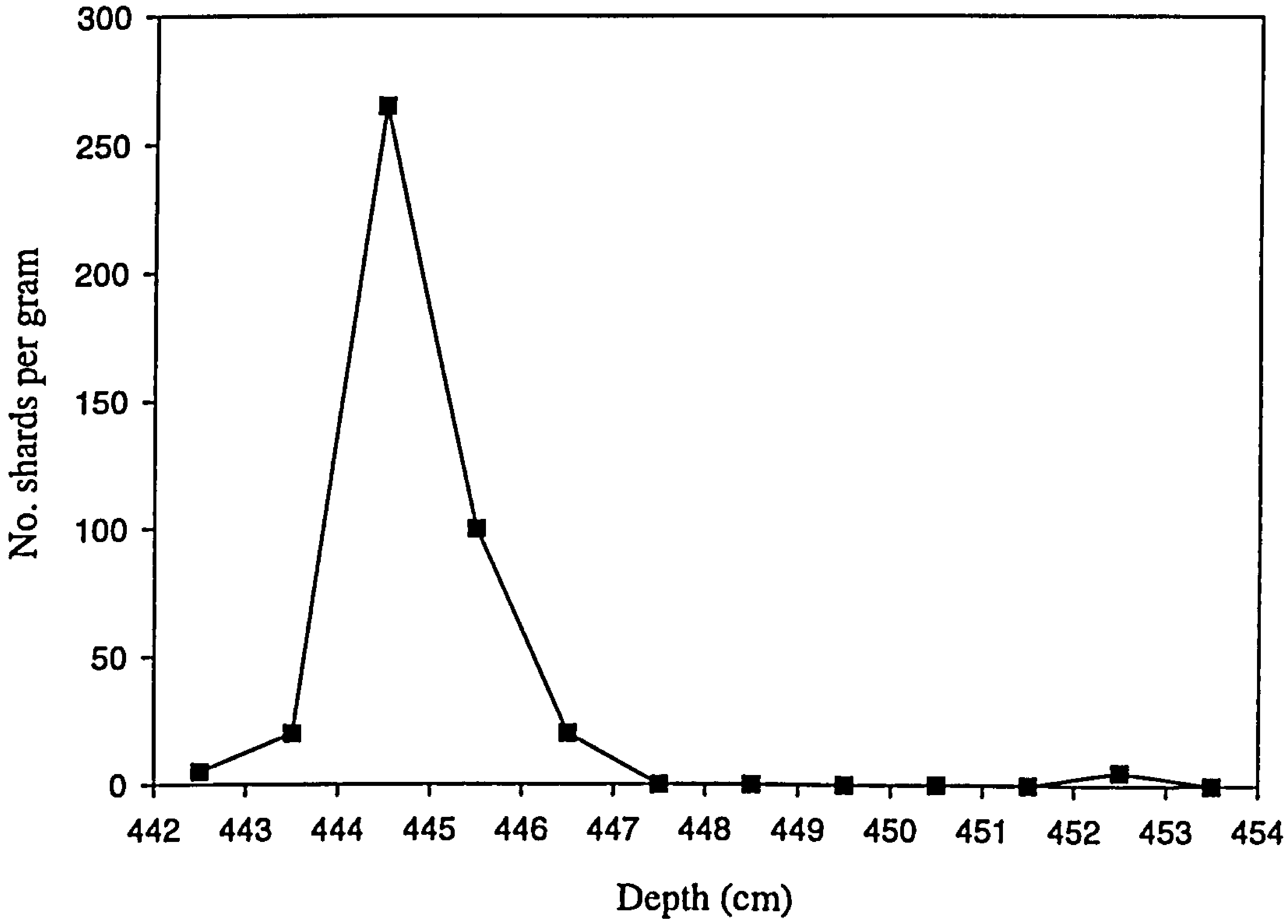


Figure 5.2g Distribution of Glen Garry tephra in 1cm peat slices from Mallachie Moss

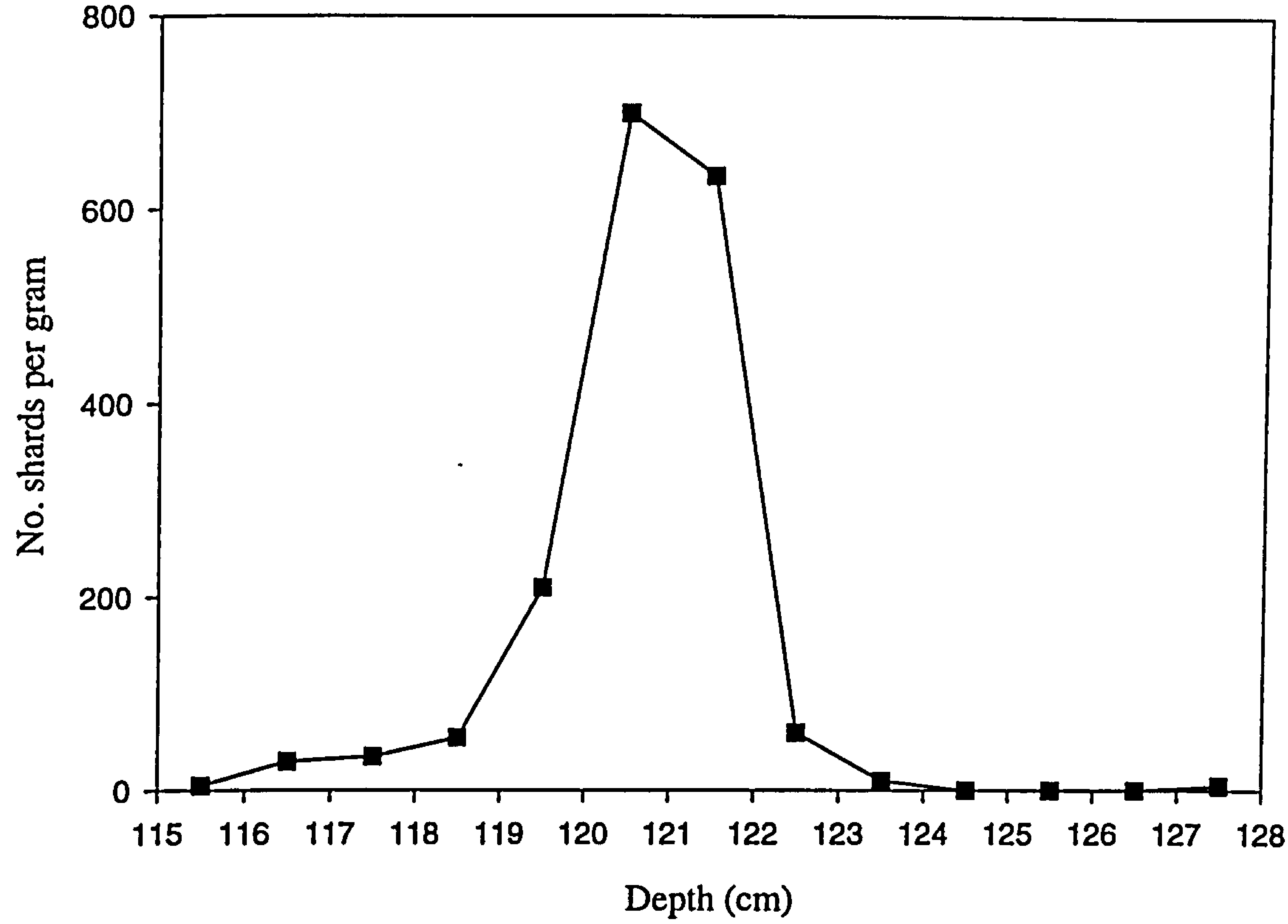


Figure 5.2h Distribution of Hekla-4 tephra in 1cm peat slices from Mallachie Moss

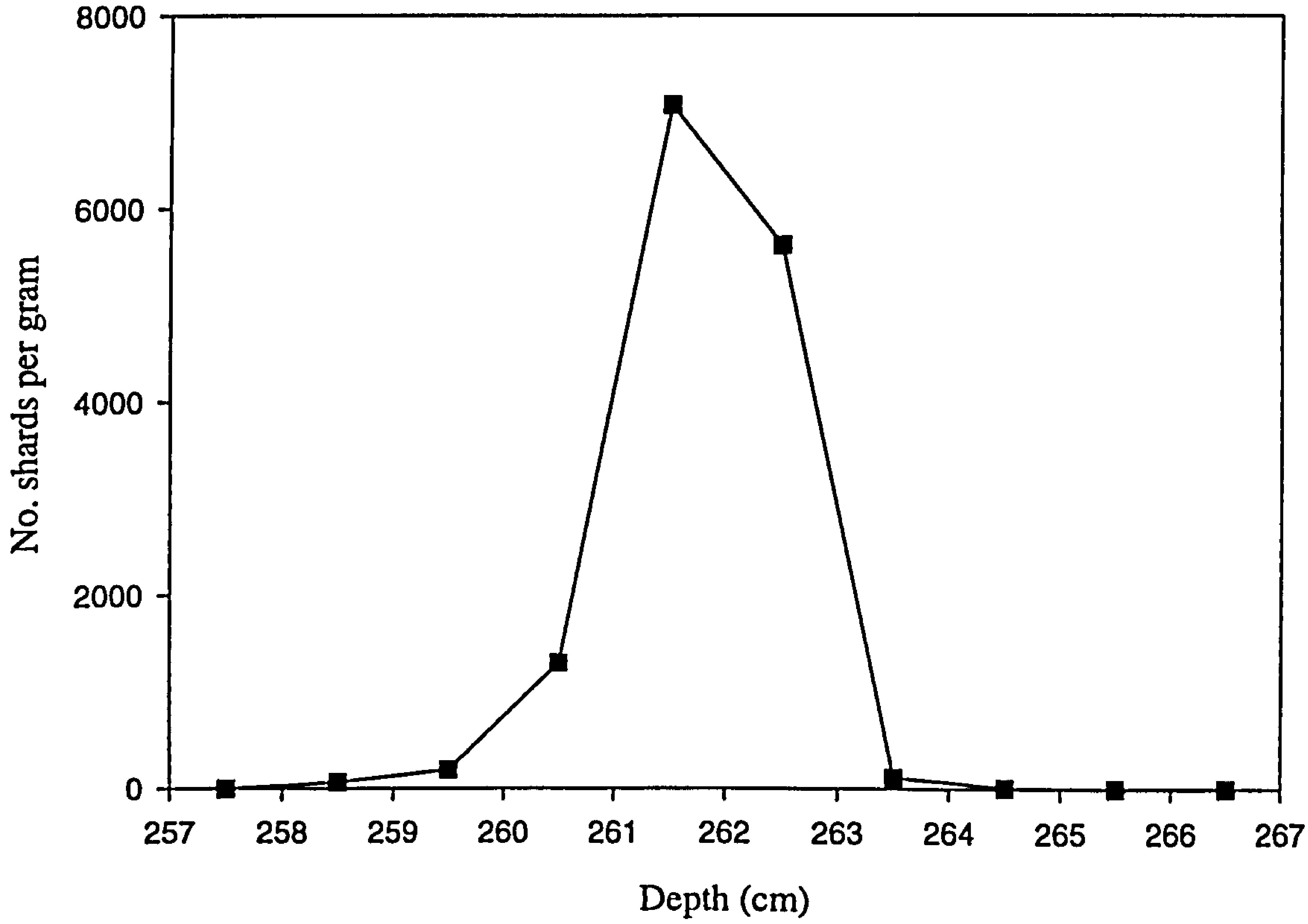


Figure 5.2i Distribution of Glen Garry tephra in 1cm peat slices from Craigmaud Moss

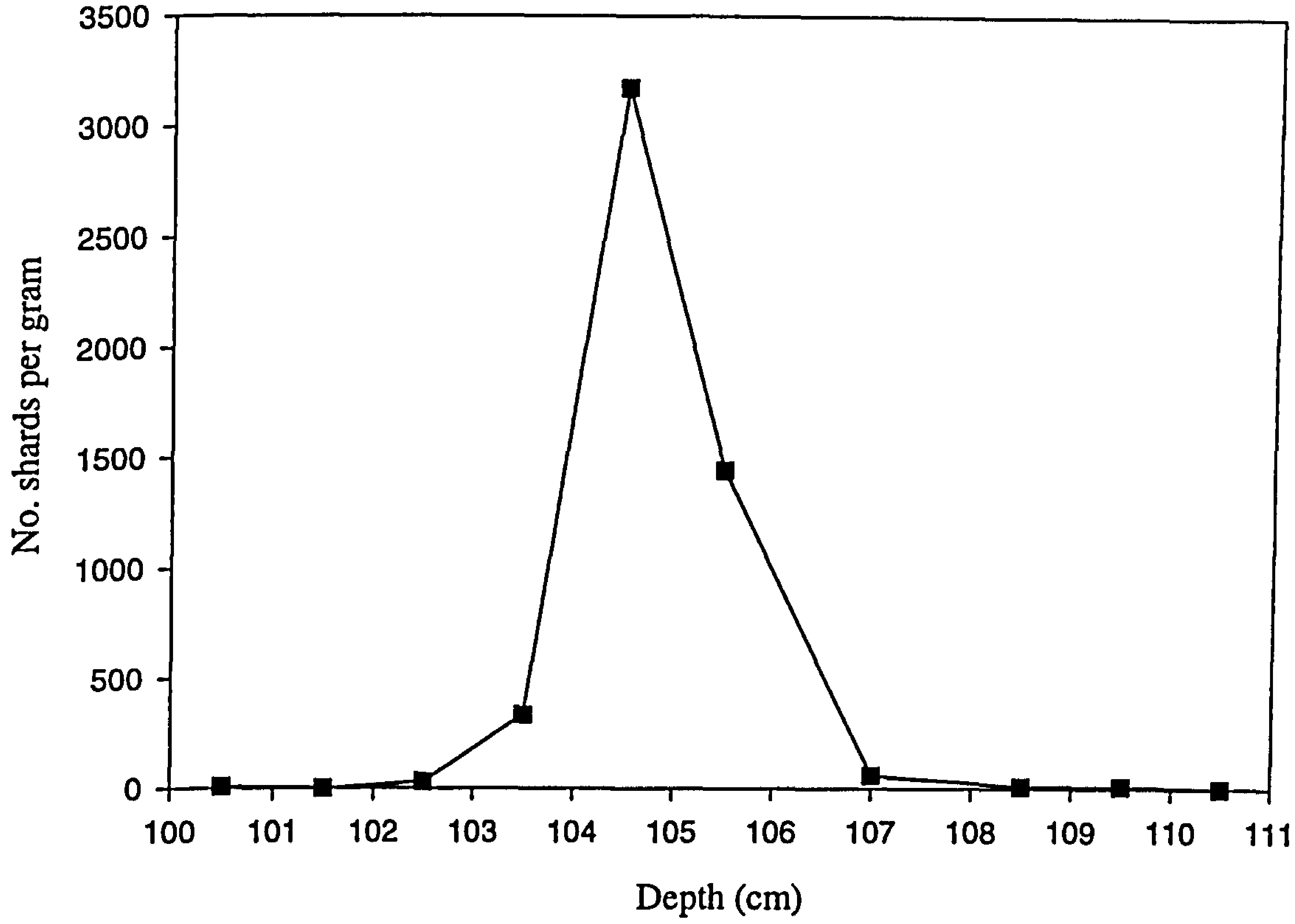


Figure 5.2j Distribution of Hekla-4 tephra in 1cm peat slices from Craigmaud Moss

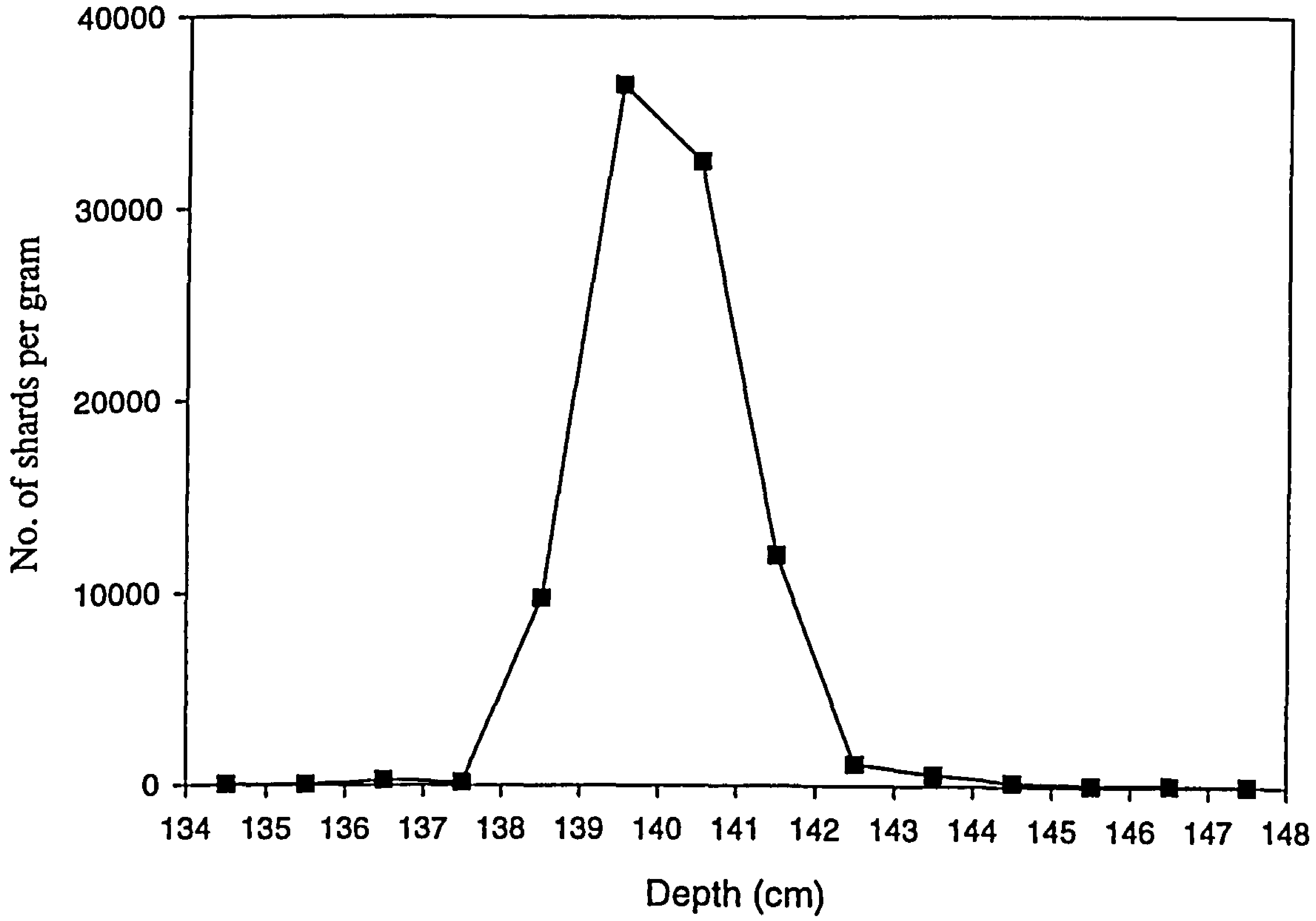


Figure 5.2k Distribution of Beinn Eighe(?) tephra in 1cm peat slices from Ben Gorm Moss

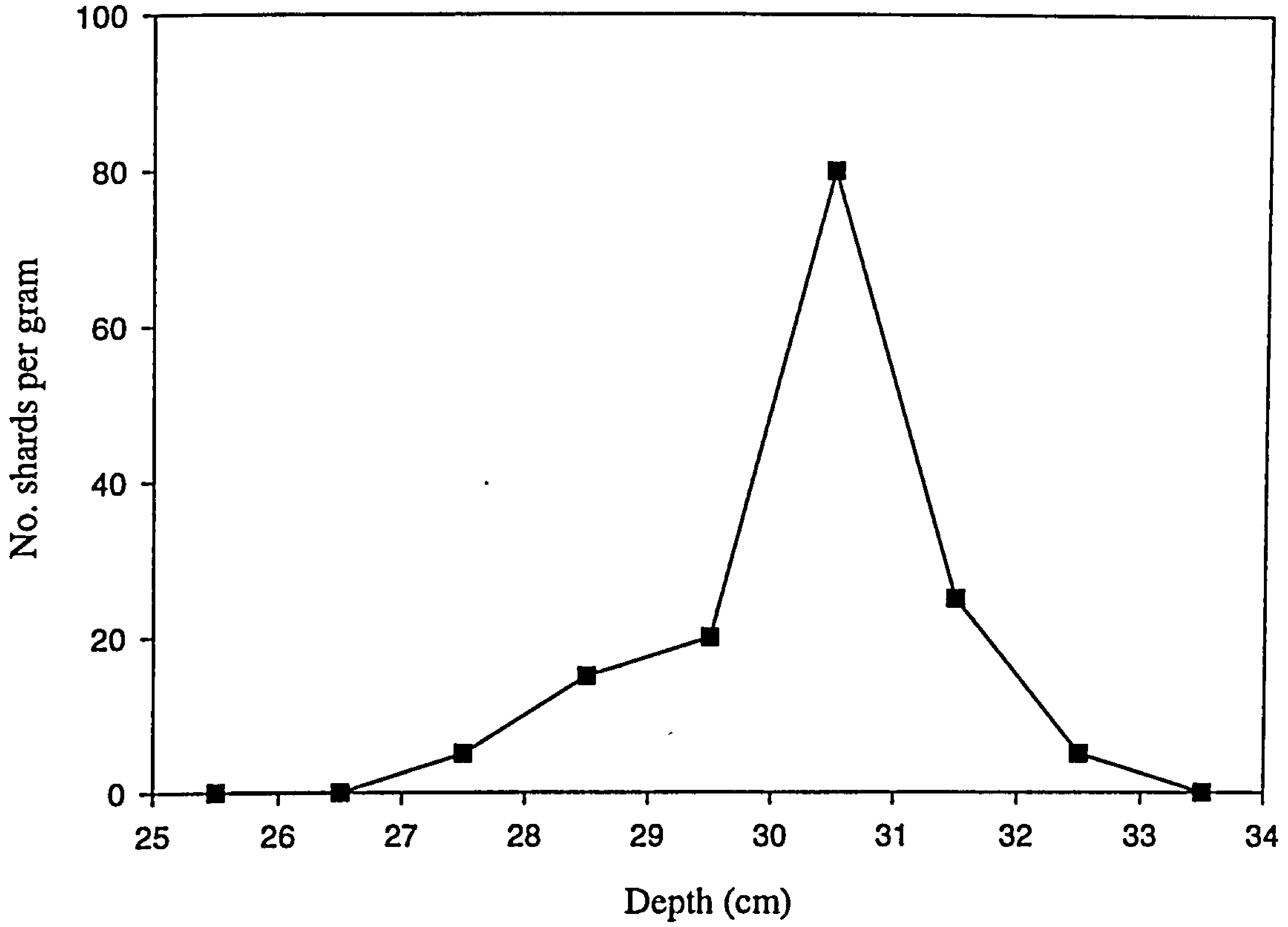


Figure 5.2l Distribution of Glen Garry tephra in 1cm peat slices from Ben Gorm Moss

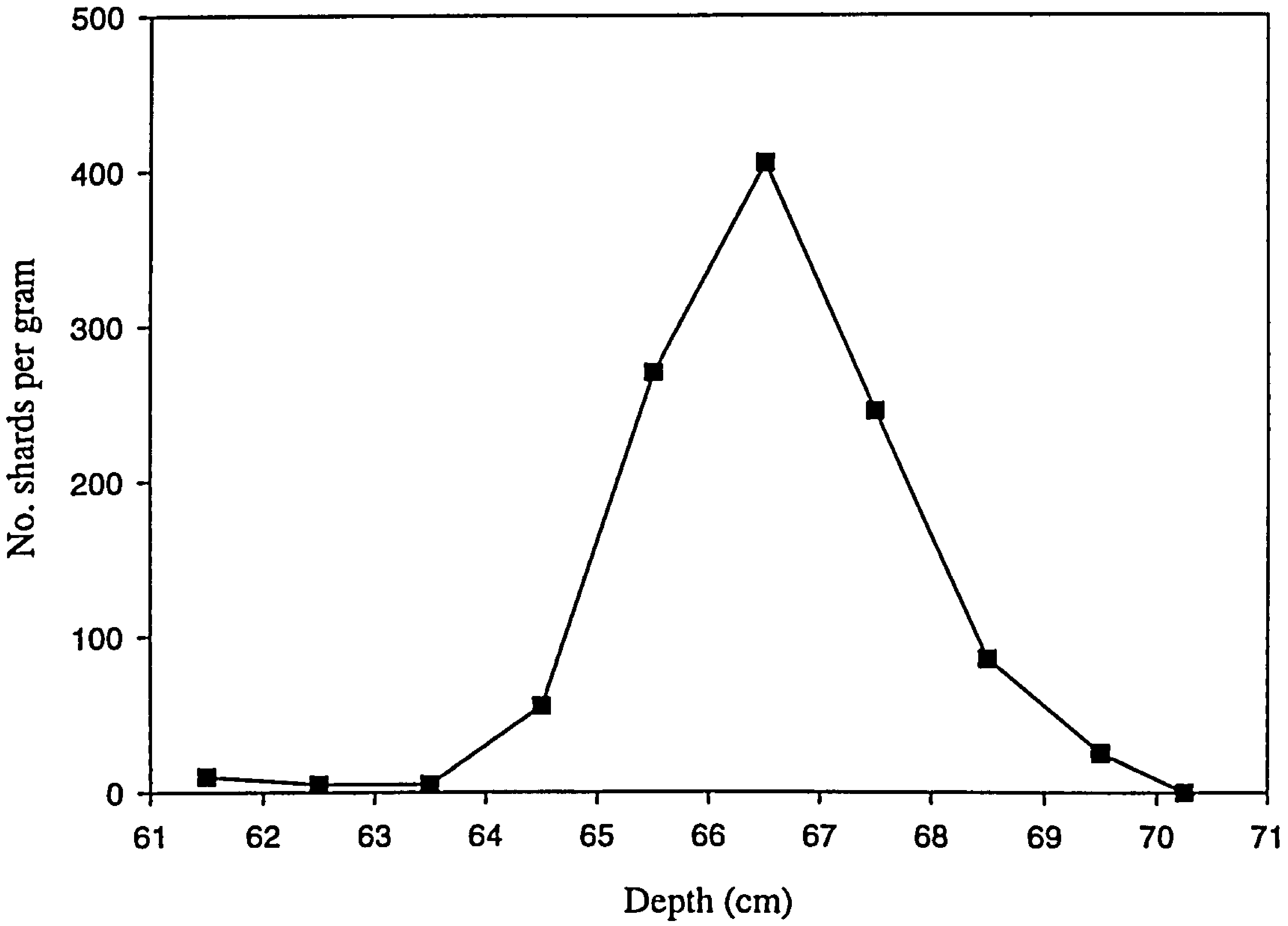


Figure 5.2n Distribution of Snaefellsjokull(?) tephra in 1cm peat slices from Ben Gorm Moss

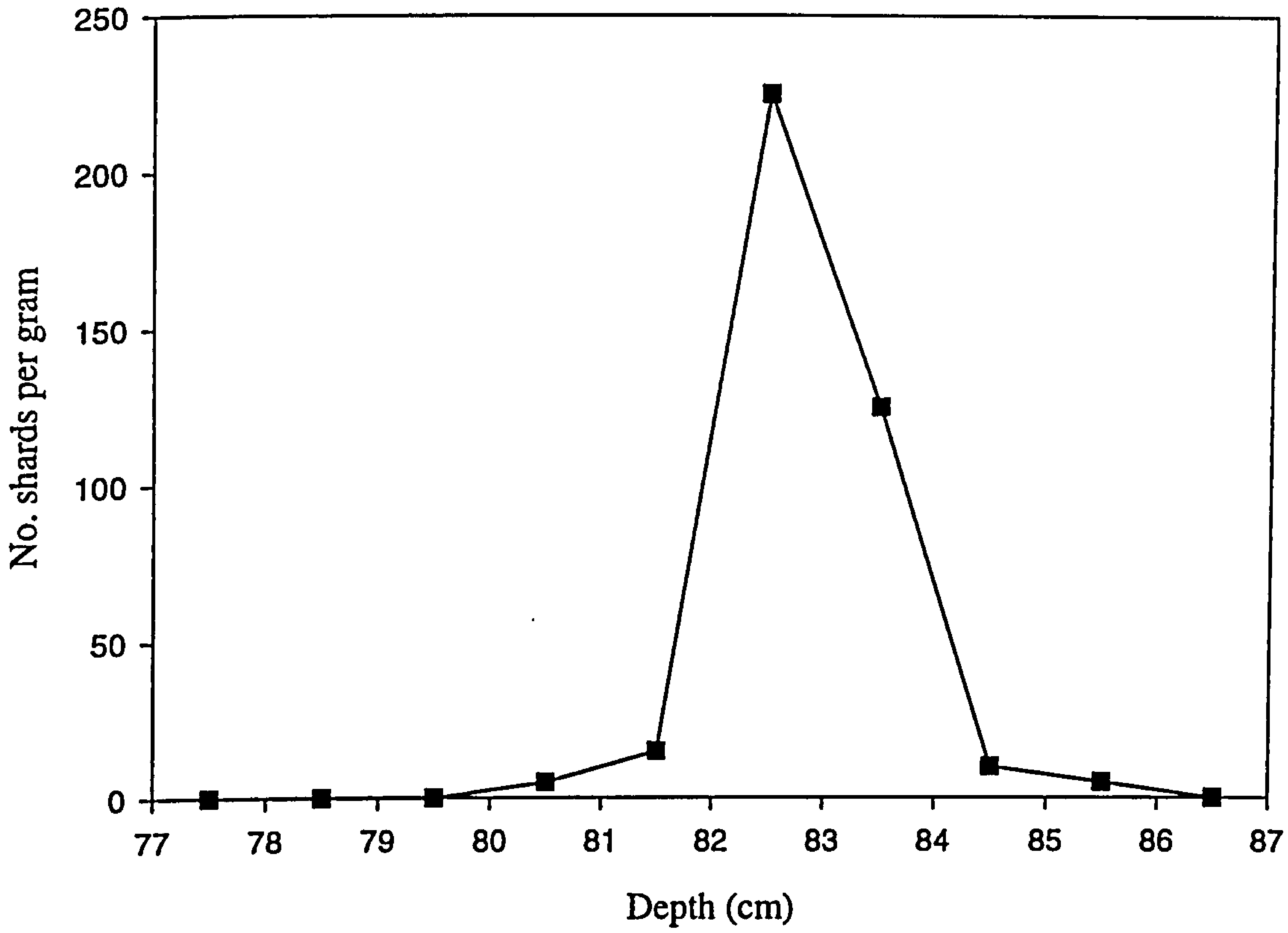


Figure 5.2m Distribution of Hekla-4 tephra in 1cm peat slices from Ben Gorm Moss

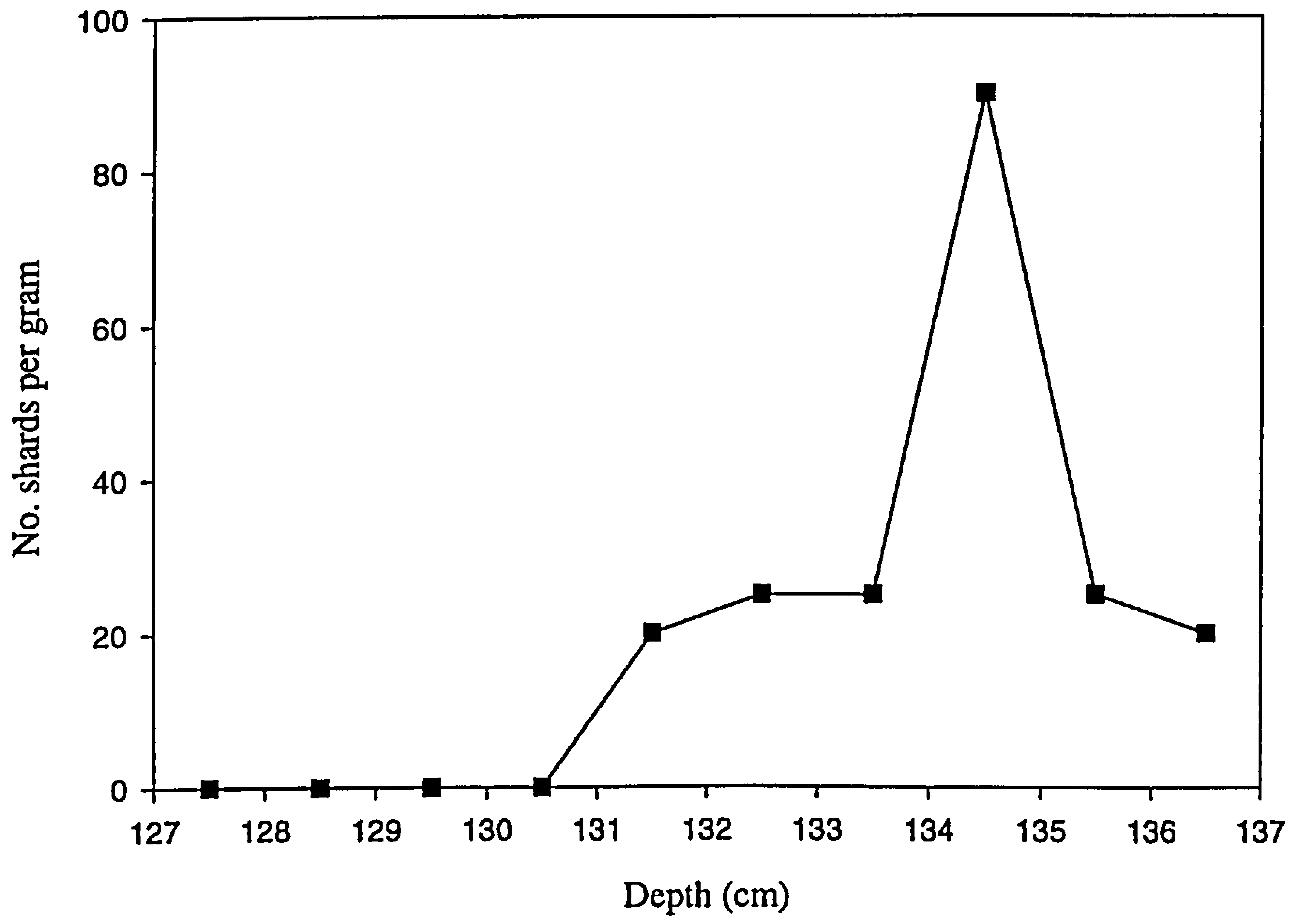


Table 5.1 Mid to late-Holocene tephras which have been identified in northern Britain and Ireland and which have published geochemistries.

Tephra	Comments	Author
Hekla AD 1510	Has currently been located in the Western Isles, Sutherland, and Northern Ireland. Youngest tephra layer so far found in Britain.	Dugmore *et al.* (1995a) Dugmore *et al.* (1996) Pilcher *et al.* (1996)
Loch Portain B tephra	SiO_2 content > 70%. Two populations may indicate two tephra falls within ± 80 years. Some similarities in geochemistry to Oræfajökull 1362 AD. Scant occurrence so far leads to the non acceptance of a distinct isochrone	Dugmore *et al.* (1995a)
Oræfajökull 1362 AD	Characterised by high Na_2O and very low MgO (Larsen *et al.*, 1999). Located in Northern Ireland and reportedly found in Scotland (Pilcher *et al.*, 1996), it has also been described from a Greenland ice-core (Palais *et al.*, 1991).	Pilcher *et al.* (1995)
Hekla 1 (AD 1104)	Found in small quantities in Irish peats. Characterised by FeO:MgO and K_2O:TiO_2 ratios (Larsen *et al.*, 1999).	Pilcher *et al.* (1995) Pilcher *et al.* (1996)
'AD 860 layer'	As yet unattributed to an Icelandic eruption. Forms two distinct geochemical populations on the basis of different CaO, MgO and K_2O proportions.	Pilcher *et al.* (1995)
Glen Garry tephra (*ca.* 2100 BP)	Found extensively in Scotland and also located in northern England, this tephra may be distinguished from other silicic tephras by K_2O, CaO, MgO, and FeO contents.	Dugmore *et al.* (1995a) Pilcher and Hall (1996)
Kebister tephra (*ca.* 3600 BP)	Only found in Shetland, and appears to be absent from Orkney and the Scottish mainland. Differs from Glen Garry and older tephras on the basis of CaO/MgO ratios.	Dugmore *et al.* (1995a)
Hekla-4 tephra (2310 ± 20 BC)	By far the most studied tephra in Britain, it can be found in Scotland, Ireland, and northern England. A relatively large range of SiO_2 content has been found, ranging from 76%-59% in Scotland. The geochemistry can be clearly differentiated from most other tephras by differing ranges of variables, notably MgO/CaO ratios. Hekla-4 has been wiggle match dated to 2310 ± 20 BC (Pilcher *et al.*, 1995).	Dugmore (1989) Dugmore and Newton (1992) Dugmore *et al.* (1992) Hall *et al.* (1994) Dugmore *et al.* (1995a) Pilcher *et al.* (1995) Pilcher *et al.* (1996) Pilcher and Hall (1996)
Hoy tephra (*ca.* 5600 BP)	To date this has only been found on the Orkney Islands. It has a narrow range of SiO_2 and characteristically high alkali content.	Dugmore *et al.* (1995a)
Lairg layers (A+B) (*ca.* 6000 BP)	Stratigraphically distinct tephra zone composed of two distinct silicic populations of glass shards. All shards exhibit low FeO and TiO_2 and some similarities can be found with the Hekla-4 tephra. Additional dates for the two populations can be found in Pilcher *et al.* (1996).	Dugmore *et al.* (1995a) Pilcher *et al.* (1996)

A summary of the profiles found in Scotland is shown in Figure 5.3

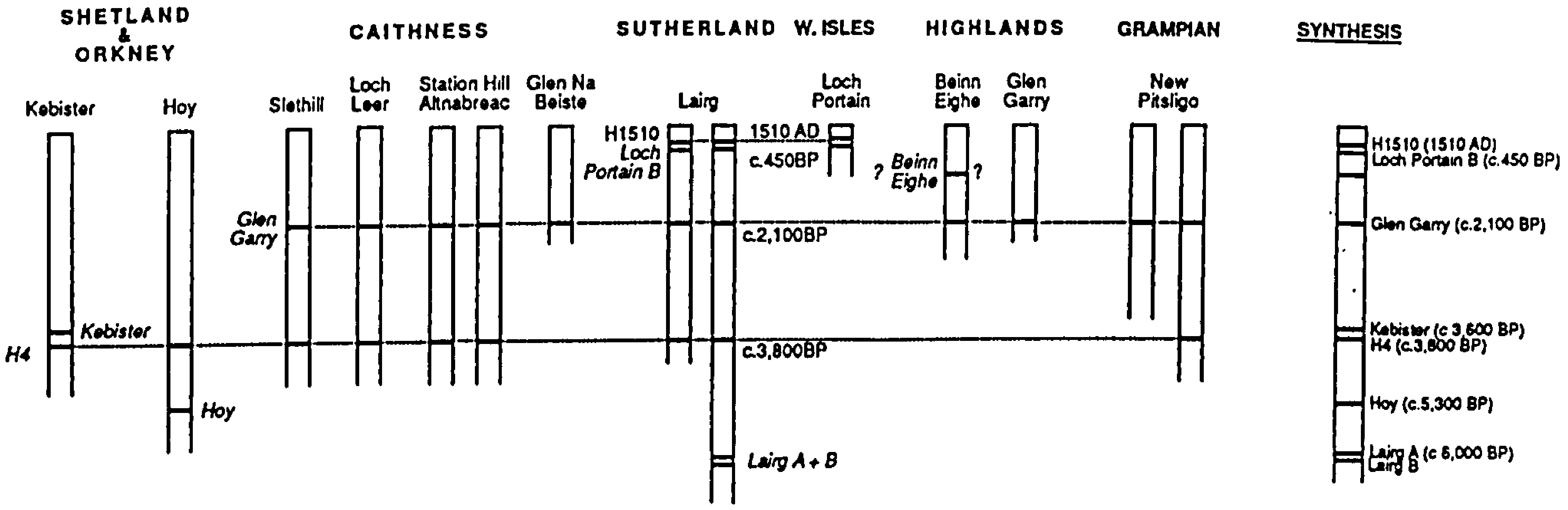


Figure 5.3 A diagrammatic summary of 14 profiles showing the tephras at each site and an idealised sequence for Scotland as a whole. The scale is arbitrary so that isochrones can be presented as horizontal lines (after Dugmore *et al.*, 1995a).

Standardising operating conditions and data presentation in order to allow comparisons between results is of primary concern to tephrochronologists (Froggatt, 1992; Hunt and Hill, 1993). It is often difficult to achieve totals of oxides around 100% due to the presence of water and highly mobile elements in volcanic glass. Therefore total percentages presented in the literature have varied in range, although some automatically normalise to 100% (Bennett *et al.*, 1992). This is to be discouraged, as normalisation to 100% effectively prevents any comparisons by outside workers, extends geochemical fields, and hides low totals (Boygle, 1994). The INQUA recommendations that a total above 90% is acceptable (Froggatt, 1992) has been disputed by Hunt and Hill (1993), who suggest a minimum cut off boundary of 95%, as lower totals may reflect high water content associated with glass alteration and poor point selection. Dugmore *et al.*, (1992) demonstrate that values of 95% and over are acceptable, as they are consistently reproducible in Britain. Data from Britain with totals having a lower threshold of 95% have since been consistently accepted (Hall *et al.*, 1994b; Pilcher *et al.*, 1995; Dugmore *et al.*, 1995a; Pilcher and Hall, 1996; Pilcher *et al.*, 1996; Turney *et al.*, 1997). However, the high levels of vesicularity found in British tephras could also result in the partial alteration of the glass, being represented by lower totals (Dugmore *et al.*, 1995a). Having stated this, the results in this thesis are presented as totals with a lower threshold of 93%, as these have proved reproducible within the data sets, with many analyses producing results in the range 94-95%. It was decided to include these results to maximise the datasets, while still maintaining data reproducibility (in agreement with A. Dugmore, pers. comm.).

All geochemical data has been compared with the relevant literature, as well as being correlated with Tephrabase: *http://www.geo.ed.ac.uk/tephra/tbasehom.html* (Edinburgh University). The analysis no. on the tables containing geochemical data refers to a number given to each shard analysed by electron probe microanalysis (EPMA). The bracketed number after this refers to either the first (08/97) or second (10/98) session of analysis.

5.2.3.2 Longbridge Moss

Two distinct isochrones were discovered at Longbridge, between depths of 60-65cm (Longbridge Moss Tephra 1, LBMT-1) and around 283-284cm (LBMT-2) stratigraphically. Unfortunately microprobe analysis of LBM-1 was problematic. One slide had the araldyte removed when being polished with 6μm diamond paste, and the second slide which had been prepared contained no identifiable tephra shards, and was therefore unable to be geochemically analysed. Analysis on LBMT-2 however produced a suite of results as shown in Table 5.2.

Table 5.2 Longbridge Moss LBMT-2: Hekla-4

Analysis No (2)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
140	74.89	0.11	13.15	2.03	0.02	0.03	1.40	3.26	2.79	97.68
141	73.91	0.13	12.80	1.81	0.03	0.05	1.41	2.28	2.74	95.16
101	73.73	0.08	12.88	1.83	0.02	0.02	1.42	3.87	2.61	96.46
144	73.65	0.09	12.80	1.97	0.08	0.05	1.26	2.96	2.48	95.34
119	73.22	0.17	12.74	2.12	0.11	0.06	1.51	3.98	2.61	96.52
113	73.02	0.07	12.75	1.88	0.07	0.02	1.25	4.08	2.75	95.89
103	72.98	0.08	12.70	1.95	0.06	0.03	1.47	4.33	2.86	96.46
139	72.79	0.17	12.59	1.95	0.09	0.03	1.38	2.14	2.85	93.99
100	72.69	0.06	12.73	1.82	0.13	0.04	1.35	4.17	2.70	95.69
111	72.48	0.14	12.66	1.83	0.10	0.04	1.30	4.20	2.91	95.66
143	72.18	0.10	12.35	1.78	0.04	0.05	1.28	2.91	2.66	93.35
121	72.18	0.10	11.97	1.95	0.05	0.05	1.35	4.38	2.98	95.01
122	71.74	0.09	12.56	1.93	0.05	0.05	1.28	4.48	2.77	94.95
118	71.72	0.12	12.92	1.97	0.07	0.06	1.36	4.08	2.85	95.15
120	71.26	0.18	12.71	2.05	0.03	0.06	1.41	4.23	2.71	94.64
110	70.63	0.08	12.54	1.88	0.03	0.05	1.42	4.22	2.50	93.35
Mean	72.69	0.11	12.68	1.92	0.06	0.04	1.37	3.72	2.74	95.33
SD	1.08	0.04	0.26	0.10	0.03	0.01	0.08	0.76	0.14	1.17

Total iron is expressed as FeO.

The data in Table 5.2 are comparable to the Hekla-4 data presented by Dugmore *et al.*, (1995a) and Pilcher *et al.*, (1995). The data are shown in Figure 5.12 along with the other data from Hekla-4 produced in this study.

5.2.3.3 Langlands Moss

Figure 5.1(f) suggests that there may be 3 isochrones present at Langlands, but further analysis revealed only 2 tephra layers, around 30-31cm (LAGT-1) and 85-86cm (LAGT-2). EPMA proved difficult on both ash horizons, with only a few shards surviving the preparation procedure and being exposed on the surface of the slide. However, geochemical data were obtained for both isochrones and are presented in Tables 5.3 and 5.4.

Table 5.3 Langlands Moss LAGT-1: 'AD 860 Layer'

Analysis No (2)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
135	73.14	0.11	11.81	1.63	0.02	0.21	1.03	3.95	3.52	95.42
136	72.77	0.17	12.25	1.53	0.04	0.07	0.79	3.81	3.65	95.08
137	72.50	0.16	12.18	1.62	0.07	0.08	0.90	3.70	3.69	94.90
132	69.66	0.20	13.33	1.69	0.00	0.59	2.25	3.81	2.82	94.35
Mean	72.02	0.16	12.39	1.62	0.03	0.24	1.24	3.82	3.42	94.94
SD	1.59	0.04	0.65	0.07	0.03	0.24	0.68	0.10	0.41	0.45

Total iron is expressed as FeO.

Table 5.4 Langlands Moss LAGT-2: Glen Garry

Analysis No (1)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
287	73.40	0.57	12.61	3.93	0.29	0.53	2.23	2.65	1.99	98.20
288	72.38	0.63	12.43	3.90	0.26	0.53	2.20	2.62	1.95	96.90
84	71.35	0.64	12.66	4.29	0.13	0.58	2.78	2.23	1.77	96.43
86	69.46	0.85	12.87	5.27	0.13	0.75	3.08	2.83	1.77	97.01
Mean	71.65	0.67	12.64	4.35	0.20	0.60	2.57	2.58	1.87	97.14
SD	1.68	0.12	0.18	0.64	0.08	0.10	0.43	0.25	0.12	0.75

Total iron is expressed as FeO.

Both isochrones have differing geochemistries as can be seen in Figure 5.4. It is important to note that one of the shards from LAGT-1 (analysis no. 132) has very different values for MgO, CaO and K_2O, suggesting that two geochemical populations could be present within this isochrone (also seen in Figure 5.10). The fact that this one analysis alone forms a separate population becomes significant when compared with other data. Pilcher *et al.* (1995) provide data which defines an 'AD lower layer' at Sluggan Bog, containing a very similar geochemical signature to LAGT-1. More significantly, the data from Sluggan includes two populations, which are outlined in Figure 5.4, and enclose the data from LAGT-1. This layer was wiggle-match dated to AD 860 ± 20 (Pilcher *et al.*, 1995) and has since been refined to AD 776-887 by reassessing the data in a more formal Bayesian approach using the Gibbs sampling option in Oxcal v2.0 (Pilcher *et al.*, 1996). The isochrone is recognised as the 'AD 860 layer' and has not yet been attributed to an Icelandic eruption, or other source (Pilcher *et al.*, 1996). The field enclosing the data from LAGT-2 are taken from tephra horizons at other sites analysed on the Edinburgh MK V (shown in Pilcher and Hall, 1996), and clearly assign LAGT-2 to the Glen Garry tephra.

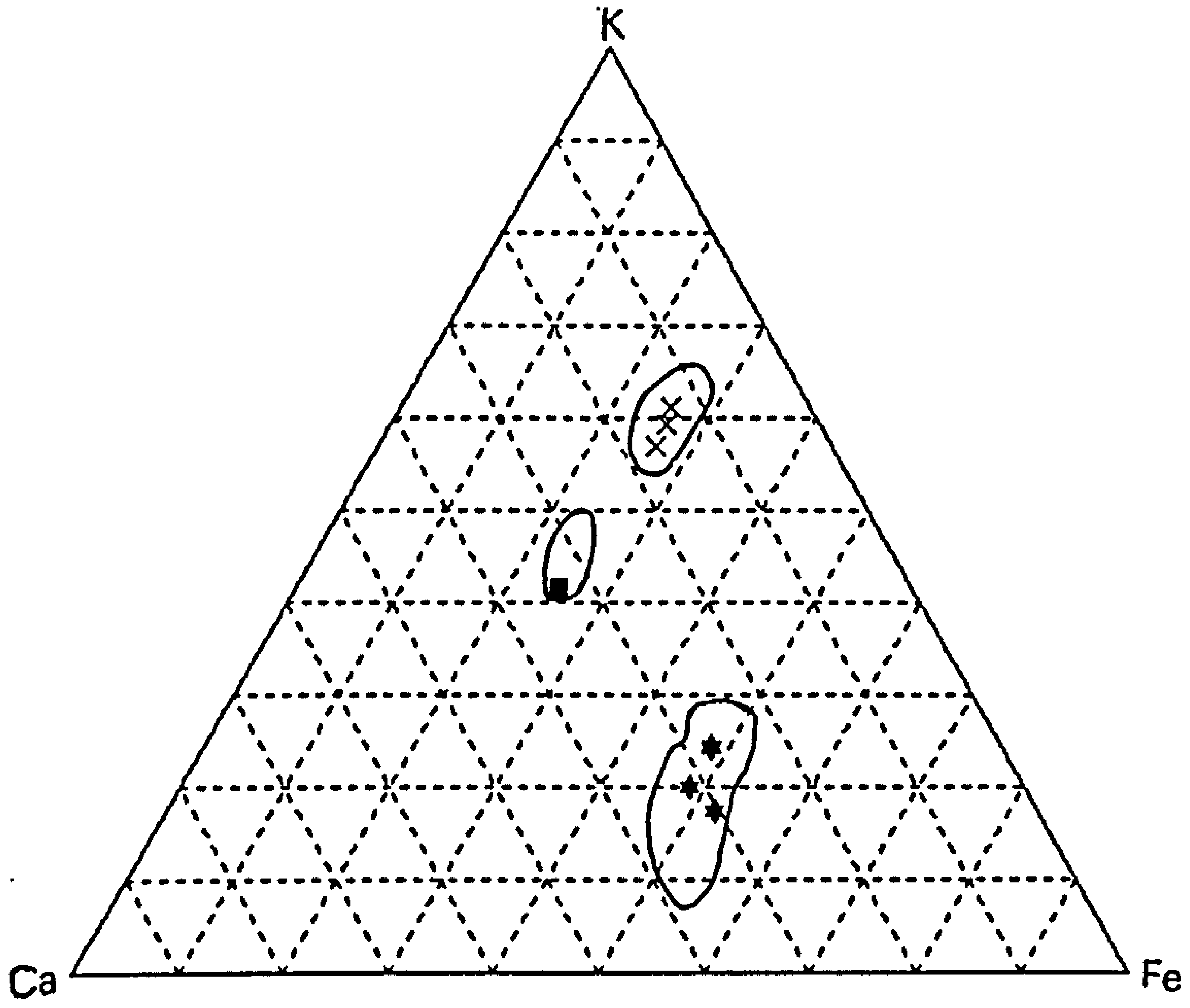


Figure 5.4 Ternary diagram showing the iron, calcium and potassium data from LAGT-1 and LAGT-2. LAGT-1 has two populations, A (crosses) and B (square), which correlate with the data fields from the 'AD 860 layer' (Pilcher *et al.*, 1995). LAGT-2 (stars) is surrounded by the data field from the Glen Garry tephra as explained in the text.

5.2.3.4 Temple Hill Moss

There are two well defined tephra isochrones at this site, around 114-115cm (TEMT-1) and 444-445cm (TEMT-2). In contrast with Langlands Moss, plenty of material was available for EMPA from TEMT-1, although TEMT-2 only yielded a few shards. The data are shown in Tables 5.5 and 5.6 respectively.

Table 5.5 Temple Hill Moss TEMT-1: Glen Garry

Analysis No (1)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
260	73.81	0.43	12.68	3.92	0.14	0.37	2.28	2.82	1.95	98.40
107	73.62	0.58	12.56	3.80	0.17	0.42	2.40	1.40	2.04	96.99
108	73.44	0.58	12.54	3.80	0.18	0.44	2.24	1.49	2.01	96.72
265	73.41	0.52	12.53	3.91	0.09	0.39	2.46	2.04	1.95	97.30
269	73.31	0.53	12.60	3.71	0.10	0.42	2.32	2.08	1.99	97.06
261	73.04	0.46	12.44	3.66	0.14	0.37	2.25	2.37	1.99	96.72
115	72.88	0.51	12.57	3.79	0.18	0.45	2.10	2.68	2.15	97.31
106	72.68	0.56	12.69	3.64	0.16	0.51	2.23	2.17	1.96	96.60
268	72.41	0.48	12.31	3.42	0.07	0.37	2.27	2.23	1.95	95.51
109	72.34	0.62	12.72	4.42	0.18	0.61	2.47	1.86	1.87	97.09
263	71.76	0.75	12.88	4.37	0.15	0.51	2.57	3.25	1.85	98.09
112	71.46	0.63	12.69	4.24	0.21	0.56	2.60	2.20	1.98	96.57
262	71.36	0.45	12.18	3.81	0.11	0.40	2.33	2.57	1.93	95.14
104	71.02	0.67	12.84	5.00	0.15	0.69	2.84	3.49	1.70	98.40
264	67.30	0.73	12.60	4.90	0.11	0.76	3.00	3.04	1.82	94.26
Mean	72.26	0.57	12.59	4.03	0.14	0.48	2.42	2.38	1.94	96.81
SD	1.63	0.10	0.18	0.46	0.04	0.12	0.24	0.60	0.10	1.15

Total iron is expressed as FeO.

Table 5.6 Temple Hill Moss TEMT-2: Lairg layer (A)

Analysis No (1)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
284	74.40	0.08	12.39	1.85	0.20	0.19	1.24	2.97	2.67	95.99
276	73.85	0.13	12.44	1.92	0.22	0.17	1.15	3.80	2.73	96.41
285	73.06	0.09	12.49	1.79	0.20	0.13	1.20	3.33	2.50	94.79
94	72.87	0.04	12.21	1.66	0.02	0.05	1.20	3.65	2.67	94.37
93	72.78	0.05	12.18	1.67	0.01	0.03	1.22	3.81	2.71	94.46
277	72.76	0.08	12.44	1.65	0.23	0.16	1.31	3.54	2.65	94.82
281	72.58	0.12	12.25	1.72	0.20	0.16	1.23	3.17	2.50	93.93
282	72.05	0.10	12.20	1.73	0.20	0.13	1.26	2.96	2.56	93.19
Mean	73.04	0.09	12.33	1.75	0.16	0.13	1.23	3.40	2.62	94.75
SD	0.74	0.03	0.13	0.10	0.09	0.06	0.05	0.35	0.09	1.04

Total iron is expressed as FeO.

The two populations contain very different geochemical characteristics, as can be seen in Figure 5.5.

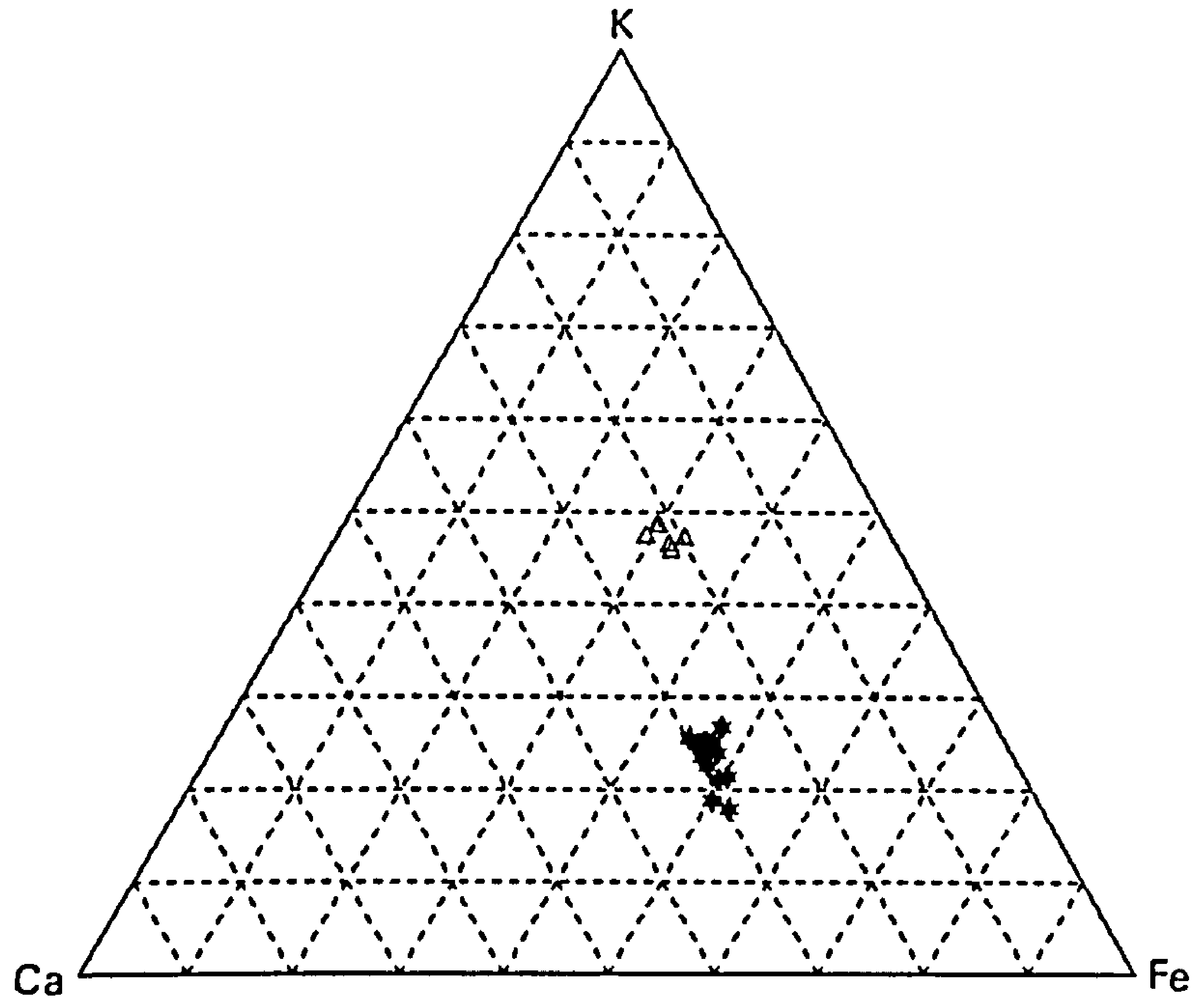


Figure 5.5 Ternary diagram of the iron, calcium and potassium geochemical data from Temple Hill Moss. Stars indicate Glen Garry glass, and triangles represent the Lairg layer.

Both isochrones form tight clusters within Figure 5.5, and comparisons with the data from other sites in this study, and data from Dugmore *et al.* (1995a) suggest that TEMT-1 is from the Glen Garry isochrone. TEMT-2, however, is more problematic. Initially it was thought that the shards were from the Hekla-4 isochrone. The ternary diagram (Figure 5.5) shows them to be located within the same population cluster as Hekla-4 shards, but the stratigraphic depth of this would suggest inconsistently high accumulation rates for the deeper peats, especially as the unambiguous Glen Garry pinning point was only at a depth of 115 cm. The stratigraphic positioning of the tephra isochrones in relation to peat accumulation rates is therefore important to note (Dugmore *et al.*, 1995b) and another solution was sought.

Sometimes particular geochemical compositions may be replicated by different eruptions from the same volcanic system (Larsen *et al.*, 1999). One other isochrone with similar geochemistry to Hekla-4 has been identified, the 'Lairg layers' which have been ^{14}C dated to *ca.* 6000 BP (Dugmore *et al.*, 1995a). The geochemistry of TEMT-2 is similar to the Lairg A layer data (medium-K glass) from Dugmore *et al.* (1995a) in Scotland and Pilcher *et al.* (1996) in Northern Ireland, as opposed to the Lairg B layer which is a high-K glass.

Dugmore *et al.* (1995a) note how close the age of the Lairg layer is to that of Hekla-5 which has been ^{14}C dated to *c.* 6200 BP (Larsen and Thorarinsson, 1977). However, it was also noted by Dugmore *et al.* (1995a) that the only published data on Hekla-5 had considerably lower K_2O values, and so as yet no correlations can be made. The radiocarbon dates presented in Section 5.3 confirm that TEMT-2 could not be tephra from Hekla-4, and suggest that it is indeed the Lairg A layer.

5.2.3.5 Shirgarton Moss

Two tephra isochrones are present at Shirgarton Moss, between 55-60cm (SGMT-1) and around 135-136cm (SGMT-2). Unfortunately SGMT-1 did not yield any geochemical data as no shards could be identified during microprobe analysis, possibly due to the small size of the shards (<30μm on average). Results were obtained from SGMT-2, although only a few shards were identified, and many became damaged by the electron beam before a full geochemical readout could be attained, resulting in low totals. The two successful geochemical analyses are shown in Table 5.7.

Table 5.7 Shirgarton Moss SGMT-2: Glen Garry

Analysis No (2)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
125	72.15	0.51	12.26	3.67	0.09	0.40	2.12	3.45	2.03	96.68
124	70.77	0.56	11.97	3.76	0.08	0.46	2.40	2.75	3.59	96.34
Mean	71.46	0.54	12.12	3.72	0.09	0.43	2.26	3.10	2.81	96.51
SD	0.98	0.04	0.21	0.06	0.01	0.04	0.20	0.49	1.10	0.24

Total iron is expressed as FeO.

Analysis no. 125 has a geochemistry very similar to that of the Glen Garry tephra, noticeable by the CaO:MgO and FeO: TiO_2 ratios. However, analysis no. 124 has similar ratios of CaO:MgO and FeO:TiO_2, but an anomalously high K_2O value. This high potassium value is not replicated in any of the Glen Garry tephra data found in Tephrabase. The low Na_2O value for analysis no. 124 is replicated by some data from Glen Garry tephra, although it cannot necessarily be explained by the migration of sodium during the analysis of silicic glass. The sodium content is measured at the start and end of each analysis, and actually decreased during analysis no. 124 by only 17% (from 1137 to 949 counts per 10 seconds), as opposed to more common decreases of over 50% (e.g. typical value for Glen Garry Na_2O being *ca.* 1500 counts at the start and only *ca.* 600

counts after 10 seconds). The alkali content of analysis 124 therefore should be treated with caution, and cannot be fully explained if the shard is from the Glen Garry tephra. However, the shard analysed as no. 125 displays all the characteristics of Glen Garry tephra, and the depth at which the tephra occurs is also stratigraphically acceptable. The radiocarbon dates for Shirgarton (see Section 5.3) suggest that Glen Garry would be a sensible choice for the tephra isochrone. It is noted that while one analysis cannot significantly represent a geochemical marker, other evidence points towards SGMT-2 being the Glen Garry tephra, and it is therefore tentatively accepted as such.

5.2.3.6 Mallachie Moss

Two quite distinct tephra horizons were recovered from Mallachie Moss, around 120-121cm (MALT-1) and 261-262cm (MALT-2). Tephra concentrations for both isochrones were high, and the glass geochemistry data are shown in Tables 5.8 and 5.9.

Table 5.8 Mallachie Moss MALT-1: Glen Garry

Analysis No (1)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
246	74.02	0.52	12.65	3.83	0.10	0.42	2.29	2.85	1.87	98.55
18	74.00	0.56	12.83	3.72	0.09	0.47	2.13	2.48	1.91	98.19
21	73.73	0.51	12.73	3.83	0.12	0.43	2.26	2.75	2.11	98.47
14	73.73	0.47	12.59	3.60	0.07	0.43	2.15	2.77	1.97	97.78
242	73.60	0.50	12.66	3.65	0.12	0.44	2.36	3.02	1.93	98.28
19	73.50	0.56	12.57	3.74	0.12	0.46	2.13	2.72	2.59	98.39
233	73.45	0.43	12.35	3.52	0.10	0.39	2.18	2.95	1.98	97.35
237	73.16	0.58	12.82	3.95	0.09	0.44	2.46	3.04	1.92	98.46
244	73.05	0.59	12.67	3.86	0.14	0.41	2.28	3.11	1.97	98.08
10	72.59	0.43	12.62	3.84	0.10	0.41	2.31	2.51	1.98	96.79
247	72.50	0.54	12.56	3.79	0.09	0.46	2.45	3.47	1.92	97.78
238	72.40	0.54	12.50	4.01	0.12	0.48	2.33	3.14	1.93	97.45
243	72.13	0.49	12.70	3.90	0.12	0.49	2.16	3.04	2.04	97.07
11	72.04	0.60	12.92	4.49	0.11	0.56	2.69	2.70	1.94	98.05
24	71.97	0.54	12.42	3.95	0.11	0.38	2.16	2.72	1.98	96.23
249	71.89	0.66	12.95	5.02	0.14	0.72	3.04	3.52	1.93	99.87
17	71.37	0.64	12.99	4.55	0.10	0.60	2.65	2.98	1.89	97.77
15	70.84	0.55	11.99	3.71	0.09	0.39	2.31	2.54	1.96	94.38
232	70.78	0.41	11.90	3.71	0.12	0.42	2.16	2.79	1.95	94.24
3	70.71	0.52	12.18	3.74	0.10	0.43	2.23	2.82	2.03	94.76
6	70.64	0.56	11.93	3.81	0.10	0.42	2.28	2.50	1.88	94.12
239	70.63	0.81	13.09	5.29	0.17	0.77	2.96	2.94	1.69	98.35
252	70.58	0.47	12.16	3.59	0.09	0.39	2.28	2.70	1.84	94.10
245	70.18	0.44	11.95	3.70	0.10	0.43	2.18	3.56	1.94	94.48
5	70.14	0.74	12.61	4.65	0.10	0.67	2.78	3.19	2.02	96.90
248	69.99	0.74	12.99	5.05	0.20	0.72	3.10	3.81	1.82	98.42
20	69.95	0.71	12.77	4.32	0.09	0.60	2.76	2.50	1.75	95.45
240	69.67	0.84	13.02	5.19	0.18	0.82	3.11	2.85	1.71	97.39
9	69.41	0.66	12.66	5.07	0.15	0.64	2.74	2.05	1.83	95.21

Mean	71.88	0.56	12.58	4.02	0.11	0.49	2.38	2.90	2.07	96.99
SD	1.45	0.14	0.36	0.67	0.03	0.15	0.39	0.38	0.59	1.65

Total iron is expressed as FeO.

Table 5.9 Mallachie Moss MALT-2: Hekla-4

Analysis No (1)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
229	74.59	0.07	13.10	1.95	0.07	0.07	1.40	3.85	2.83	97.93
228	74.24	0.13	12.80	2.06	0.07	0.02	1.27	3.97	2.77	97.33
49	74.16	0.12	13.41	2.01	0.08	0.04	1.31	3.63	2.86	97.62
44	74.12	0.11	13.34	1.99	0.12	0.06	1.26	3.89	2.85	97.74
45	73.94	0.13	13.16	1.98	0.11	0.05	1.36	3.61	2.82	97.16
223	73.87	0.07	12.91	1.97	0.06	0.04	1.27	4.12	2.61	96.92
222	73.64	0.09	12.90	2.02	0.10	0.03	1.33	3.89	2.70	96.70
187	73.29	0.10	12.79	1.97	0.08	0.07	1.44	3.47	2.69	95.90
219	73.28	0.07	12.75	1.91	0.04	0.05	1.38	4.00	2.83	96.31
54	73.05	0.14	13.44	2.04	0.08	0.06	1.32	3.48	2.91	96.52
210	73.03	0.14	12.87	1.98	0.12	0.03	1.31	4.01	2.62	96.11
53	73.02	0.10	13.16	1.83	0.06	0.04	1.19	3.48	2.84	95.72
47	73.00	0.11	12.92	1.84	0.16	0.04	1.34	3.21	2.80	95.42
224	72.97	0.08	12.86	1.90	0.10	0.01	1.40	3.44	2.72	95.48
218	72.90	0.15	12.90	1.87	0.07	0.05	1.35	3.54	2.66	95.49
32	72.79	0.06	13.15	2.02	0.08	0.05	1.39	3.27	2.81	95.62
36	72.72	0.03	12.41	1.68	0.11	0.06	1.31	3.37	2.88	94.57
46	72.68	0.08	12.88	2.03	0.10	0.04	1.29	3.14	2.95	95.19
52	72.64	0.14	12.98	1.91	0.08	0.05	1.26	3.24	2.69	94.99
220	72.53	0.10	12.69	1.81	0.10	0.06	1.19	3.55	2.86	94.89
185	72.50	0.07	12.63	1.96	0.09	0.04	1.32	3.53	2.79	94.93
38	72.29	0.11	12.76	1.95	0.15	0.06	1.23	3.35	2.71	94.61
188	72.15	0.10	12.64	1.93	0.09	0.03	1.25	4.05	2.70	94.94
42	72.13	0.12	13.15	1.84	0.08	0.02	1.33	3.51	3.18	95.36
40	71.77	0.14	13.04	1.83	0.15	0.05	1.27	3.19	2.85	94.29
193	71.35	0.09	13.06	2.37	0.13	0.02	1.50	3.72	2.63	94.87
43	70.79	0.10	12.78	1.94	0.08	0.04	1.25	3.06	4.61	94.65
Mean	72.94	0.10	12.94	1.95	0.10	0.04	1.32	3.58	2.86	95.83
SD	0.91	0.03	0.25	0.12	0.03	0.02	0.07	0.31	0.38	1.10

Total iron is expressed as FeO.

By comparing the data from Tables 5.8 and 5.9 with Dugmore *et al.* (1995a) and Tephrabase, it is suggested that MALT-1 and MALT-2 are composed of Glen Garry and Hekla-4 tephras respectively. The two isochrones can be separated geochemically by plotting a number of variables against each other, but one that is commonly used is the CaO:MgO ratios, which therefore allow direct comparisons with other literature. Although MgO values tend to be low, which is problematic if the analytical tools are poor (Larsen *et al.*, 1999), Glen Garry and Hekla-4 tephras in Scotland can be distinguished by this ratio (Dugmore *et al.*, 1995a), as shown in Figure 5.6.

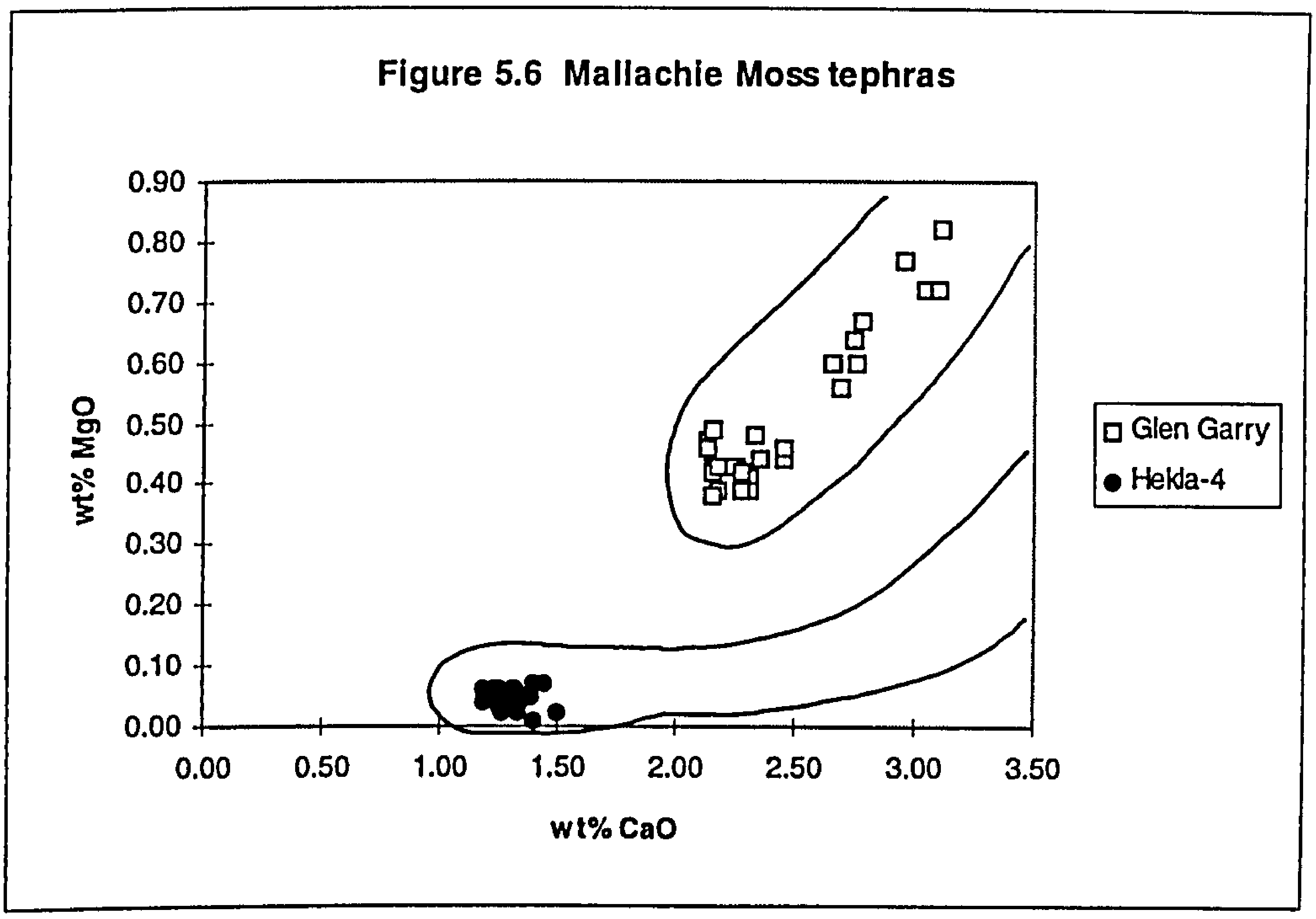


Figure 5.6 Mallachie Moss tephras - the Glen Garry and Hekla-4 tephras can be distinguished by the ratio of calcium to magnesium. The outlines indicate the populations of Glen Garry and Hekla-4 tephras from Dugmore *et al.* (1995a).

5.2.3.7 Craigmaud Moss

Two very distinctive populations of tephra can be seen from Figure 5.1(l), with peak concentrations occurring around 104-105cm (CMDT-1) and 139-140cm (CMDT-2). Tephra has been analysed from a nearby site at New Pitsligo where Glen Garry and Hekla-4 tephras were found (Dugmore *et al.*, 1995a), and it was therefore expected that these isochrones should be present at Craigmaud. Tephra loading was highest at this site, so consequently many shards have been analysed. The geochemical data are presented in Tables 5.10 and 5.11.

Table 5.10 Craigmaud Moss CMDT-1: Glen Garry

Analysis No (1)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
117	74.26	0.56	12.59	3.70	0.14	0.41	2.11	2.96	2.04	98.77
125	74.11	0.60	12.72	3.59	0.10	0.44	2.31	3.18	2.01	99.06
122	73.98	0.48	12.59	3.73	0.11	0.42	2.23	3.26	2.07	98.87
255	73.77	0.54	12.48	3.88	0.13	0.40	2.24	2.90	2.35	98.69
256	73.56	0.54	12.40	3.80	0.07	0.45	2.30	3.68	2.02	98.82
137	73.46	0.51	12.41	3.76	0.11	0.45	2.28	3.26	2.03	98.27
134	73.16	0.51	12.47	3.67	0.06	0.39	2.09	3.17	1.91	97.43
144	72.55	0.49	12.37	3.61	0.08	0.40	2.18	2.48	1.60	95.76
119	72.32	0.50	12.33	3.90	0.11	0.41	2.25	2.47	1.99	96.28
131	72.02	0.70	12.38	3.98	0.10	0.45	2.40	2.98	1.95	96.96
130	71.96	0.68	13.00	4.25	0.14	0.58	2.53	2.96	1.79	97.89
61	71.77	0.60	13.01	4.18	0.12	0.48	2.62	3.01	1.92	97.71
138	71.67	0.49	12.08	3.49	0.10	0.39	2.03	2.51	1.94	94.70
60	71.66	0.61	13.07	4.37	0.13	0.50	2.51	2.85	1.88	97.58
121	70.95	0.76	12.64	5.01	0.19	0.64	2.79	3.06	1.80	97.84
140	70.87	0.77	12.89	4.63	0.12	0.65	2.83	3.05	1.77	97.58
57	70.67	0.67	13.05	4.66	0.10	0.62	2.99	2.89	1.84	97.49
141	70.66	0.79	12.91	4.89	0.19	0.64	2.97	2.79	1.81	97.65
56	70.63	0.80	13.22	4.17	0.12	0.64	2.87	2.23	1.70	96.38
120	70.15	0.73	12.97	4.97	0.10	0.69	2.93	3.09	1.69	97.32
257	66.40	1.06	13.38	6.28	0.21	1.19	4.12	3.33	1.54	97.51
127	64.06	1.24	13.29	7.11	0.24	1.58	4.66	3.42	1.36	96.96
136	62.72	1.32	13.32	7.56	0.16	1.69	4.68	3.01	1.33	95.79
126	62.26	1.40	13.41	8.02	0.20	1.90	5.35	3.39	1.14	97.07
Mean	70.98	0.70	12.81	4.53	0.13	0.66	2.78	3.03	1.85	97.47
SD	3.49	0.29	0.39	1.38	0.05	0.45	0.95	0.37	0.33	1.08

Total iron is expressed as FeO.

Table 5.11 Craigmaud Moss CMDT-2: Hekla-4

Analysis No (1)	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
77	75.55	0.15	13.40	2.02	0.20	0.04	1.42	4.15	2.95	99.88
80	75.21	0.12	13.36	2.07	0.16	0.06	1.38	3.98	3.07	99.41
78	75.10	0.18	13.41	2.05	0.14	0.05	1.34	3.61	3.65	99.53
183	74.99	0.07	13.19	1.98	0.09	0.01	1.33	3.89	2.84	98.39
153	74.71	0.08	13.08	1.93	0.11	0.01	1.35	4.06	2.75	98.08
70	74.45	0.11	13.58	1.98	0.13	0.10	1.37	3.42	2.72	97.86
168	74.30	0.11	12.65	2.04	0.08	0.03	1.38	3.71	2.93	97.23
171	74.17	0.11	12.90	1.97	0.06	0.03	1.35	3.84	2.92	97.35
182	74.14	0.18	12.99	1.89	0.10	0.03	1.38	4.03	3.01	97.75
181	74.01	0.09	12.96	1.98	0.13	0.01	1.32	3.97	2.67	97.14
73	73.80	0.15	13.18	2.15	0.13	0.04	1.23	3.39	2.57	96.64
151	73.76	0.12	13.18	1.97	0.05	0.04	1.30	3.55	3.09	97.06
75	73.68	0.13	13.35	2.23	0.06	0.04	1.17	4.00	2.84	97.50
67	73.56	0.10	13.20	2.01	0.13	0.06	1.25	2.99	2.99	96.29
149	73.36	0.10	12.93	1.92	0.08	0.00	1.33	3.29	2.74	95.75
166	73.21	0.08	12.82	2.05	0.07	0.02	1.19	3.80	3.34	96.58
169	73.17	0.13	13.05	1.83	0.05	0.04	1.14	2.76	2.65	94.82
173	73.10	0.10	12.64	1.88	0.09	0.02	1.30	4.41	2.71	96.25
175	73.08	0.15	12.71	1.76	0.03	0.03	1.21	3.68	3.13	95.78
177	72.86	0.11	12.70	1.80	0.10	0.04	1.27	3.70	2.70	95.28
157	72.84	0.11	12.40	2.04	0.03	0.00	1.31	3.34	2.56	94.63

65	72.78	0.19	13.31	2.04	0.13	0.05	1.41	3.68	2.93	96.52
159	72.70	0.10	13.03	2.06	0.08	0.04	1.37	4.01	2.61	96.00
179	72.69	0.08	12.54	1.91	0.07	0.04	1.13	3.88	2.88	95.22
165	72.61	0.11	12.62	1.80	0.05	0.05	1.30	3.77	2.65	94.96
150	72.55	0.13	13.77	2.79	0.11	0.06	1.79	3.83	2.60	97.63
163	72.32	0.11	12.40	2.06	0.10	0.07	1.27	3.00	4.86	96.19
69	72.32	0.15	12.90	1.90	0.17	0.06	1.32	3.10	2.78	94.70
160	72.31	0.07	12.62	1.89	0.06	0.05	1.25	4.11	2.75	95.11
180	72.29	0.14	12.64	1.91	0.04	0.03	1.30	3.48	2.67	94.50
72	71.89	0.13	13.02	2.10	0.15	0.06	1.36	3.62	2.85	95.18
66	71.06	0.24	14.10	3.36	0.18	0.06	2.03	3.63	2.52	97.18
178	70.71	0.22	13.61	3.39	0.17	0.03	2.04	4.23	2.69	97.09
164	70.41	0.22	13.77	3.78	0.14	0.05	2.24	3.62	2.45	96.68
64	68.29	0.32	14.02	3.96	0.22	0.08	2.38	3.02	2.39	94.68
Mean	73.03	0.14	13.08	2.19	0.11	0.04	1.42	3.67	2.87	96.54
SD	1.46	0.05	0.44	0.56	0.05	0.02	0.31	0.39	0.43	1.43

Total iron is expressed as FeO.

The two geochemical populations clearly correlate well with those from Mallachie Moss, and as expected, show the geochemical characteristics of the Glen Garry and Hekla-4 tephras. The CaO:MgO ratios are plotted in Figure 5.7.

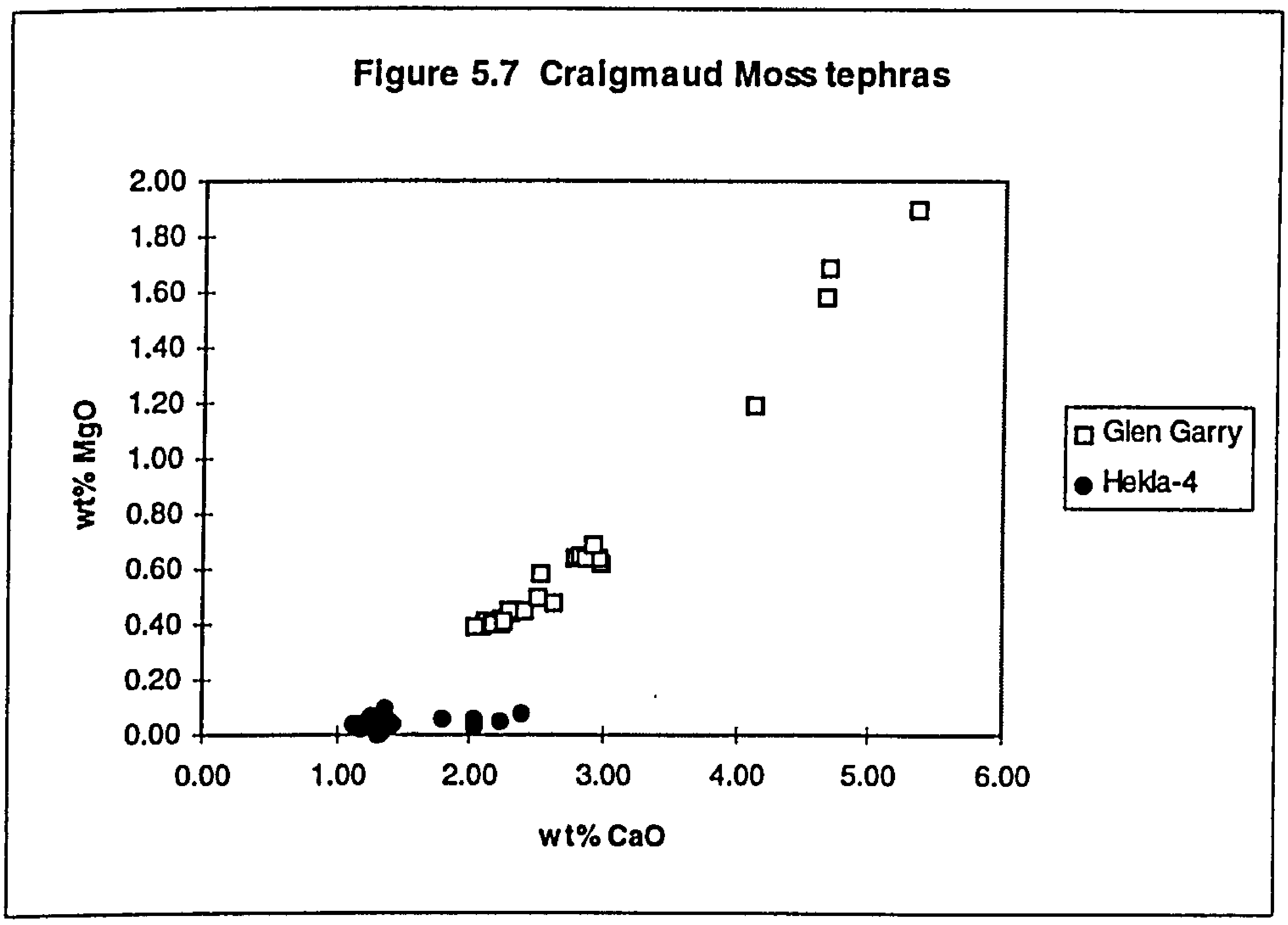


Figure 5.7 The Craigmaud tephras - calcium and magnesium geochemical data are plotted to show the different populations of Glen Garry and Hekla-4 tephras.

Although the geochemical signatures can be seen clearly, the spread of data within the populations is greatly increased compared to the Mallachie data. This increased spread is due to the less silicic (more dacitic) glass shards being identified at Craigmaud. Rhyolitic tephra (silicic content >70%) is most commonly found in Scotland, with the dacitic component generally having a silicic range between 64-70%. The terminology is based on the recommendations of Le Maitre (1989).

The geochemical populations shown in Figure 5.7 compare favourably with other sites in Scotland (Dugmore *et al.*, 1995a). The dacitic component is found in both the Hekla-4 and Glen Garry tephras at this site, although the geochemical data for Craigmaud (Tables 5.10 and 5.11) shows that the dacitic shards are more commonly found, and are comparatively less silicic, in the Glen Garry tephra. This has implications for tephra dispersal, which is discussed in Section 5.2.4.

5.2.3.8 Ben Gorm Moss

The blanket mire at Ben Gorm contains at least four tephra isochrones, and there is a suggestion from Figure 5.1(k) that there may be as many as six, making it comparable with the six tephra layers recovered from Talla Moss (Chambers *et al.*, 1997a), although the geochemistry for these isochrones has not yet been published. Four isochrones from Ben Gorm have been geochemically analysed, including those around 30-31cm (BGMT-1), 66-67cm (BGMT-2), 82-83cm (BGMT-3) and 134-135cm (BGMT-4). The geochemical data are shown in Tables 5.12-5.15.

Table 5.12 Ben Gorm Moss BGMT-1 (31cm)

Analysis No	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
80	74.94	0.20	12.28	1.48	0.03	0.04	0.73	3.74	3.76	97.20
32	69.93	0.43	12.80	2.83	0.03	0.38	1.79	3.27	2.57	94.03
81	68.55	0.24	12.91	3.53	0.15	0.23	1.40	4.64	3.48	95.13
82	67.20	0.88	13.80	5.31	0.25	0.86	2.67	3.39	2.91	97.27
83	66.95	1.05	13.78	5.13	0.21	0.84	2.59	3.62	2.75	96.92
36	65.09	0.97	13.45	5.23	0.13	0.81	2.42	2.61	5.98	96.69
Mean	67.54	0.71	13.35	4.41	0.15	0.62	2.17	3.51	3.54	96.01
SD	3.43	0.38	0.61	1.58	0.09	0.36	0.77	0.67	1.26	1.32

Total iron is expressed as FeO.

Table 5.13 Ben Gorm Moss BGMT-2: Glen Garry

Analysis No	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
161	73.26	0.46	12.48	3.7	0.14	0.44	2.32	3.49	1.88	98.17
162	72.94	0.54	12.23	4.09	0.09	0.45	2.37	3.68	1.95	98.34
16	72.13	0.64	12.6	4.27	0.08	0.46	2.66	2.62	1.77	97.23
17	72.06	0.55	12.27	3.32	0.11	0.44	2.61	2.11	1.69	95.16
11	71.92	0.49	12.23	3.59	0.12	0.47	2.4	2.94	2.02	96.18
18	71.87	0.45	12.29	3.79	0.06	0.40	2.5	3.27	2.15	96.78
19	71.61	0.41	12.17	3.52	0.03	0.33	2.28	2.63	1.88	94.86
163	71.06	0.48	11.82	3.63	0.05	0.42	2.17	3.58	1.96	95.17
2	69.92	0.71	13.14	5.08	0.15	0.82	2.97	3.17	1.73	97.69
4	69.63	0.76	12.98	4.70	0.10	0.67	2.88	2.74	1.69	96.15
3	68.67	0.72	12.52	5.12	0.08	0.68	2.84	3.6	1.83	96.06
25	67.88	0.77	12.63	4.60	0.10	0.66	2.97	2.78	1.80	94.19
10	67.39	0.85	13.00	5.80	0.10	1.35	4.33	2.96	1.52	97.30
159	66.12	1.00	11.96	7.10	0.16	2.55	6.00	3.53	1.43	99.85
6	65.19	0.96	12.69	5.87	0.14	1.06	3.86	3.12	1.51	94.40
Mean	70.11	0.65	12.47	4.55	0.10	0.75	3.01	3.08	1.79	96.50
SD	2.54	0.19	0.38	1.08	0.04	0.57	1.02	0.45	0.20	1.61

Total iron is expressed as FeO.

Table 5.14 Ben Gorm Moss BGMT-3: (81cm)

Analysis No	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
155	75.55	0.23	12.46	1.47	0.01	0.09	0.84	4.06	3.72	98.43
72	75.17	0.14	12.29	1.69	0.05	0.06	0.72	3.71	3.83	97.66
61	74.63	0.18	12.66	1.70	0.13	0.10	0.76	3.75	3.95	97.86
66	74.53	0.21	12.34	1.52	0.04	0.09	0.78	3.61	3.59	96.71
151	74.33	0.18	11.80	1.26	0.00	0.04	0.59	3.70	3.92	95.82
152	74.16	0.16	11.67	1.36	0.06	0.05	0.66	3.62	4.83	96.57
55	74.08	0.12	11.55	1.30	0.04	0.03	0.54	3.27	4.05	94.98
65	73.85	0.14	11.77	1.15	0.06	0.06	0.54	3.30	3.96	94.83
156	73.79	0.22	12.17	1.62	0.07	0.13	0.83	3.98	3.63	96.44
56	73.77	0.08	11.84	1.31	0.00	0.05	0.50	3.56	3.98	95.09
43	73.66	0.09	11.87	1.16	0.04	0.04	0.58	3.62	3.86	94.92
153	73.37	0.15	11.54	1.54	0.01	0.11	0.68	3.74	3.66	94.80
62	73.10	0.15	12.08	1.38	0.02	0.07	0.62	3.36	3.71	94.49
45	73.08	0.16	11.89	1.10	0.04	0.04	0.68	3.32	3.90	94.21
71	73.08	0.16	12.05	1.46	0.05	0.10	0.68	3.66	3.68	94.92
59	73.06	0.15	12.00	1.33	0.06	0.06	0.67	3.46	3.76	94.55
64	72.88	0.13	11.09	1.03	0.06	0.03	0.51	3.36	3.91	93.00
67	72.43	0.08	12.17	1.56	0.00	0.07	0.78	3.72	3.80	94.61
49	72.25	0.28	11.77	1.47	0.04	0.11	0.72	2.99	3.80	93.43
68	72.11	0.16	11.91	1.58	0.06	0.08	0.76	3.26	3.67	93.59
150	72.02	0.10	13.33	1.07	0.00	0.03	0.94	4.99	3.31	95.79
69	71.99	0.18	11.90	1.50	0.06	0.12	0.84	3.00	3.73	93.32
Mean	73.50	0.16	12.01	1.39	0.04	0.07	0.69	3.59	3.83	95.27
SD	1.01	0.05	0.45	0.20	0.03	0.03	0.12	0.42	0.28	1.49

Total iron is expressed as FeO.

Table 5.15 Ben Gorm Moss BGMT-4: Hekla-4

Analysis No	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
78	72.85	0.08	12.79	1.88	0.12	0.04	1.32	3.88	2.90	95.86
89	72.16	0.08	12.46	1.98	0.14	0.02	1.23	3.52	2.72	94.31
91	71.83	0.05	12.70	1.83	0.12	0.07	1.25	3.95	2.71	94.51
76	71.43	0.07	12.32	1.64	0.08	0.03	1.31	3.62	2.50	93.00
86	70.66	0.16	11.73	2.81	0.11	0.05	0.43	3.48	4.24	93.67
85	70.05	0.21	11.87	2.68	0.10	0.08	0.42	3.94	4.16	93.51
92	63.67	1.63	14.29	6.65	0.22	1.55	3.83	4.58	2.61	99.03
Mean	70.38	0.33	12.59	2.78	0.13	0.26	1.40	3.85	3.12	94.84
SD	3.10	0.58	0.85	1.76	0.04	0.57	1.15	0.38	0.75	2.06

Total iron is expressed as FeO.

The four ash horizons can be differentiated using the CaO:MgO ratios, as shown in Figure 5.8, enabling comparisons with the tephras from Mallachie and Craigmaud.

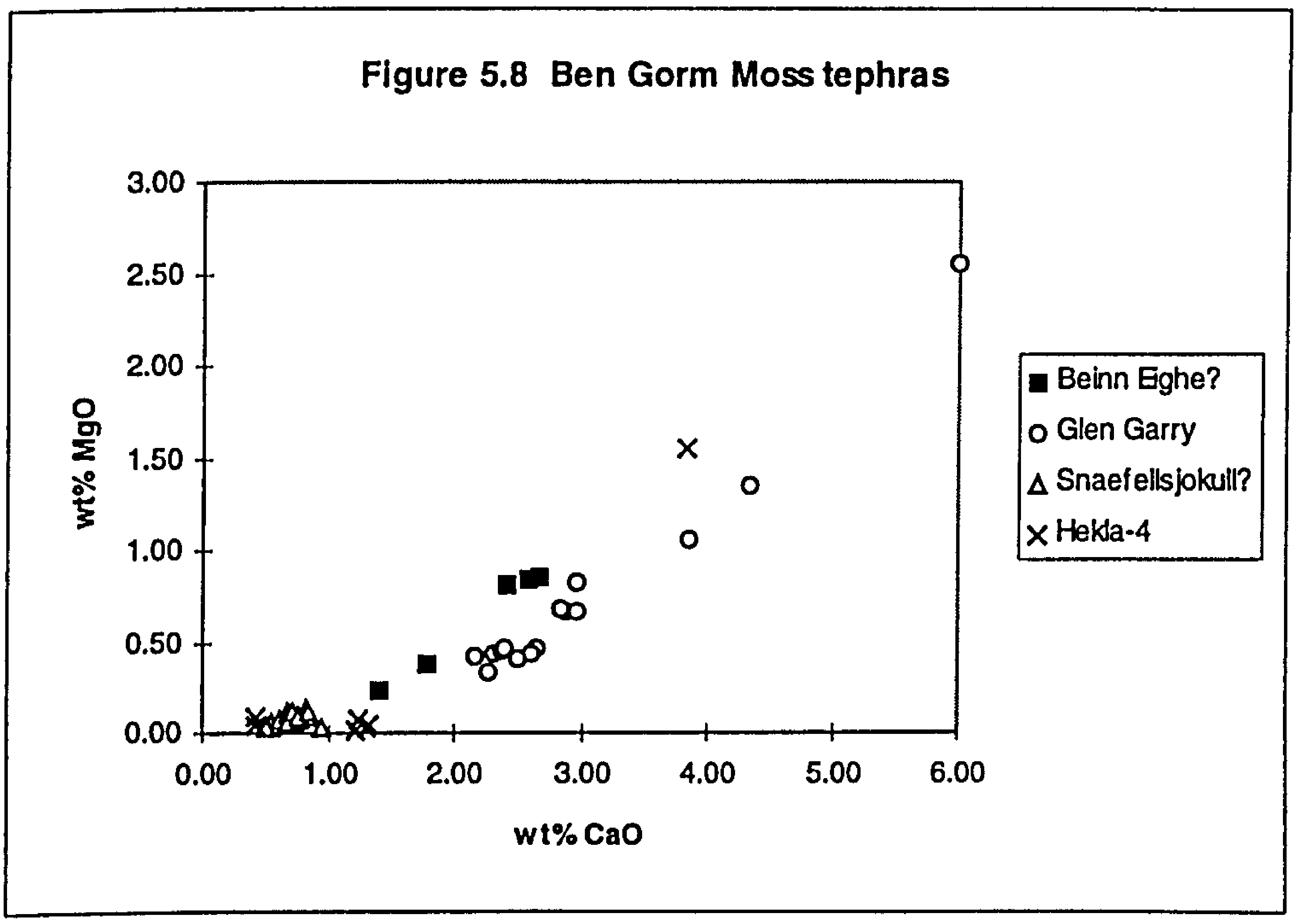


Figure 5.8 Ben Gorm tephras - the ratios of calcium and magnesium can be plotted to differentiate distinctive populations of the tephra isochrones.

The populations of Glen Garry (BGMT-2) and Hekla-4 (BGMT-4) tephras can be clearly delineated from Figure 5.8, as well as comparisons with the isochrones from other sites, Tephrabase, and Dugmore *et al.* (1995a). It is important to note, however, that some of the analyses for BGMT-4 do not initially appear to be from Hekla-4. The dacitic shard from analysis 92, although containing what appear to be very high values of TiO_2, MgO and CaO, compared with the rest of the analyses, does have geochemical ratios comparable with other dacitic and andesitic (*ca.* 59-63% silica) shards from the Hekla-4 type site at Kalfafell (KAL-Y) analysed on the MK V at Edinburgh (Dugmore *et al.*, 1992), which can be seen in Figure 5.9.

Less explicable are the shards from analyses 85 and 86, which have comparatively much lower CaO values and much higher K_2O values. The only comparable geochemical data on Tephrabase is that from the Lairg B tephra (Dugmore *et al.*, 1995; Pilcher *et al.*, 1996). As the Lairg A tephra is very similar to Hekla-4 geochemically, it could be suggested that BGMT-4 combines both Lairg layers. However, the radiocarbon dates from Ben Gorm (see Section 5.3) place a tephra isochrone at this depth within the age confines of Hekla-4 and not as old as the Lairg layers. Another possibility could be that there was an airfall event around the same time as Hekla-4, possibly from Hekla itself given that the Lairg A tephra is very similar geochemically to Hekla-4. Dwyer and Mitchell (1997) describe two tephra layers in western Ireland, which have been radiocarbon dated to around 4000 BP. Only one of the horizons yielded tephra of sufficient size to be geochemically analysed, and the indication was that the population (possibly two populations) did not match exactly the published geochemistry of Hekla-4. However, the geochemistry of these analyses has not yet been published, and therefore no comparisons can be made with the data from BGMT-4, although the occurrence of another ash layer around 4000 BP allows speculation that two populations may be present. Finally it is worth noting that any instrumental drift leading to these erroneous analyses is considered extremely unlikely, as the probe drift was constantly checked during analysis using an andradite standard.

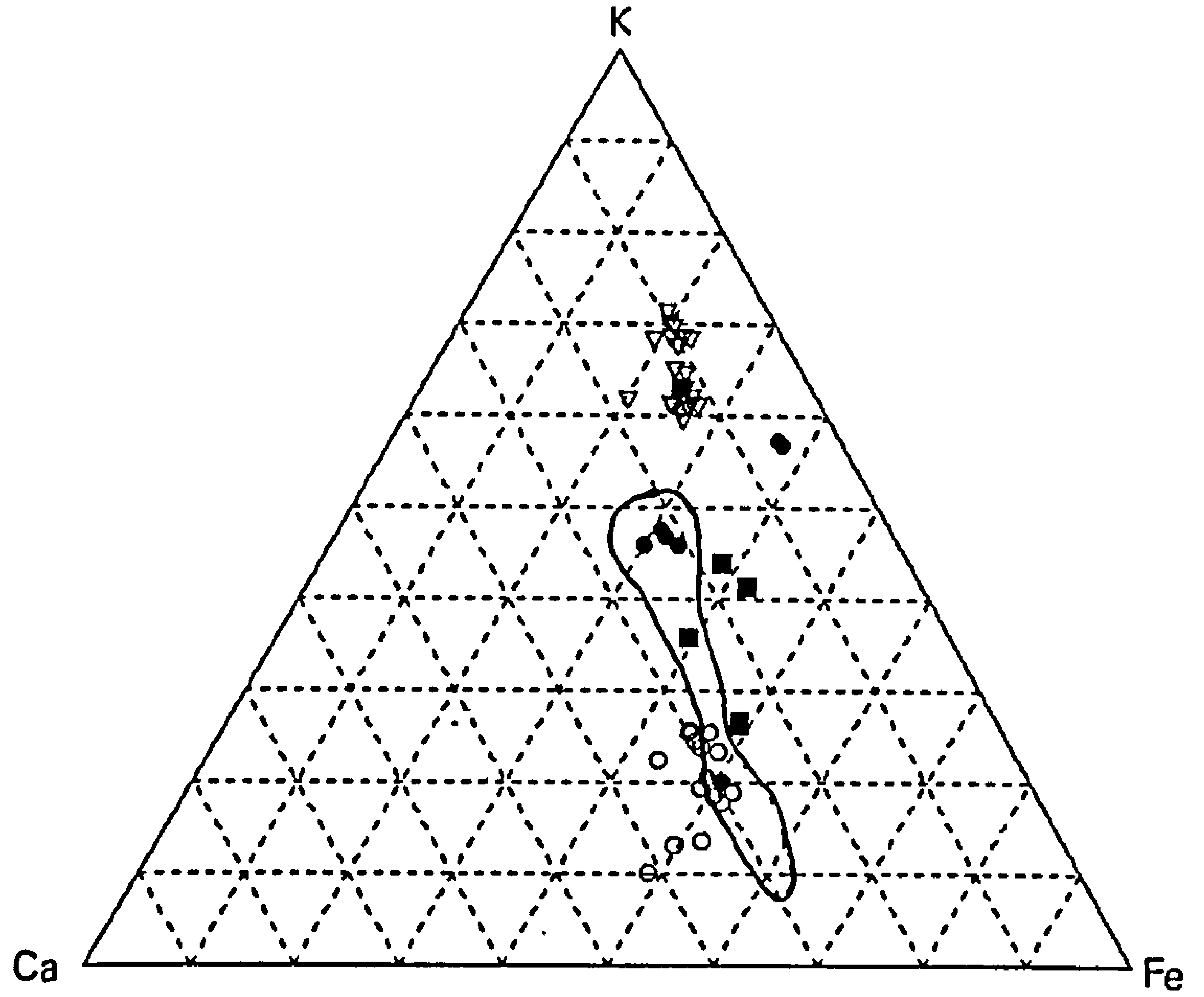


Figure 5.9 Ternary diagram showing the proportion of iron, potassium, and calcium oxides from glass shards at Ben Gorm Moss, Isle of Skye. Beinn Eighe(?) tephra is represented by squares, Glen Garry by clear circles, Snaefellsjökull(?) by triangles, and Hekla-4 by filled circles. The outline of the Hekla-4 population is shown (after Dugmore *et al.*, 1992).

BGMT-1 and BGMT-3 are more problematic. All the tephras found in this study which are not either Glen Garry or Hekla-4 isochrones (BGMT-1, BGMT-3, LAGT-1 and TEMT-2) are shown in Figures 5.10(a-c), which plots FeO against CaO, MgO, and TiO_2 respectively, in order to try and identify population structures within the data.

BGMT-1 can be clearly distinguished from the other tephras in Figures 5.10(a-c). It has a similar geochemistry to the Glen Garry tephra (although the lower limit of FeO is less in BGMT-1) with the major exception that potassium values for BGMT-1 are much higher. However, the other general similarities with the Glen Garry tephra suggest that these two isochrones may have the same volcanic origin. Another interpretation could be that BGMT-1 forms three separate populations (Figures 5.10), representing three separate eruptions. These may possibly have been from the same source, but ash from earlier eruptions lying on the slopes of the volcano may have been incorporated into an ash cloud

Figure 5.10(a-c) Comparison of tephras only found at one site.
Note BGMT-1 (blue) and BGMT-3 (green).

Figure 5.10(a) Binary Plot of FeO against CaO

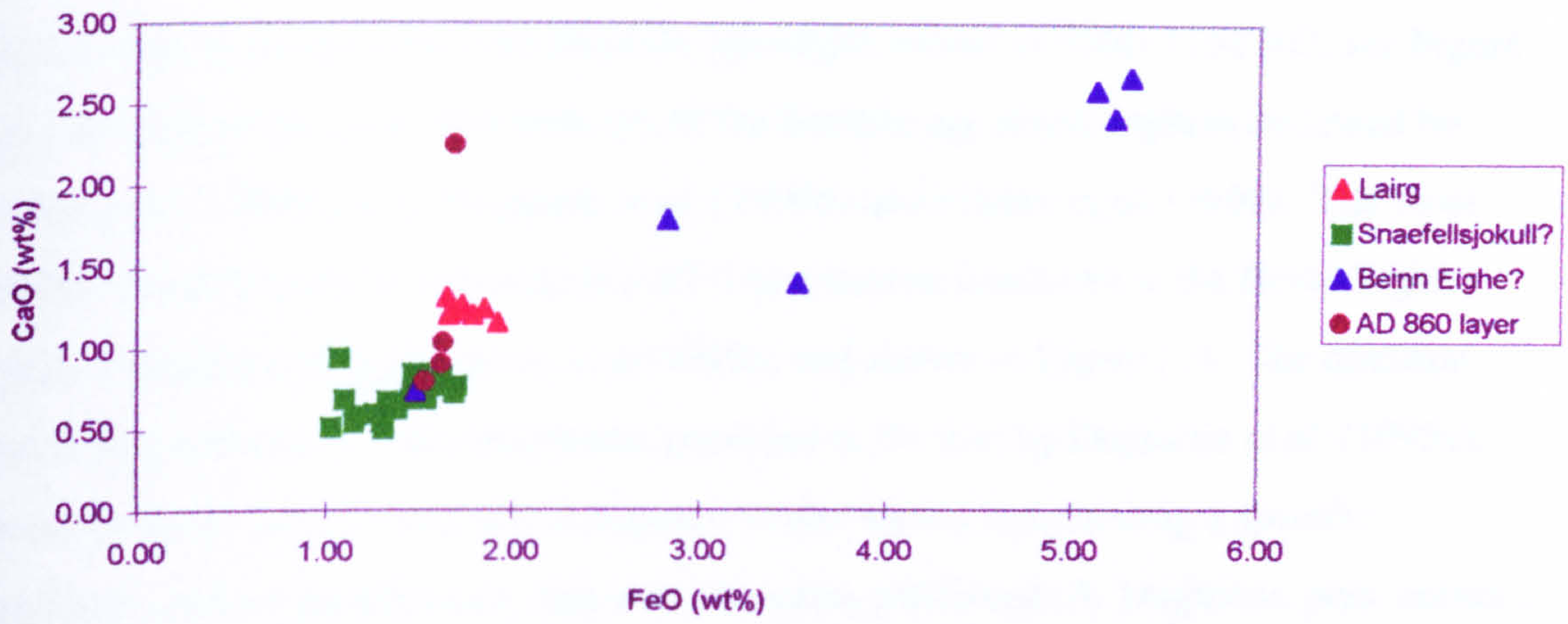


Figure 5.10(b) Binary plot of FeO against MgO

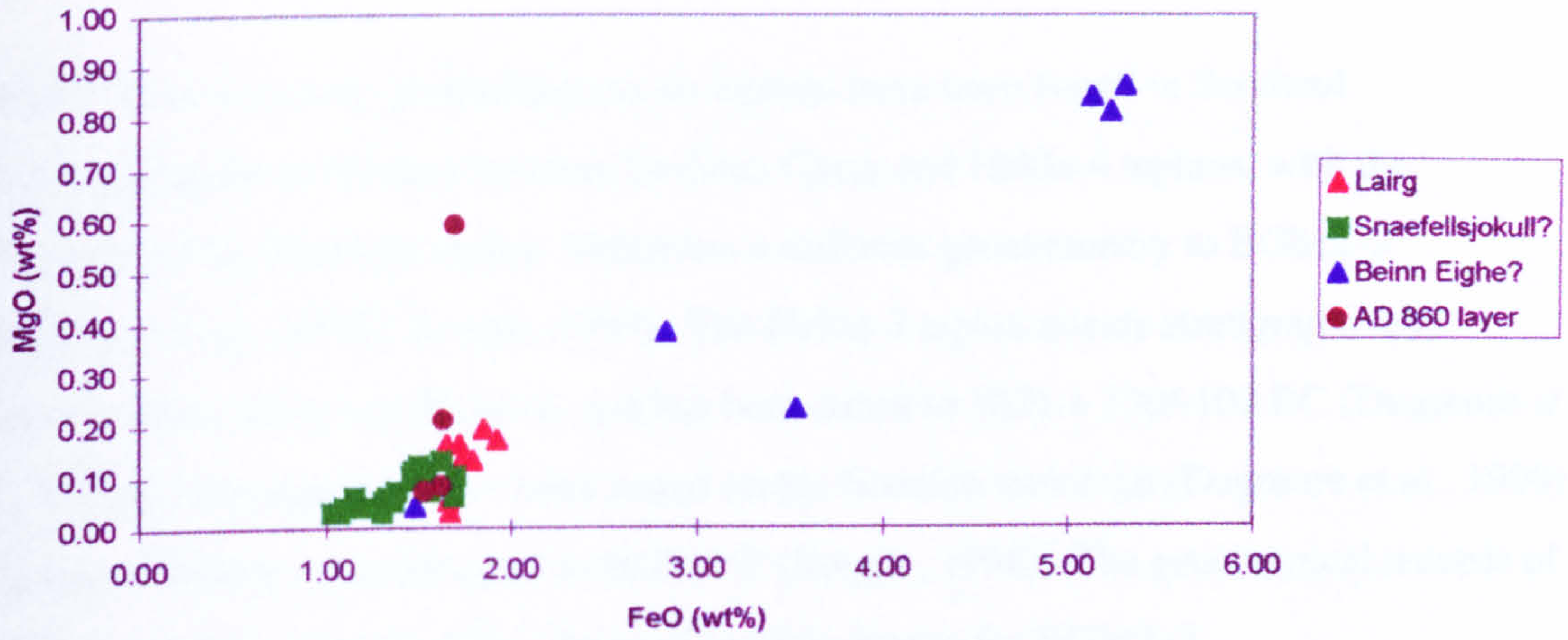
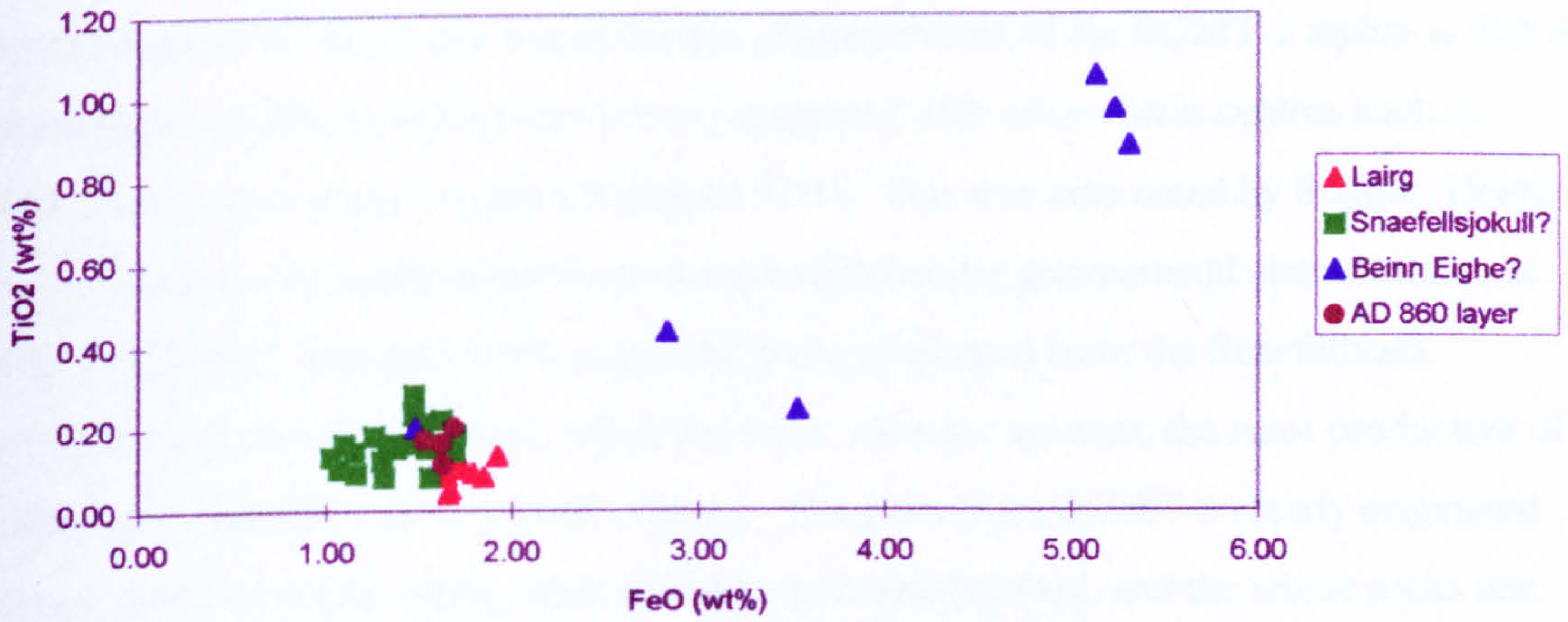


Figure 5.10(c) Binary plot of FeO against TiO2



of a later eruption, if the event was large enough. However, more data are necessary before the origins of this tephra can be fully interpreted. BGMT-1 is clearly of historic origin (with an interpolated date from the age-depth model of 1080-1150 AD, see Figure 5.21) but cannot be correlated with any of the historic-age silicic tephras reviewed by Larsen *et al.* (1999), or by Dugmore *et al.* (1995a) and Pilcher *et al.* (1996). The most realistic possibility for correlating BGMT-1 to a known isochrone is the Beinn Eighe tephra, documented by Dugmore et al (1995a) and shown in Figure 5.3. The question marks with reference to this isochrone, provided in the text by Dugmore *et al.* (1995a), were due to the lack of data and confidence in this tephra representing a specific isochrone, and no geochemical data were therefore published (A. Dugmore, pers. comm.). However, the occurrence of a tephra at similar stratigraphic depth, and the close proximity of the sites, suggests that BGMT-1 could be the Beinn Eighe tephra.

BGMT-3 is even more problematic as no tephras have been found in Scotland stratigraphically positioned between the Glen Garry and Hekla-4 tephras, with the exception of the Kebister tephra, which has a different geochemistry to BGMT-3 (Dugmore et al., 1995a; Boygle, 1998). The Hekla-3 tephra occurs stratigraphically between Glen Garry and Hekla-4, and has been dated to 1021 + 130/-100 BC (Dugmore *et al.*, 1995b) although it has not been found on the Scottish mainland (Dugmore et al., 1999) and has a different geochemistry to BGM3-T (Boygle, 1998). The geochemical records of Icelandic tephras were therefore the best possible source for BGMT-3.

The geochemistry of BGMT-3 is highly silicic (>71%) and displays a tight population cluster (Figures 5.10a-c), but one of the key characteristics of the BGMT-3 tephra is that it has alkali ratios (Na_2O:K_2O) close to one, compared with other silicic centres such as Hekla which have ratios around 1.5 (Figure 5.11). This was also noted by Boygle (1994) who identified a tephra from northern Iceland with similar geochemical characteristics to BGMT-3 (Sv2q). Boygle (1994) suggested Sv2q originated from the Snaefellsnes Volcanic Zone, western Iceland, which has three volcanic systems, the most productive of which is the Snaefellsjökull central volcano. The glass from BGMT-3 clearly originated from an evolved acidic centre, such as Hekla or Snaefellsjökull, and the silicic rocks and tephra from Snaefellsjökull have alkali ratios close to one (Imsland, 1978). Three major acidic tephras have been identified in the Snaefellsjökull area (Jóhannesson *et al.*, 1981), which are termed Sn-1, Sn-2, and Sn-3. Sn-1 and Sn-2 have been radiocarbon dated to 1750 ± 150 BP (cal. 100 BC - 650 AD) and 3960 ± 100 BP (cal. 2900 - 2100 BC) respectively, whilst Sn-3 has an estimated age of 7000 - 9000 BP (Steinthórsson, 1967). The age-depth model for Ben Gorm Moss (Figure 5.21) suggests an interpolated age of BGMT-3 of 420 BC. However, radiocarbon dating peats in Iceland is difficult, due mainly to disturbances following eruptions and tephra contamination (Boygle, pers. comm.), suggesting that the single dates taken from Snaefellsjökull could have an even larger range than quoted. The tephra with a date closest to BGMT-3 is Sn-1. However, no geochemical data have been published for any of the Snaefellsjökull tephras, and so a conclusive identification of BGMT-3 cannot be made until the geochemical data are available.

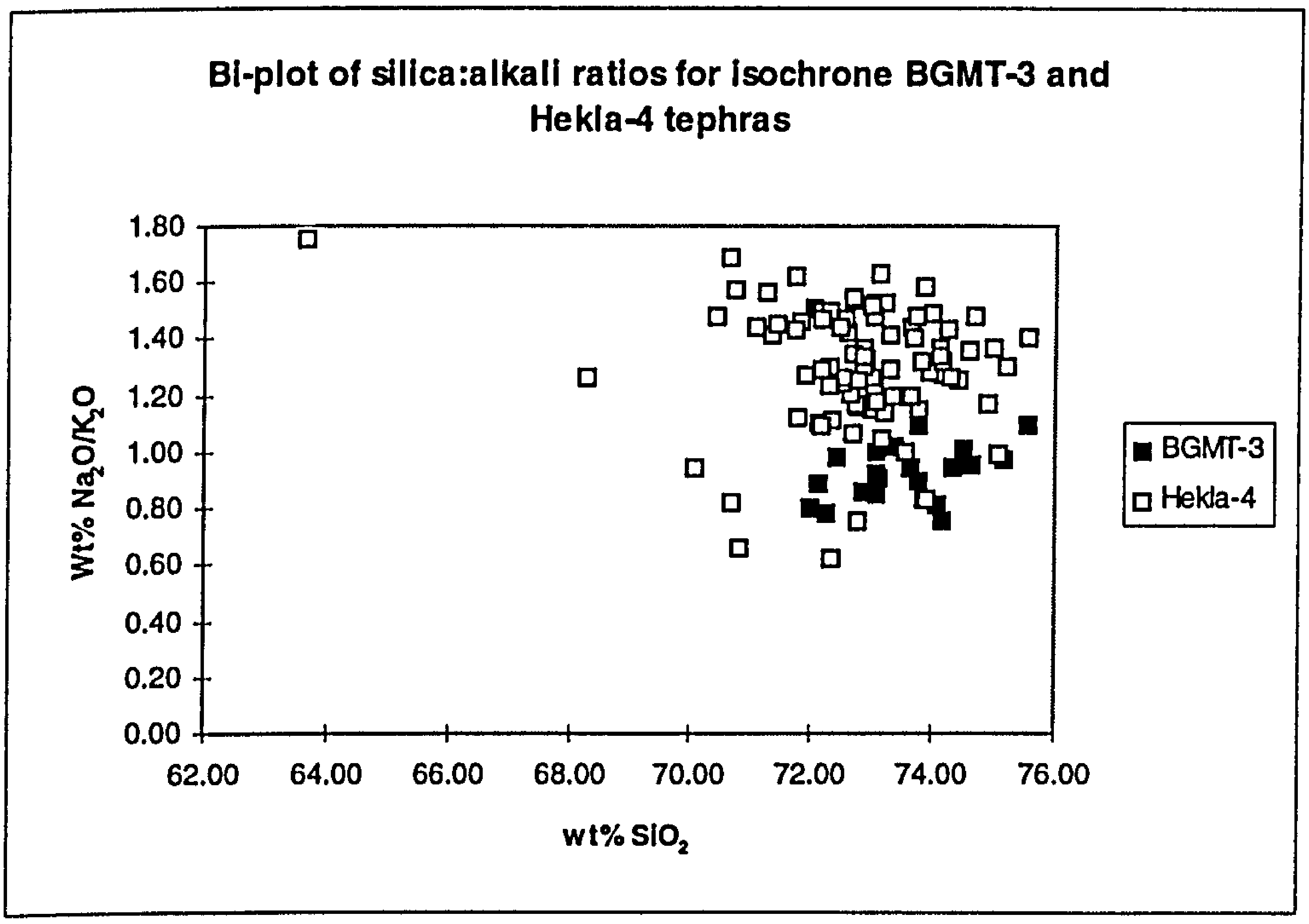


Figure 5.11 Bi-plot of silica:alkali ratios for Hekla-4 tephra (data from this thesis) and BGMT-3. Sample BGMT-3 has alkali ratios around 1, suggesting Snaefellsjökull as a possible source.

A summary of the data described above is shown in Table 5.16.

Table 5.16 Summary of the tephra isochrones identified and used in this study.

Site	Grid Reference	Depth (cm) to clay base	Tephra isochrone (cm)	Tephra isochrone
Longbridge Moss	NY 3053 5694	684	60-65	*Unknown*
			282-283	Hekla-4
Langlands Moss	NS 2634 6513	>800	30-31	AD 860 layer
			85-86	Glen Garry
Temple Hill Moss	NT 3113 6613	541	114-115	Glen Garry
			444-445	Lairg A
Shirgarton Moss	NS 2645 6963	471	55-60	*Unknown*
			134-135	Glen Garry
Mallachie Moss	NH 2963 8170	796	120-121	Glen Garry
			261-262	Hekla-4
Craigmaud Moss	NJ 3885 8588	>275	104-105	Glen Garry
			139-140	Hekla-4
Ben Gorm Moss	NG 1432 8658	137	30-31	Beinn Eighe?
			66-67	Glen Garry
			82-83	Snaefellsjokull?
			134-135	Hekla-4

5.2.4 Geochemical and spatial variability of tephra distribution

The geochemical variability of silica within a defined tephra isochrone allows inferences to be made about tephra dispersal. During major Icelandic eruptions the silicic content of the ash will decrease over time (Thorarinsson, 1967). These changes can be detected in the distal fallout material, allowing reconstructions of the path of the ash cloud to be established. This highlights the potential for locating certain isochrones within distinctive regions, and enables zones where fallout material is present to be mapped.

As silica content changes throughout a major Icelandic eruption, the less siliceous, dacitic phase of an eruption tends to occur towards the end of the eruptive phase. Larsen and Thorarinsson (1977) suggest that the silica content for Hekla-4 ranges between *ca.* 74% initially to 57% towards the end of the eruption. Thus, when dacitic shards are recovered from a tephra deposit, there is evidence that material from the later stages of the eruption reached the area where the ash was deposited. Whereas when only rhyolitic shards are recovered, deposition (and therefore probably dispersal) only occurred from the initial stages of the eruption. A plot of the Hekla-4 and Glen Garry SiO_2:K_2O ratio from each site is shown in Figure 5.12, indicating the relative amount of rhyolitic to dacitic proportions of ash can be traced across Scotland.

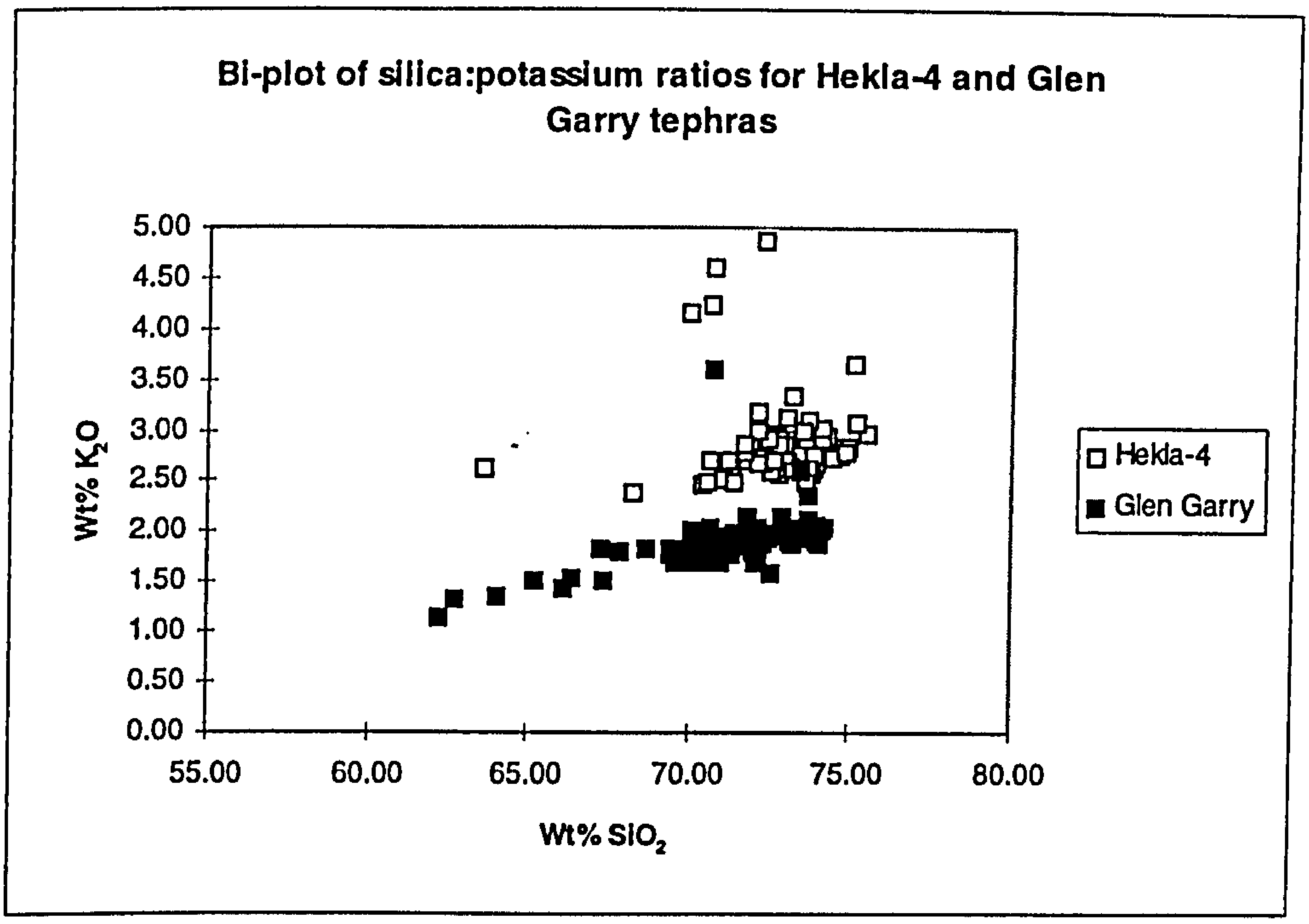


Figure 5.12 Bi plot of silica:potassium ratios for Hekla-4 and Glen Garry tephras (data from thesis). The relative dispersals of the tephras can be clearly seen.

An indication of tephra loading and range of percentage silica content for sites where Hekla-4 and Glen Garry tephras have been found are plotted in Figures 5.13 and 5.14. The dispersal pattern shown by Hekla-4 tephra in Figure 5.13 suggests that most areas in the east of Scotland were covered by the initial fallout, as the highest percentage of silica content at each site is either just over or under 75%. However, the west coast site at Ben Gorm Moss appears to have only low concentrations of tephra, suggesting that it may have been on the periphery of the fallout cloud.

A key feature shown in Figure 5.13 depicts a gap in the fallout pattern around central Scotland. Hekla-4 tephra was not recovered from the three sites in this region (SGM, LAG and TEM) despite extensive searches. No literature exists on Hekla-4 having been found in this region. Negative evidence, such as this, is by no means ideal, but here it is suggested as more than a coincidence. Hekla-4 tephra has been located in southern

Figure 5.13 Hekla 4:
Tephra loading and silicic content

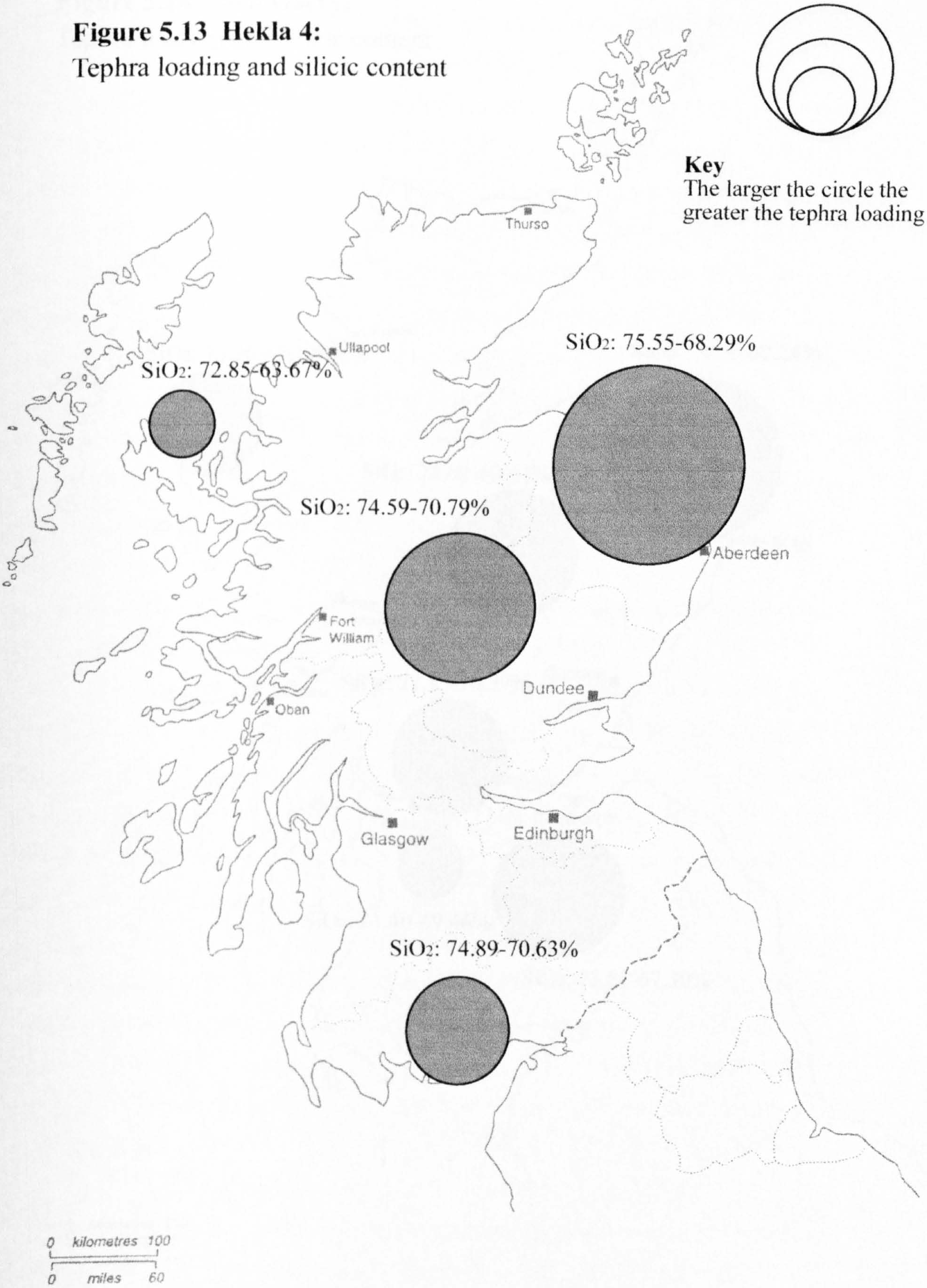
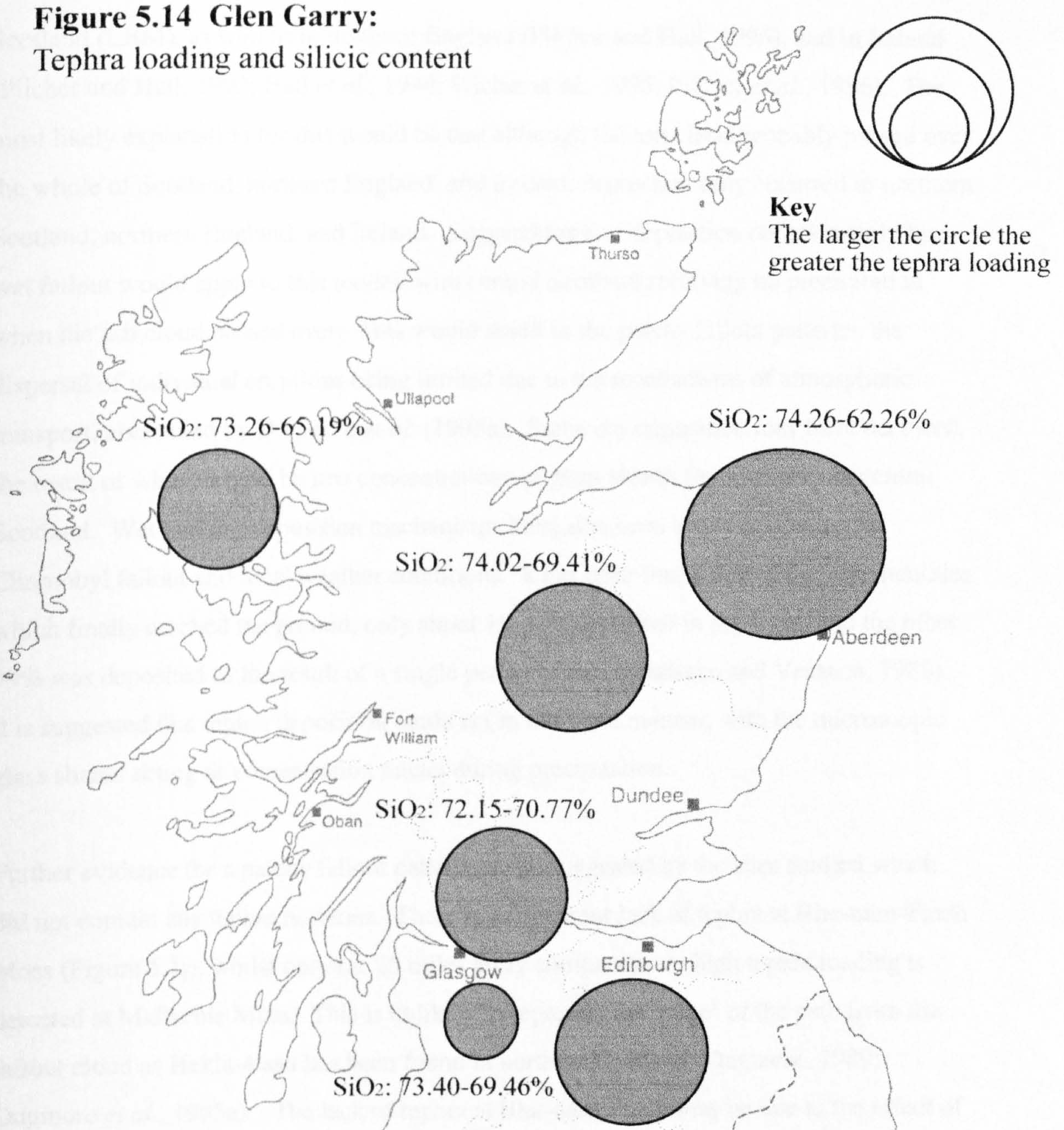


Figure 5.14 Glen Garry:
Tephra loading and silicic content



Scotland (LBM), as well as in northern England (Pilcher and Hall, 1996), and in Ireland (Pilcher and Hall, 1992; Hall *et al.*, 1994; Pilcher *et al.*, 1995; Pilcher *et al.*, 1996). The most likely explanation for this would be that although the ash cloud probably passed over the whole of Scotland, northern England, and Ireland, deposition only occurred in northern Scotland, northern England, and Ireland. Suggestions that deposition occurs mainly by wet fallout would apply to this model, with central Scotland receiving no precipitation when the ash cloud passed over. This would result in the patchy fallout patterns, the dispersal of individual eruptions being limited due to the mechanisms of atmospheric transport, also noted by Dugmore *et al.* (1995a). Some dry deposition may have occurred, the result of which would be low concentrations of glass shards found at sites in central Scotland. Wet and dry deposition mechanisms have also been tested in relation to Chernobyl fallout and local weather conditions. It has been found that of the radionuclides which finally reached the ground, only about 1% was deposited in dry form, and the other 99% was deposited as the result of a single period of rain (Mattsson and Vesanen, 1988). It is suggested that tephra deposition could act in the same manner, with the microscopic glass shards acting as condensation nuclei during precipitation.

Further evidence for a patchy fallout pattern can be suggested by the sites studied which did not contain any tephra horizons. There is an apparent lack of tephra at Blar-nam-Fiadh Moss (Figure 5.1j), whilst only *ca.* 25 miles away comparatively high tephra loading is detected at Mallachie Moss. This is unlikely to represent the 'edge' of the path from the fallout cloud as Hekla-4 ash has been found in northern Scotland (Dugmore, 1989; Dugmore *et al.*, 1995a). The lack of tephra at Blar-nam-Fiadh may be due to the effect of a rain shadow, produced by the Black Isle, which is situated to the northwest of the site. Additionally, no clearly defined tephra isochrones can be identified at Red Moss of Balerno (Figure 5.1h), whilst less than 5 miles away relatively high concentrations of tephra have been recovered from Temple Hill Moss (Figure 5.1e), possibly reflecting local variability in fallout.

However, certain characteristics pertaining to the dispersal patterns can be seen within this patchy nature of distribution. It is suggested that the initial extremely silicic phase of the Hekla-4 eruption reached the northeast corner of Scotland first, as shown by the variation in silica content between sites (Figure 5.13). Northern Skye has a low tephra loading, suggesting that this area was on the periphery of the ash cloud. The tephra from the main part of the eruption appears to have been carried over Scotland and northern England, towards Ireland, displaying anticyclonic behaviour, as suggested by Larsen and Thorarinsson (1977). The latter phase of the eruption may be interpreted from the range of dacitic shards which have been analysed. The tephra analysed from Craigmaud exhibits slight dacitic nature, and one shard has been found at Ben Gorm of only 63.67% silica, indicating that the latter phase of the eruption still reached northwest Scotland. Dacitic and some andesitic shards have been recovered from northern Scotland (Dugmore *et al.*, 1995a) and therefore confirm that the latter phase of the eruption can be detected in these areas.

The distribution of Glen Garry tephra dispersal is illustrated in Figure 5.14. The data suggests that most of Scotland was covered by the initial phase of the eruption. Glen Garry tephra has also been located on the Western Isles (Dugmore *et al.*, 1995a) as well as in northern England (Pilcher and Hall, 1996), but has not yet been reported in Ireland. Again, the initial phase of the eruption, indicated by the highest amounts of silicic glass, is found in the northeast part of Scotland, spreading further south during the eruption. This pattern appears to be replicated for the later phase of the eruption, although there is evidence for a change in dispersal patterns for the latter part of the eruption, as dacitic glass shards are found on the eastern part of Scotland, with the lowest silicic levels again being located in the northeastern region. It would therefore appear that the circulation patterns changed throughout the eruption, from a more northerly airflow over Scotland to a more northeasterly flow by the end of the eruption. The tephra loading data corroborates this explanation, with the highest loading occurring in the north and northeast, and the lowest amounts in west central Scotland.

5.2.5 Tephrochronology summary

Fourteen ash layers have been geochemically analysed from seven ombrogenous mires in Scotland. At least six different volcanic eruptions have been geochemically characterised, the most common of which are the Glen Garry and Hekla-4 tephras. These isochrones can be used to correlate precisely between the palaeoecological and palaeoclimatological records derived from the peat stratigraphies, and from other sites in northwest Europe where the isochrones have been identified.

5.3 Radiocarbon dating

Growth rates of peat are not solely governed by the rate of aerobic decay of peat, but more so by the length of time over which it operates. Radiocarbon dates are simply the average of these rates (Clymo, 1978). Variation in accumulation rates will occur over short timescales due to the prevailing climatic conditions (Stoneman, 1993). However, modelling this variability is extremely difficult, and rather than trying to reconstruct each minor variation an average has been sought. Fitting a trend line to the data allows the models to effectively smooth this variability over the longer timescales, and with a good fit (high r^2 value) it is the best estimate of average rates of accumulation (Barber *et al.*, 1994b).

Table 5.17 Radiocarbon dates (37) with the full calibrated 2 sigma range.

Lab. No.	Site	Depth below peat surface (cm)	^{14}C date BP	δ^{13}C‰	Calibrated range AD/BC[*] (2 sigma)	Mid-point of the 2 sigma calibrated range
SRR-6324	Longbridge Moss	32-40	625 ± 40	-26.7	1280-1400 AD	1340 AD
SRR-6325	Longbridge Moss	44-52	890 ± 45	-26.9	1030-1230 AD	1130 AD
SRR-6326	Longbridge Moss	60-68	1000 ± 45	-26.9	950-1170 AD	1060 AD
SRR-6327	Longbridge Moss	108-116	1830 ± 45	-26.9	70-330 AD	200 AD
SRR-6328	Longbridge Moss	140-148	2265 ± 45	-26.1	400-200 BC	300 BC
SRR-6329	Longbridge Moss	208-216	3680 ± 45	-25.9	2200-1930 BC	2065 BC
SRR-6330	Longbridge Moss	248-256	3585 ± 45	-26.8	2130-1780 BC	1955 BC
SRR-6397[*]	Longbridge Moss	228-236	3600 ± 45	-27.0	2140-1820 BC	1980 BC
SRR-6331	Langlands Moss	16-24	1055 ± 40	-27.5	890-1030 AD	960 AD
SRR-6332	Langlands Moss	52-60	1500 ± 45	-27.7	440-640 AD	540 AD
SRR-6333	Langlands Moss	64-72	1755 ± 45	-27.2	140-390 AD	265 AD
SRR-6334	Langlands Moss	168-176	2920 ± 45	-26.8	1300-990 BC	1145 BC
SRR-6335	Langlands Moss	240-248	3240 ± 50	-27.2	1670-1420 BC	1545 BC
SRR-6336	Temple Hill Moss	68-76	1885 ± 40	-25.6	20-230 AD	125 AD
SRR-6337	Temple Hill Moss	180-188	3065 ± 40	-31.9	1430-1220 BC	1325 BC
SRR-6338	Temple Hill Moss	252-260	4155 ± 40	-25.5	2890-2600 BC	2745 BC
SRR-6339	Temple Hill Moss	328-336	4600 ± 45	-27.7	3510-3100 BC	3305 BC
SRR-6340	Temple Hill Moss	416-424	5855 ± 45	-27.1	4900-4600 BC	4750 BC
SRR-6341	Shirgarton Moss	31-33	725 ± 40	-25.4	1220-1390 AD	1305 AD
SRR-6342	Shirgarton Moss	72-80	795 ± 45	-25.7	1120-1290 AD	1205 AD
SRR-6343	Shirgarton Moss	120-128	1960 ± 40	-27.5	90 BC-130 AD	20 AD
SRR-6344	Shirgarton Moss	196-204	2630 ± 45	-33.5	910-760 BC	835 BC
SRR-6345	Shirgarton Moss	240-248	3010 ± 45	-26.8	1410-1120 BC	1265 BC
SRR-6346	Mallachie Moss	20-28	105.60 ± 0.60% mod C	-26.2	*Out of calibration range*	
SRR-6347	Mallachie Moss	44-52	1260 ± 40	-26.9	660-880 AD	770 AD
SRR-6348	Mallachie Moss	96-104	1470 ± 45	-27.5	450-660 AD	555 AD
SRR-6349	Mallachie Moss	149-155	2795 ± 45	-26.6	1060-830 BC	945 BC
SRR-6350	Mallachie Moss	212-220	3470 ± 45	-27.1	1910-1680 BC	1795 BC
SRR-6351	Craigmaud Moss	11-19	390 ± 45	-27.6	1430-1640 AD	1535 AD
SRR-6352	Craigmaud Moss	24-32	315 ± 40	-26.9	1470-1650 AD	1560 AD
SRR-6353	Craigmaud Moss	68-76	1225 ± 40	-27.7	680-890 AD	785 AD
SRR-6354	Craigmaud Moss	76-84	1385 ± 40	-27.9	560-700 AD	630 AD
SRR-6355	Craigmaud Moss	120-128	2800 ± 40	-28.2	1050-840 BC	945 BC
SRR-6356	Ben Gorm Moss	15-17	290 ± 45	-27.7	1470-1800 AD	1635 AD
SRR-6357	Ben Gorm Moss	49-51	1475 ± 40	-29.3	450-660 AD	555 AD
SRR-6358	Ben Gorm Moss	71-73	2155 ± 40	-27.5	370-100 BC	235 BC
SRR-6359	Ben Gorm Moss	88.5-91.5	2605 ± 45	-28.5	900-550 BC	725 BC
SRR-6360	Ben Gorm Moss	121-123	3225 ± 45	-28.6	1620-1420 BC	1520 BC

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[*] Check date on apparent aberrant age from SRR 6330, kindly provided by Brian Miller, NERC Radiocarbon Laboratory, East Kilbride.

The age-depth models for each site are shown in Figures 5.15-5.21, with the full 2 sigma ranges (Dumayne *et al.*, 1995) and depth ranges of the radiocarbon dates shown. The tephra isochrones are used as pinning points for the radiocarbon dates, and the pine rise/SCP chronology are used to aid the chronological control for the last 200 years. Pine pollen was counted from the same slides used for testate amoebae analysis, with sampling intervals therefore only every 8cm, in a crude attempt to reconstruct the pine rise of *ca.* 1800 AD in the Scottish Borders (see Mauquoy and Barber, 1999a). SCP analysis has been used at Shirgarton Moss as only a subtle pine rise could be detected. It is suspected that the planting of pine was very localised in the central belt of Scotland at *ca.* 1800 AD, hence the detection of a pine rise from testate amoebae slides at these sites could be ambiguous (R. Tipping, pers. comm.).

5.3.1 Longbridge Moss

Figure 5.15 shows the age-depth model for Longbridge which assumes a linear relationship with a roughly even accumulation rate of *ca.* 15 years/cm, implying the same order of magnitude for peat growth as the ombrotrophic bogs of Cumbria (Barber *et al.*, 1994; Mauquoy and Barber, 1999a). Two of the dates (SRR-6329 and SRR-6330) show the same radiocarbon age, although with a difference of 40cm stratigraphically, indicating that one of the dates is probably inaccurate. The stratigraphic position of the Hekla-4 tephra agrees with the linear model presented in Figure 5.15, and would clearly not fit into a model with date SRR-6329 being accurate. The precision of the Hekla-4 tephra therefore pins date SRR-6330 to the linear model, and suggests that SRR-6329 is too old. Therefore, although SRR-6329 is plotted in Figure 5.15 it is not included in the age-depth model. SRR-6329 could be much older than the model suggests due to the reservoir effect (Kilian *et al.*, 1995), as an average of *ca.* 7% Ericaceae rootlets are present at that depth. However, the date is so much older than the model predicts that the NERC Radiocarbon Laboratory at East Kilbride offered to analyse another sample to further test the model (Brian Miller, pers. comm.). The new date, however, was also older than predicted by the age/depth model, possibly also due to the reservoir effect as levels of Ericaceae are still high throughout this zone around 10% (see Figure 6.4).

The pine rise position at 24cm is generally in agreement with the model, although the low resolution sampling interval of 8cm infers that 1800 AD could be nearer to 16cm than 24cm (*cf.* Mauquoy and Barber, 1999a). The top 10cm of the stratigraphy will not be used in any time series analysis as this is composed almost entirely of fresh *Sphagnum*, as yet unhumified, and would therefore have a much faster accumulation rate than the rest of the profile.

5.3.2 Langlands Moss

The age-depth model for Langlands Moss is shown in Figure 5.16. A second order polynomial model best fits the data with an r^2 value of 0.988, and the Glen Garry and AD 860 tephra pin the radiocarbon dates to the model. This gives average accumulation rates of between *ca.* 11 years/cm for the lower section and *ca.* 16 years/cm for the upper peat.

The AD 860 tephra is located at 30-31cm depth, inferring that the last 1000 years of accumulation occurred within the top 30 cm, which is clearly unrealistic. Thus, either a hiatus exists within the stratigraphy, or the top of the mire has been cut/damaged. The extinction of *S. imbricatum* occurs between 16-20cm, which has been radiocarbon dated to between 890-1030 AD, and is in agreement with the age-depth model. Within the two sigma radiocarbon range, this date is coherent with *S. imbricatum* extinction dates studied by Stoneman *et al.* (1993) and Mauquoy and Barber (1999b). The extinction is also associated with a wet shift, identified by the *S.s. Cuspidata* phase described in Section 6.72; a feature which has also been described by Mauquoy and Barber (1999a). It is therefore suggested that this part of the stratigraphy is still intact.

The top 10cm of the stratigraphy have unusually high amounts of UOM (see Figure 6.13, Section 6.7) which could indicate that this section has been disturbed. As parts of the mire have previously been forested and drainage ditches dug (see Section 4.2), it is possible that the top layers of the peat have compacted due to water loss, and may also have lost material in subsequent phases of drying out and desiccation. This is therefore suggested as the explanation for the lack of peat covering the last 1000 years. The model used for time series analysis will only contain data from the *S. imbricatum* extinction and below.

5.3.3 Temple Hill Moss

Figure 5.17 shows the age-depth model from Temple Hill Moss with a second order polynomial fit, which pins the radiocarbon dates to the Lairg and Glen Garry tephras, and gives an average accumulation rate of *ca.* 16.5 years/cm. The pine rise at 24 cm ensures that the chronology at the top of the mire is in accordance with the rest of the profile. The top 10cm will not be used in any time series analysis due to the increased accumulation rates for this part of the stratigraphy.

5.3.4 Shirgarton Moss

The stratigraphically lower dates from the profile and the Glen Garry tephra agree well with each other chronologically, and suggest a linear interpretation of peat growth. However, the upper two dates appear discordant with each other and hence SCP analysis was used to provide a check on the more recent accumulation, and also to investigate whether the bog had been cut at any point. The results from the SCP analysis are illustrated in Figure 5.22, and clearly show an initial take off period around 28-24cm, the rapid increase between 20-18cm, and the peak in concentration at 16cm, which can be correlated with the dates in Table 3.2 from Rose *et al.* (1995). When plotted on Figure 5.18 with the rest of the radiocarbon and tephra dates a second order polynomial can be fitted to the age-depth model with an r^2 of 0.98. This shows good agreement between the model and the dates, and gives an average accumulation rate of *ca.* 14 years/cm. The top 16 cm of peat is almost entirely composed of unhumified *S. magellanicum* and, from the SCP evidence, covers the last 20 years. Hence, a vastly increased growth rate can be inferred for the top of the profile, and will therefore not be used in any time series analysis.

However, the problem of the two most recent dates disagreeing and being bisected by the polynomial model still exists. A polynomial fit through the date between 31-33 cm (SRR-6341: 1220-1390 AD) would suggest a cessation of peat growth through the MWP. This level also marks the extinction of *S. imbricatum* (see Figure 6.31) and the date agrees with the extinction evidence from the sites studied by Stoneman *et al.* (1993) and Mauquoy and Barber (1999b). However, peat growth should have increased again after the MWP in the LIA due to an onset of wetter, cooler conditions, and thus a date slightly younger than this would be expected. It is therefore argued that bisecting these dates smoothes the potentially high variability of accumulation over these times, and that date SRR-6342 could be too young (probably due to root contamination of younger carbon), whereas date SRR-6341 may be slightly too old.

5.3.5 Mallachie Moss

A second order polynomial ($r^2 = 0.967$) was found to give a slightly better fit than a linear interpolation ($r^2 = 0.963$) for the age-depth model of Mallachie Moss (Figure 5.19), giving an average accumulation rate of *ca.* 16 years/cm. The stratigraphically lower radiocarbon dates agree well with the tephra dates according to the model, although the upper two dates are bisected by the polynomial fit, in a manner similar to Shirgarton Moss. It is thought that a hiatus occurs in the stratigraphy somewhere between 25-45cm, as there is an exceptionally high amount of UOM, charcoal and *Cenococcum* present between these depths (see Figure 6.40, Section 6.10). In the field it was noticed that this layer of high UOM is spatially extensive over the mire, with only fresh, comparatively unhumified *Sphagnum* overlying it. It is assumed that the overlying *Sphagnum* is of comparatively recent origin, and that the upper layers of stratigraphy have been disturbed, either by burning events and subsequent desiccation, or by peat cutting. The time of the disturbance is ambiguous, and therefore time series analysis will only utilise data from 50cm and below to avoid any potential errors.

5.3.6 Craigmaud Moss

The age-depth relationship for Craigmaud is shown in Figure 5.20, and a second order polynomial ($r^2 = 0.986$) is used to fit the radiocarbon dates with the tephra isochrones. Accumulation rates vary throughout the stratigraphy, from *ca.* 17 years/cm in the upper part of the stratigraphy to the slower rates of *ca.* 25 years/cm in the lower parts.

The upper part of the stratigraphy exhibits accumulation rates of the same magnitude as the Cumbrian mires studied by Mauquoy (1997), although there appears to be a hiatus in accumulation somewhere below 120cm as accumulation rates are much slower. This may be a natural cessation in growth due to the comparative aridity of the site (possibly an indication of Sub-boreal dryness, Godwin, 1954), although large levels of charcoal are present from 85cm to the base, suggesting periodic burning of the mire. Time series analysis will therefore only utilise data covering the stratigraphy between 4-100cm.

5.3.7 Ben Gorm Moss

A linear plot ($r^2 = 0.994$) best fits the age-depth model for Ben Gorm Moss, with an average accumulation rate of *ca.* 30 years/cm (Figure 5.21). The uncharacterised tephras around 30-31cm and 82-83cm (possibly Beinn Eighe and Snaefellsjökull respectively) have been added to Figure 5.21 from interpolating their ages with respect to the age-depth model, and so are plotted as a function of the model. Peat hagging at the surface may have caused some shrinkage, and therefore the top 10cm are omitted from the data used for time series analysis.

The rate of accumulation at Ben Gorm appears to be comparable with the Scottish west coast blanket mires studied by Anderson (1998) and Moine Mhor in the Cairngorms (Barber *et al.*, 1999). However, it appears faster than at Talla Moss in the Scottish Borders (Chambers *et al.*, 1997a), which is probably due to the hyper-oceanic climate of the site. Anderson (1998) adopts an approach for adjusting the age-depth models using humification data, although this is not attempted here as even though the peat may accumulate faster in a wet phase (lower humification) it is difficult to date the onset of a dry phase (higher humification) as a cessation in growth may have occurred.

5.4 Conclusions

Tephrochronology can be used to aid chronological control, and 'pin' specific radiocarbon dates to an age-depth model for peat growth. An age-depth model has been constructed for each of seven ombrotrophic mires, and a linear or second order polynomial trend line fitted, in order to interpolate the dates of specific levels. These data have been used to construct time series, for analysis with the palaeoecological data in Chapter 6.

Figure 5.15 Longbridge Moss age/depth model.

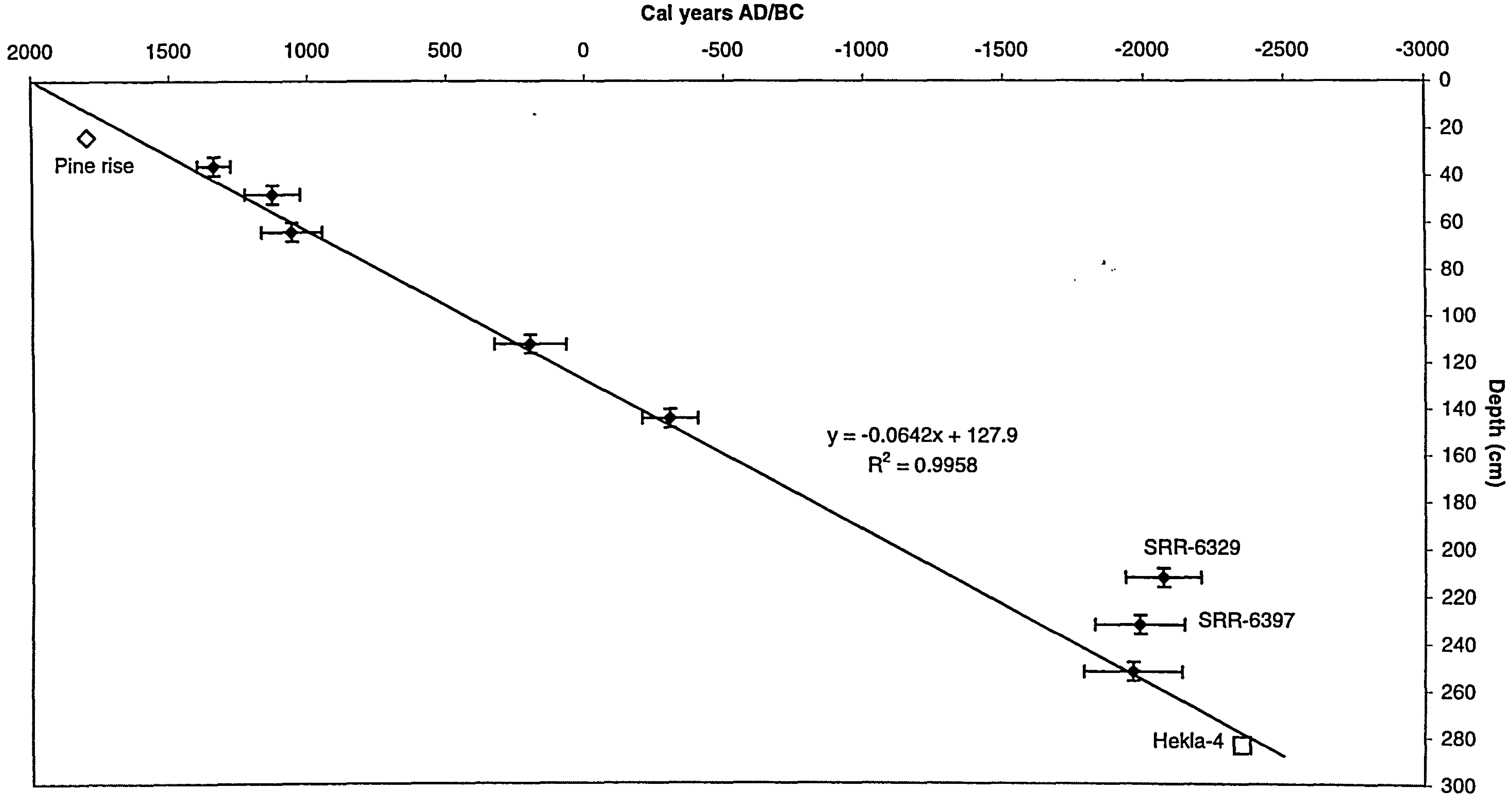


Figure 5.16 Langlands Moss age/depth model.

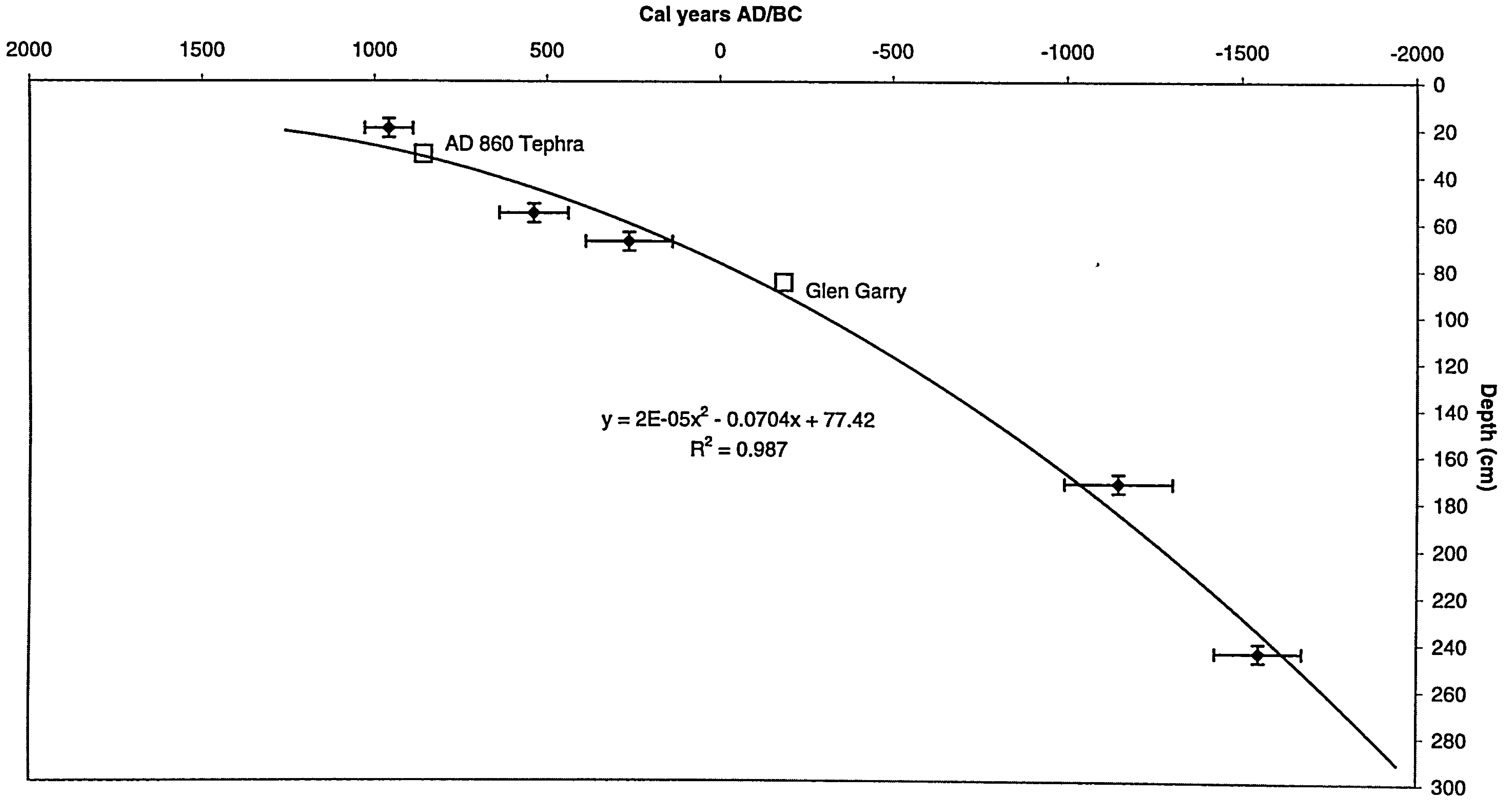


Figure 5.17 Temple Hill Moss age/depth model.

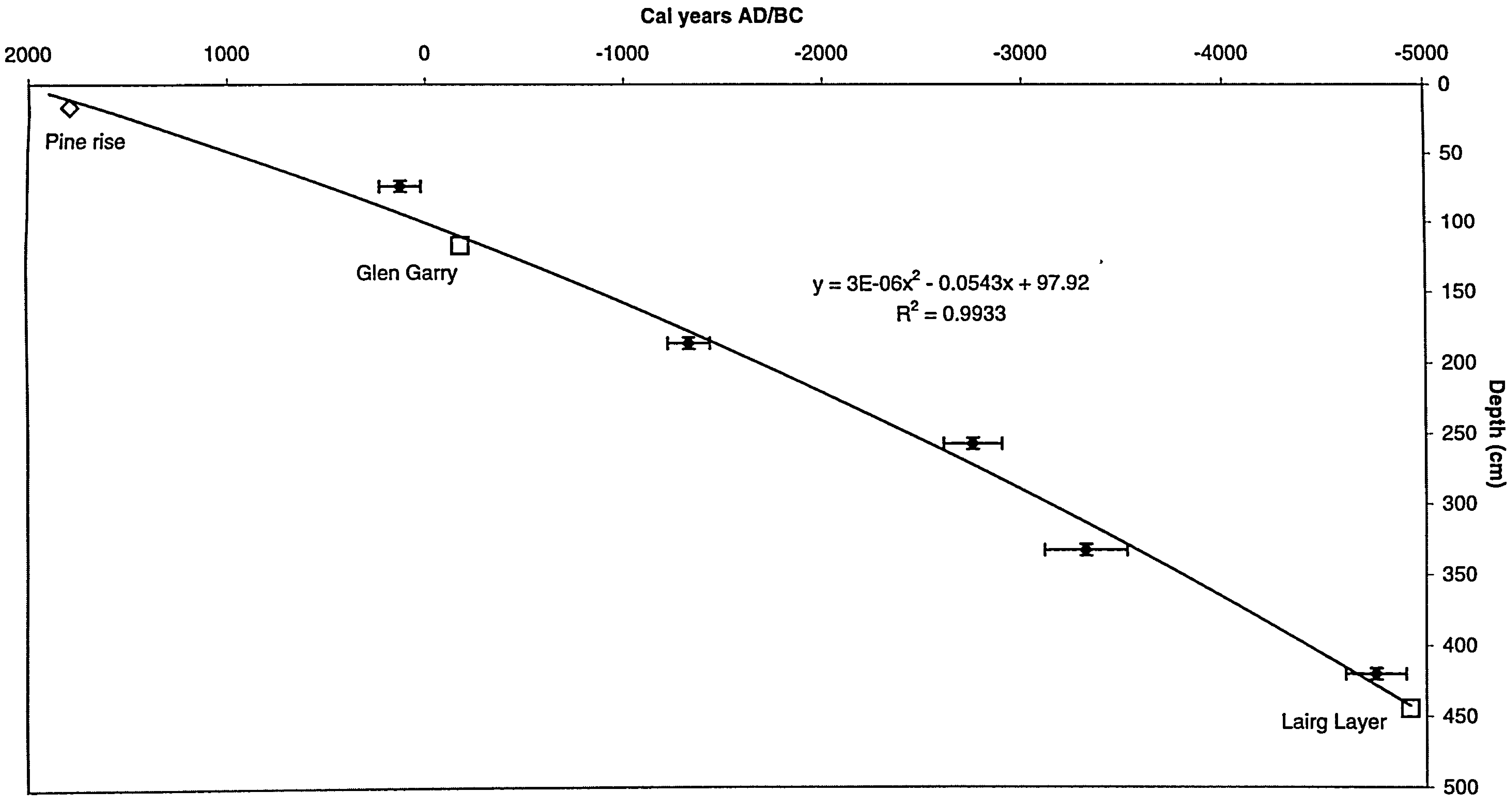


Figure 5.18 Shirgarton Moss age/depth model.

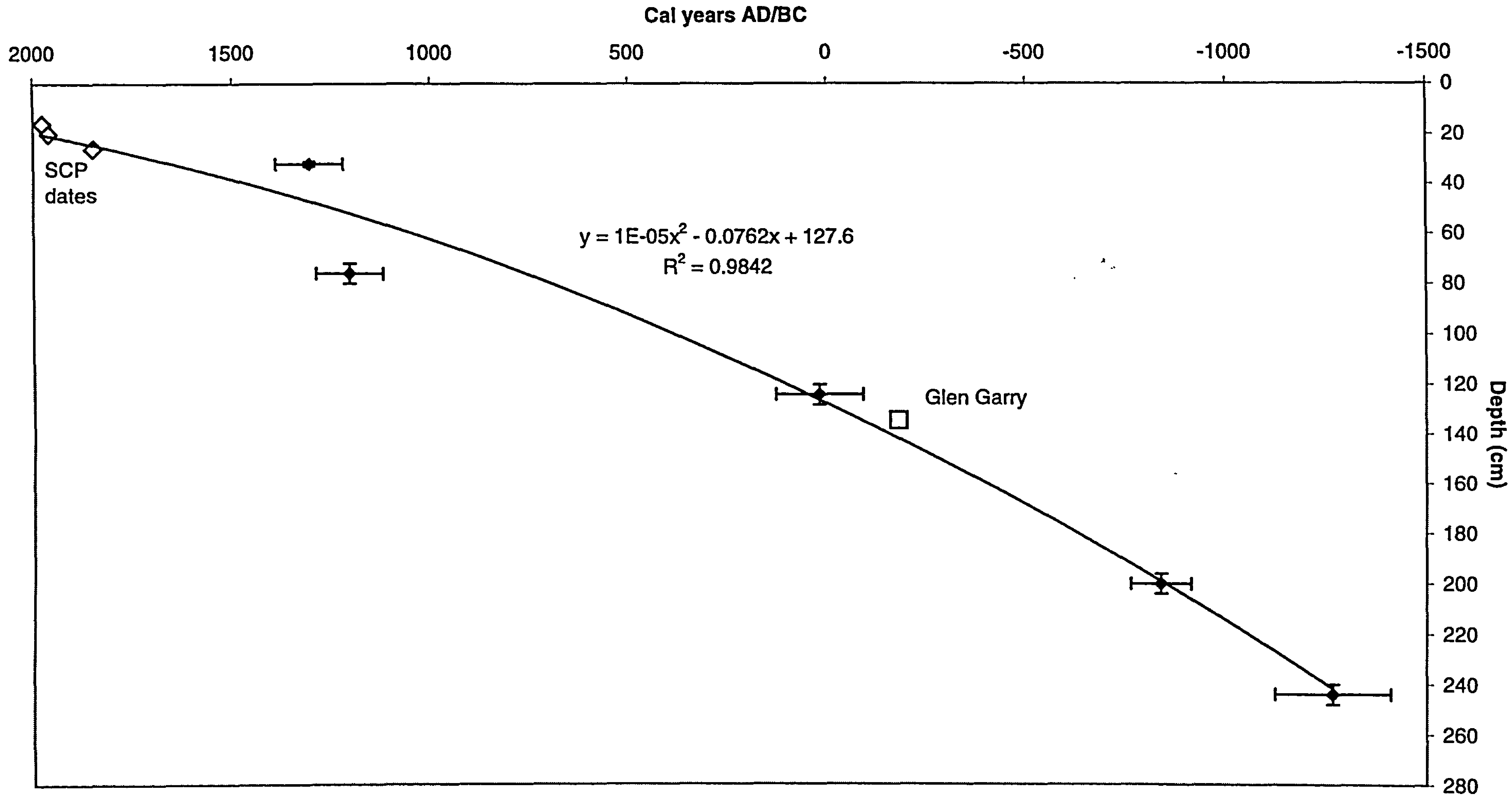


Figure 5.19 Mallachie Moss age/depth model.

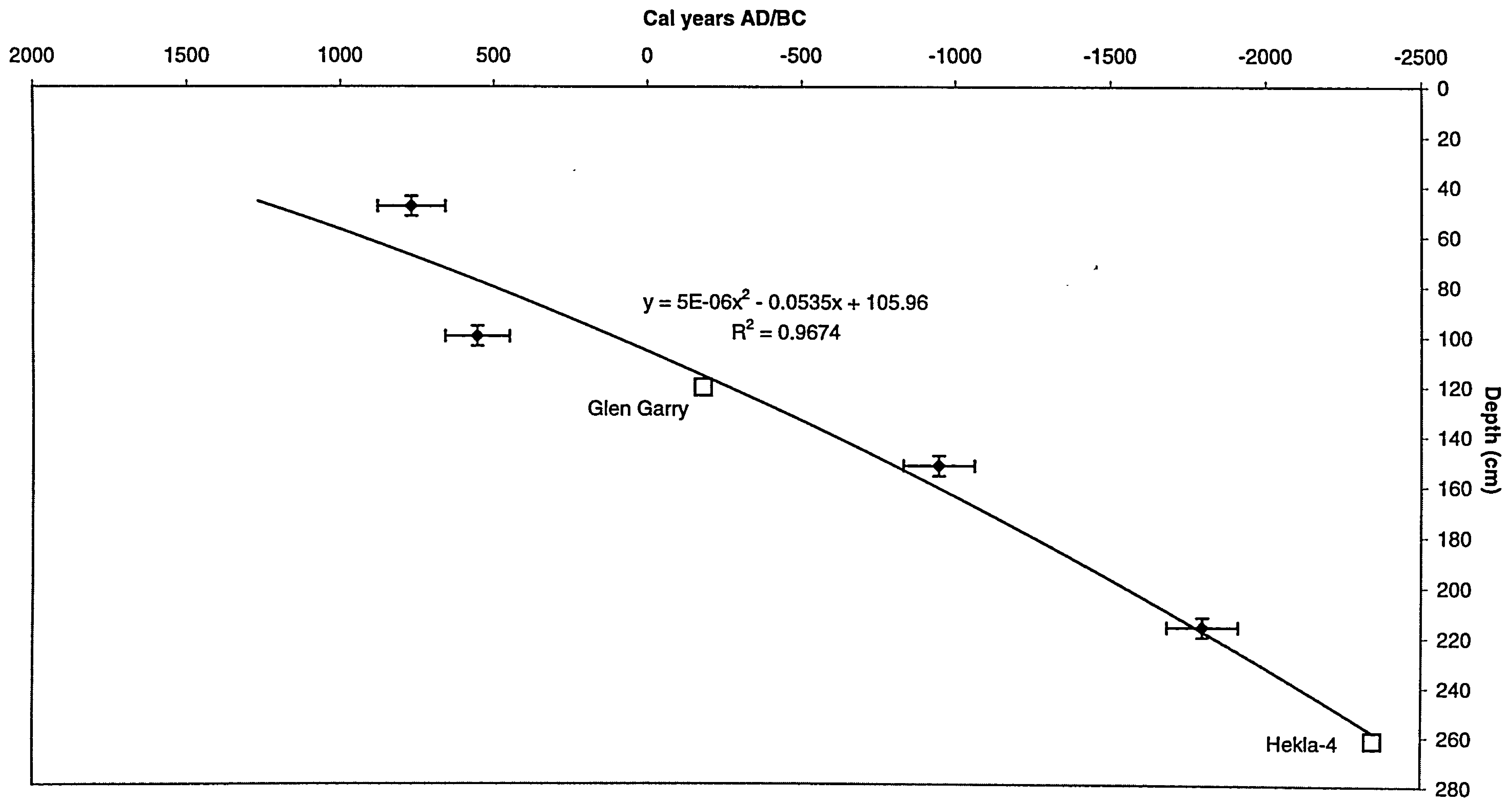


Figure 5.20 Craigmaud Moss age/depth model.

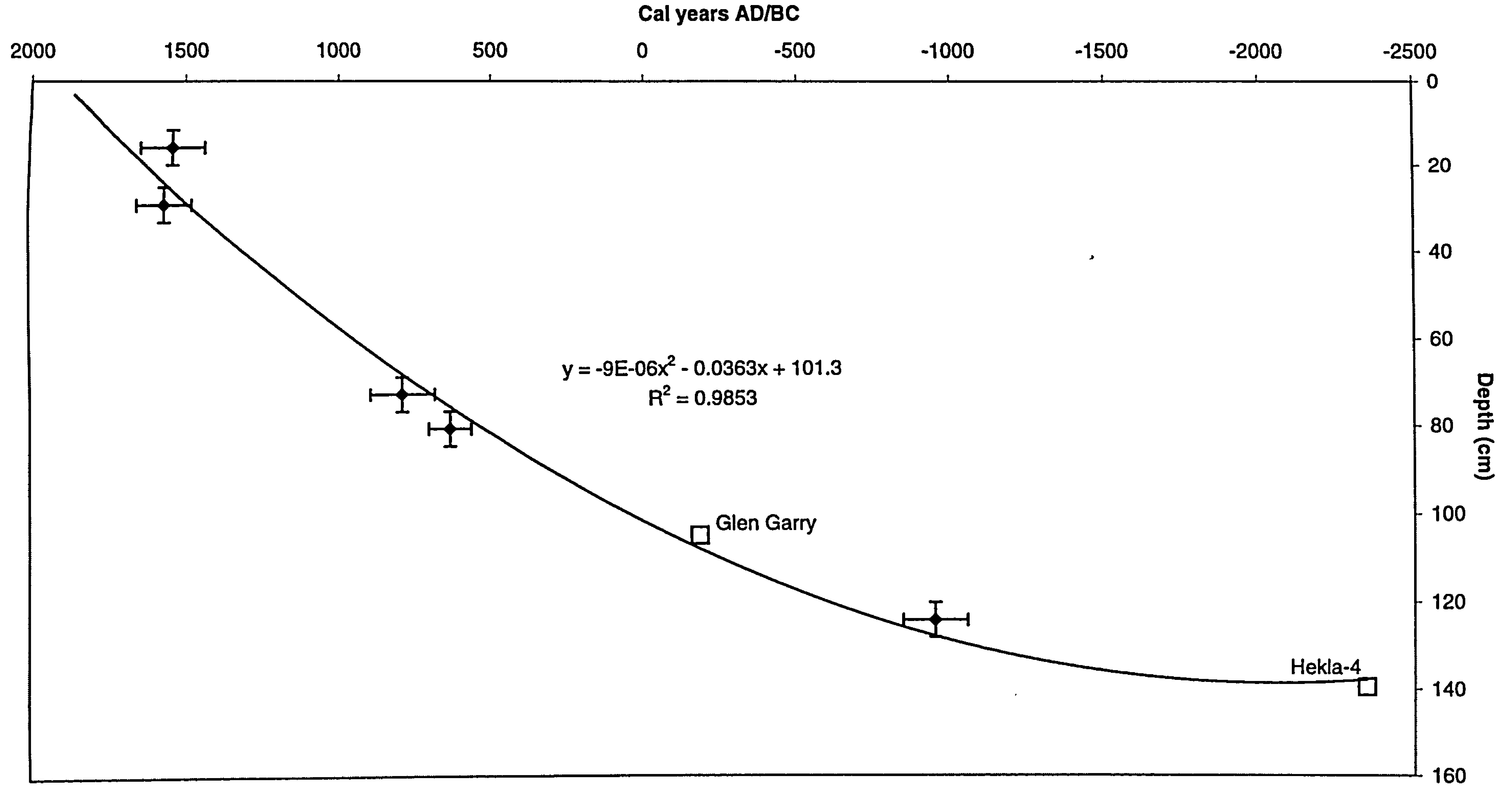


Figure 5.21 Ben Gorm Moss age/depth model.

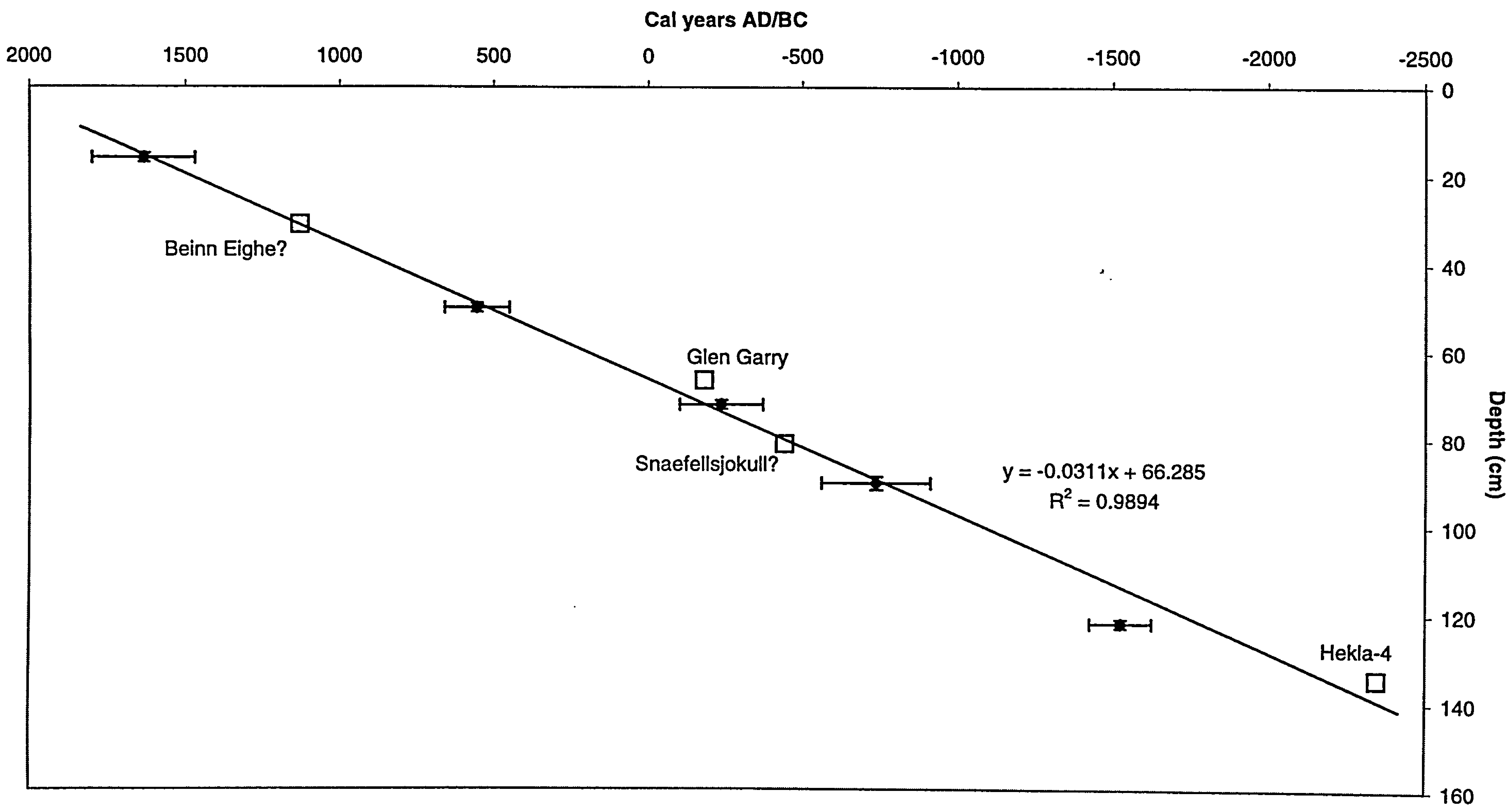
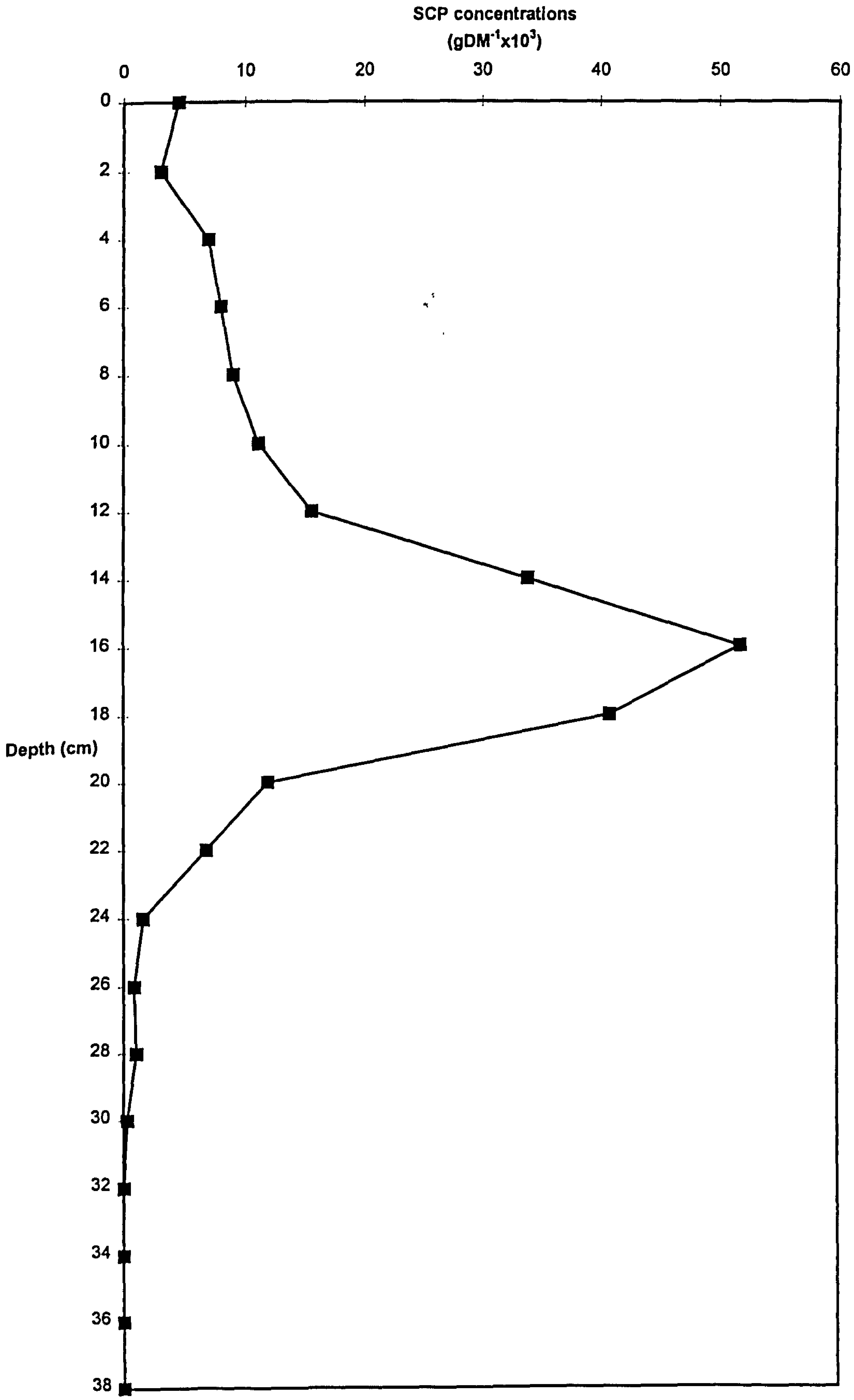


Figure 5.22 Shirgarton Moss - SCP stratigraphy



Chapter 6. Results

The results section forms three main parts. The first part (sections 6.1 to 6.5) describes the ecologies of all the species identified from the macrofossil and testate amoebae analyses. Details are also provided with respect to the understanding and interpretation of humification analysis, detrended correspondence analysis, and the testate amoebae transfer function. The second part of this chapter (sections 6.6 to 6.12) documents the stratigraphy, plant macrofossil analyses, humification analyses and testate amoebae analyses for each site. The multivariate statistical techniques and generation of proxy effective precipitation curves are described, and a synthesis of the data, incorporating the relationship to reconstructed mire water levels, is undertaken. The third part of the results (sections 6.13 and 6.14) analyses the datasets for periodicities and provides an overall summary of the data, which is discussed in relation to regional climatic changes in Chapter 7.

The species assemblages are displayed in macrofossil and testate amoebae TILIA diagrams for each site, and explained in detail in relation to inferred climatic changes. Subdivisions of the TILIA diagrams into zones was determined by visual inspection of each profile, as it was always obvious where the main changes in species assemblages occurred. Numerical zonation techniques were therefore not considered necessary, and it has been noted that subtle changes are often undetected using these methods (Birks and Birks, 1980).

6.1 Components of the macrofossil diagrams

Ecological studies on ombrotrophic mires act as a precursor to palaeoecological studies, as it is imperative to understand the ecological environments that species are able to tolerate before palaeoecological inferences can be made. A large proportion of observations used in this thesis originate from early research on peatlands, when the water table depths of individual species was noted as well as the ecology of individual species and assemblages. Key texts include Tansley (1939); Godwin and Conway's

(1939) work on Tregaron Bog, Wales; and the research undertaken in the Silver Flowe peat bogs, southwest Scotland - specifically by Ratcliffe and Walker (1958) and Boatman (1983). Other notable work has been undertaken on Clara Bog, Ireland (van der Molen and Wijmstra, 1994; van der Molen *et al.*, 1994), on Åkhult mire, Sweden (Malmer, 1962), and in North America (Andrus *et al.*, 1983; Janssens *et al.*, 1992). It must, however, be noted that in most of the studies the water level was only measured once (Rydin, 1993).

Each component is described in detail in relation to past water table depths, and hence inferences to climatic changes. At the end of this section a summary has been constructed detailing mire water table depth data of individual species (Table 6.1), as a guide to the ranges of moisture tolerances of species assemblages.

6.1.1 UOM

The unidentifiable organic matter (UOM) component consists of material that could not be identified to taxon level. High levels of UOM can be generally related to well humified peat with high proportions of decomposition usually related to a drier climate. However, care must be taken when interpreting the significance of UOM as it can also be indicative of algal muds. Oxygenation by pool algae, as well as a luxuriant *Sphagnum* carpet, may enhance the decomposition of pool vegetation, converting the dead algae into algal muds (Johnson *et al.*, 1990). Hence the UOM component can have a bimodal distribution along a wet/dry gradient.

It is difficult to differentiate between algal muds and well humified UOM using macrofossil analysis in the laboratory, as the peat oxidises rapidly after extraction in the field to a dark black/brown colour. However, these algal muds can be clearly seen in the field stratigraphy (e.g. Ratcliffe and Walker, 1958), as demonstrated later in this Chapter, and hence should be recorded in the field in order to avoid confusion when interpreting the macrofossil records (*cf.* Hughes *et al.*, submitted).

6.1.2 Ericaceae

The Ericaceae curve generally relates to the fine rootlets found in the peat matrix, most likely composed of *Calluna vulgaris* or *Erica tetralix*, or possibly *Vaccinium oxycoccus*, *Vaccinium myrtillus*, or *Empetrum nigrum*. The more woody material can often be identified to species level if the petioles are preserved on the sample, and these are represented as either wood undiff. (if no petioles can be found), *Calluna* wood or *Erica* wood. Extra information can also be gained by identifying the *Calluna* and *Erica* leaves, seeds and flowers which are represented on a 1-5 scale in the macrofossil diagrams.

Calluna is able to survive in a continuously growing *Sphagnum* mat as it has the ability to rejuvenate below ground by forming adventitious roots, making it potentially immortal in that environment (Barber, 1981; Wallen, 1987). This adaptation also allows *Calluna* to escape sites which have become unfavourable and re-establish itself, a prerequisite for mire survival (Wallen, 1987; Gimingham, 1988; MacDonald *et al.*, 1995). Casparie (1972) notes that *Calluna vulgaris* is very rarely absent in ombrogenous peat, and therefore can be regarded as one of the common mire plants. *Erica tetralix* has also been shown to rejuvenate quickly after a fire, occupying over half a mire surface with juvenile plants only months after the fire had occurred (Barkman, 1992).

On mires, *Calluna vulgaris* has a distinct distribution with respect to hydrology, generally dominating the hummock tops which allows the stem to act as scaffolds to the weak stemmed Sphagna, preventing the structure from collapse (Wallen, 1987). The lower limit of *Calluna* dominance ('*Calluna* limit') has been described as the boundary between hummocks and lawns signifying the lack of tolerance for waterlogged conditions (Malmer, 1962). A study from Åkhult mire in Sweden suggests that *Calluna* roots from hummocks are found to extend to the water table, whereas *Calluna* growing on lawn communities exhibits roots extending to 15cm below the water table, i.e. down to the minimum water table (Wallen, 1987).

Most hummocks in the Silver Flowe are dominated by *Calluna vulgaris* (Boatman, 1983), although *Calluna* and *Erica tetralix* can appear in lawn situations, but the plants are usually sparse and stunted (Ratcliffe and Walker, 1958). Malmer (1962) suggests that *Erica tetralix* can tolerate slightly wetter conditions than *Calluna*, and Økland (1990) notes that whilst *Calluna* is dominant on hummock tops, *Erica tetralix* has an optimum position on the lower hummocks and upper lawn microforms. *Erica tetralix* has been shown to occur with *Sphagnum papillosum* and *Rhynchospora alba* communities, suggesting it can withstand slightly wetter growing conditions (van der Molen and Hoekstra, 1988). Hammond *et al.* (1990) observe that 90% of occurrences for *Calluna vulgaris* are at 30 cm above the water table, ranging from 10-52cm, whereas *Erica tetralix* has a mean occurrence at 35cm above the water table, with ranges of 5-52cm. The identification of peaks in Ericaceae, and more specifically *Calluna vulgaris*, from the subfossil material can therefore be associated with relatively low water table depths, and a drier/warmer climate.

6.1.3 Monocots undiff.

The identification of epidermal material from the sedges *Eriophorum vaginatum*, *Eriophorum angustifoilum*, *Rhynchospora alba*, and *Scirpus cespitosus* enable them to be differentiated from each other and other possible monocotyledons. However, epidermal tissue was not always present in the samples, in which case the existing stems, rhizomes and roots were counted as monocots undiff. Where samples contained remains of one specific sedge, as well as undifferentiated material, it is likely that the material is also from the same plant, but cannot necessarily be proved due to the lack of diagnostic epidermal material.

6.1.4 *Eriophorum vaginatum*

Eriophorum vaginatum can grow over a wide range of hydrological conditions (Wein, 1973) and has the broadest habitat niche when compared with all other species found on the mire surface (Økland, 1990). Although mainly found on hummocks, *E. vaginatum* has been found to grow in fairly dense tussocks on drier lawns close to the water table (Ratcliffe and Walker, 1958). *E. vaginatum* is able to tolerate dry mire conditions, where the water table is very low, as the rhizomes can extend to depths of 60cm (Boggie *et al.*, 1958; Gore and Urquhart, 1966). This was emphasised on Tregaron Bog, parts of which were termed '*Eriophorum* islands' by Godwin and Conway (1939), often being associated with *Scirpus cespitosus*. These islands varied from 0.5 to 3m in height, often being associated with a moss cushion of *S. capillifolium* var. *rubellum*. Hammond *et al.* (1990) note how although 90% of occurrences for *E. vaginatum* are at 24cm above the water table, the range over which it is found is between 0-28cm, indicating the wide range of hydrological conditions it can tolerate.

Although *Eriophorum vaginatum* is often found colonising hummocks, its occurrence may not be linked with desiccated peat without further qualification (Casparie, 1972). If it is found with high levels of *Calluna*, then dry conditions may be inferred, but, it is also known that *Eriophorum vaginatum* is adaptable to conditions with a strongly fluctuating water table (Casparie, 1972). *Eriophorum vaginatum* is often dominant in peat communities where the water table can be high enough to flood the hollows between tussocks in winter, whereas the upper horizons may dry out considerably in summer (Tansley, 1939; Gimingham, 1964), indicating a preference for fluctuating water tables. It's ability to colonise shallow pools has also been documented (Osvald, 1923; Barber, 1981). Drought is not really a problem for the species due to their deep rooting habit and the high moisture holding ability of the organic substrate, and often the species becomes more luxuriant and tussocky as a bog dries out (Wein, 1973). The tussock habitat of *Eriophorum vaginatum* has the ability to conserve nutrients (Jonasson and Chapin, 1991) and possesses a favourable thermal regime as it is an effective receiver of insolation (Chapin *et al.*, 1979), especially suited to maximising CO_2 uptake at the beginning of the growing season (Robertson and Woolhouse, 1984). These attributes are thought to be responsible for its high productivity and widespread distribution, making it a difficult species for *Sphagnum* to replace.

Eriophorum vaginatum is able to tolerate fire well, as the shoots are protected by the tussock growth form, and it is able to recover quicker than the more woody Ericaceae (Ratcliffe, 1959; Barber, 1981). This can be identified in the macrofossil remains when *Eriophorum vaginatum* is associated with, or found just after macroscopic charcoal. The presence of *Eriophorum vaginatum* in the macrofossil record is thus usually indicative of a drier climatic phase, although the species assemblages associated with *E. vaginatum* will aid further interpretations.

6.1.5 *Eriophorum angustifolium*

There is direct space competition between *Eriophorum angustifolium* and *Eriophorum vaginatum/Scirpus cespitosus*, with rhizomes of *Eriophorum angustifolium* usually unable to penetrate tussocks of *Eriophorum vaginatum/Scirpus cespitosus*. The shoots of *Eriophorum angustifolium* are therefore confined to the relatively wet hollows, often found growing in association with *Rhynchospora alba* (Phillips, 1954).

Within the Silver Flowe, the *Sphagnum* mat in the middle of a dwindling pool is invaded at an early stage by *E. angustifolium*, although the sedge appears more sparse in open, deep pools (Ratcliffe and Walker, 1958; Boatman and Tomlinson, 1977). However, *E. angustifolium* often grows as well on hummock tops as it does at pool levels (Ratcliffe and Walker, 1958). It is distributed on Mongan Bog, Ireland, throughout hummock and hollow habitats, although shows a preference for the wetter hollows (Madden and Doyle, 1990).

Godwin and Conway (1939) found marked abundances of *Eriophorum angustifolium* at Tregaron Bog, and suggest that it is characteristically a colonist of drying pools. It is noted as an important coloniser of bare peat, or mud bottoms of shallow pools, although it is eventually replaced by *Eriophorum vaginatum*, allowing *E. angustifolium* to spread into the surrounding area (Phillips, 1954). *Eriophorum angustifolium* can also be found growing in patches over *Sphagnum* floating mats, as observed in a bog in southern Estonia (Frenzel and Rudolph, 1998).

A well protected stem apex allows *Eriophorum angustifolium* to survive burning episodes, although the species will gradually disappear due to secondary effects, such as the lowering of the water table, or competition with *Sphagnum* species which it cannot endure (Phillips, 1954). The identification of subfossil epidermal material of this sedge thus infers a relatively high mire water table.

6.1.6 *Rhynchospora alba*

In the Silver Flowe peat bogs *Rhynchospora alba* is located when the *Sphagnum* mat reaches or rises above the water level (Ratcliffe and Walker, 1958). It grows in a narrow water level range, marking the boundary between pool and lawn, disappearing at an early stage of hummock development. Godwin and Conway (1939) also document a sprinkling a *R. alba* within shallow pools of 10-15cm in depth, and Madden and Doyle (1990) note how on Mongan Bog *Rhynchospora alba* is almost confined to hollows, a feature which is also documented by Tansley (1939). It is rarely found on Åkhult mire at levels above 15-20cm above the water level (Malmer, 1962), having a median depth to the water table of 5-7cm (Økland, 1990), or 8cm as documented by Hammond *et al.* (1990). Backéus (1985) notes that specimens of *R. alba* are much larger at the edge of pools than on other microforms on ombrotrophic mires, suggesting that this is their optimal habitat. Identification of subfossil material is thus interpreted as having grown in a relatively wet environment.

6.1.7 *Scirpus cespitosus*

Scirpus cespitosus was recorded in bogs from as early as the mid-sixteenth century, when Linnaeus (1751) characterised it as one of the main hummock species on Swedish bogs, being associated with *Calluna vulgaris* and reindeer lichen (*Cladonia rangiferina*). *Scirpus cespitosus* is found in southwest Sweden from the water level upwards (Malmer, 1962), and optimum abundance is in the upper lawn/lower hummock microforms, being virtually absent from pools (Økland, 1990), although stunted growth forms can appear on lawn microforms (Ratcliffe and Walker, 1958). *Scirpus cespitosus* abundance has been documented from Tregaron Bog, Wales (Godwin and Conway, 1939) and forms low tough tussocks up to 60cm in diameter, but not usually over 15cm in height, whereas 90% of species occurred 25cm above the water table on a raised bog in County Kildare (Hammond *et al.*, 1990). Tansley (1939) suggests that *Scirpus* peat is somewhat drier than *Eriophorum* peat, indicating it has a preference for hummock environments on raised bogs, and it also appears to be a main component of blanket bogs, often occurring with *Calluna*. Rodwell *et al.* (1991) describe it as a lawn species tolerant of waterlogging, whilst Eddy *et al.* (1969) found it growing in wetter areas of the Moor House blanket mire.

It would therefore appear that on raised mires, *Scirpus cespitosus* is indicative of drier conditions, but other associated vegetative remains within the assemblage are required to qualify this. In blanket mires, it would appear that *Scirpus cespitosus* is able to tolerate relatively wetter conditions, possibly due to greater water flow.

6.1.8 *Odontoschisma sphagni*

This was the only liverwort found in the macrofossil remains, as was also observed by Mauquoy (1997), and only occurred in the surface sample of Langlands Moss. Ratcliffe and Walker (1958) note the appearance of *O. sphagni* in sparse populations on the edges of pool-hummock complexes, and more commonly in lawns and medium to tall hummocks. This species is noted as occurring in a dry habitat in Clara Bog, associated with *Eriophorum vaginatum* (van der Molen *et al.*, 1994). Its sparse occurrence in the macrofossil data precludes the record of this species from having any influence on the results.

6.1.9 *Aulacomnium palustre*

Restricted to hummocks on mires in southeast Norway, and found in abundance with *Polytrichum strictum*, this moss appears to be optimally located in more minerotrophic sites (Økland, 1990). It has also been noted to grow in small patches on hummock sides (Andrus *et al.*, 1983), although it has been noted to tolerate a wide range of conditions on mires (Dickson, 1973).

6.1.10 *Dicranum scoparium*

Smith (1980) describes the distribution of *Dicranum scoparium* as being frequent in the wetter parts of Britain. It has also been described as having a wide ecological range, but on bogs tends to occur under tall *Calluna*, and often in association with *Hypnum jutlandicum* (Corley, 1992), indicating a preference for hummock, and comparatively dry environments.

6.1.11 *Drepanocladus fluitans*

Primarily a species of base-poor pools (Dickson, 1973), *Drepanocladus* leaves were found within a *Sphagnum cuspidatum* layer at peat-face I from Emmen 25 (Casparie,1972), inferring wet conditions at the time of formation. Van der Molen *et al.*

(1994) locate *Drepanocladus fluitans* in the community of a wet habitat on Clara Bog, along with *Rhynchospora alba* and *Sphagnum cuspidatum*, suggesting it is indicative of wet conditions. It has also been identifed in relative abundance in association with aquatic species in peats from New Zealand (McGlone and Wilmshurst, 1999). However, only very low amounts of this species were found at the sites studied.

6.1.12 *Hypnum cupressiforme*

Hypnum cupressiforme has been found on hummocks in association with *Sphagnum capillifolium* var. *rubellum* on blanket mires from the Silver Flowe (Boatman, 1983). *Hypnum cupressiforme/jutlandicum* have also been correlated with strong *Calluna* layering, indicative of the shading and dryness promoted by increased canopy growth (MacDonald *et al.*, 1995). Other examples document *Hypnum cupressiforme* found on slopes between hummocks and hollows on Clara Bog, Co. Offaly (van der Molen and Wijmsta, 1994), and it has been associated with *Calluna* hummocks at Tregaron Bog, Wales (Godwin and Conway, 1939) and scattered on hummocks on a small bog near Loch Maree, Scotland (Tansley, 1939). It is thus considered to grow in dry habitats.

6.1.13 *Pleurozium schreberi*

This moss has been recorded on *Sphagnum capillifolium* var. *rubellum* hummocks in the Silver Flowe (Boatman, 1983). It has also been found to be restricted to hummocks with infrequent distribution in southeast Norway, growing only where the depth to the water table is greater that 12cm, and often found in pure patches under *Calluna* stands (Økland, 1990). *P. schreberi* was noted by Andrus *et al.* (1983) to mainly grow in small patches on hummock sides, reflecting the dry habitat on which it grows (Smith, 1980).

6.1.14 *Pohlia nutans*

Økland (1990) found *Pohlia nutans* to be restricted to hummocks on a southeastern mire in Norway, with infrequent distributions mainly occurring in *Sphagnum* species as opposed to other mosses. Smith (1980) suggests that the species can grow in wet or dry habitats. However very little subfossil material was recovered from any of the sites.

6.1.15 *Polytrichum juniperum/alpestre* type

These species are difficult to distinguish between in the subfossil state, especially if the leaf tips have been eroded away (Dickson, 1973); indeed most non-British authorities treat *Polytrichum alpestre* as a variety or subspecies of *Polytrichum juniperum* (Smith, 1980). *Polytrichum alpestre* has been recorded on *S. capillifolium* var. *rubellum* hummocks in the Silver Flowe (Boatman, 1983) and is best known as a former of dense tussocks on ombrogenous bogs (Dickson, 1973). *Polytrichum juniperum* tends to grow on dry acid sites, usually well drained (Dickson, 1973; Smith, 1980), and its occurrence is thus interpreted as indicating a relatively low water table depth.

6.1.16 *Racomitrium lanuginosum*

Racomitrium lanuginosum is an extremely oceanic species, occurring as a community dominant mainly in highly oceanic areas (Lindsay *et al.*, 1988). This moss has been identified on the tallest hummocks within the Silver Flowe, often in a drying out phase (Ratcliffe and Walker, 1958). It is suggested that these drying out hummocks were originally composed of *S. rubellum*, *S. fuscum* or *S. imbricatum* (Ratcliffe and Walker, 1958), although Gimingham *et al.* (1961) suggest that hummocks which are dominated by *Racomitrium lanuginosum* appear to be virtually devoid of *Sphagnum* species. Ratcliffe (1964) regarded *Racomitrium lanuginosum* hummocks to be an indication of burning, and considers a mire with an abundance of these hummocks to be degenerate. Moore (1977), however, claims that *Racomitrium lanuginosum* has been present in the stratigraphy of Claish Moss over the last 5000 years, and is therefore a natural component of the mire. Tallis (1995) suggests that phases of growth of *Racomitrium lanuginosum* from blanket mires in the Southern Pennines are coincident with the Early Medieval Warm Period, having been dated to between 1150-1300 AD. *Racomitrium lanuginosum* has also been identified in the stratigraphy of a montane blanket bog at Moine Mhor in the Cairngorms, and was found to alternate with *Sphagnum*-rich peat, representing relatively dry and wet periods respectively (Barber *et al.*, 1999).

Godwin and Conway (1939) associate *R. lanuginosum* with *Calluna* hummocks at Tregaron Bog, Wales, and it is also found scattered on hummocks in Scotland (Tansley, 1939), indicating the dry-growing stages of ombrotrophic bogs it represents in the subfossil record (Dickson, 1973).

6.1.17 *Rhytidiadelphus loreus*

This moss exhibits a markedly western range in the British Isles, often growing above the tree-line, and subfossils have been reported mainly from the Atlantic or later periods (Dickson, 1973). The only subfossil remains found in this study were from Ben Gorm Moss, the most westerly site within the study.

6.1.18 Identifiable *Sphagnum*

Sphagnum is the most successful bryophyte globally, and considering the amount of carbon incorporated within it, it is possibly the most successful plant genus of any kind (Clymo, 1997). *Sphagnum* species have been described as "water-loving plants with numerous adaptations for dealing with periodic drought conditions" (Andrus, 1986, p.416). Even though different species are involved in oceanic and continental climates, it is suggested that the mechanisms behind interspecific competition are similar, even between remote areas (Rydin, 1993). Therefore the factors that determine species composition at the start of a plant's growth, may be decisive for the community structure for a long time (Rydin, 1997). Depth to the water table has been determined to be the most important factor for *Sphagnum* growth (see section 3.4), which is controlled by climate in ombrogenous bogs, and thus the interpretation of changes in *Sphagnum* assemblages as a response to changes in climate becomes valid.

Water table depth is more important than chemical status in controlling the distribution of *Sphagnum* species. In years of average precipitation hollow species such as from section *Cuspidata* are more productive than the hummock species of section *Sphagnum*, whereas in dry years the reverse is true, with the hollow species more subject to summer desiccation (Gerdol, 1995). The sharp segregation along the water table depth gradient displayed by hollow Sphagna is determined by a higher degree of competition, whereas the hummock species appear to be segregated by physiological tolerance (Bragazza, 1997). Hummock species have a better ability to supply the photosynthetically active parts with water under dynamic conditions (Rydin, 1985; Hayward and Clymo, 1982), which suggests that the higher evaporation rate found in hummocks (Clymo, 1973) is well compensated by the superior water transport capacity of hummock species (Rydin, 1985). There is also a difference in densities between species within hummocks, with *ca.* 360-630 capitula dm^{-2} for section *Acutifolia*, whereas section *Sphagnum* have only 140-300 capitula dm^{-2}, with even smaller ranges being found in the lawn and pool species (van der Molen *et al.*, 1994).

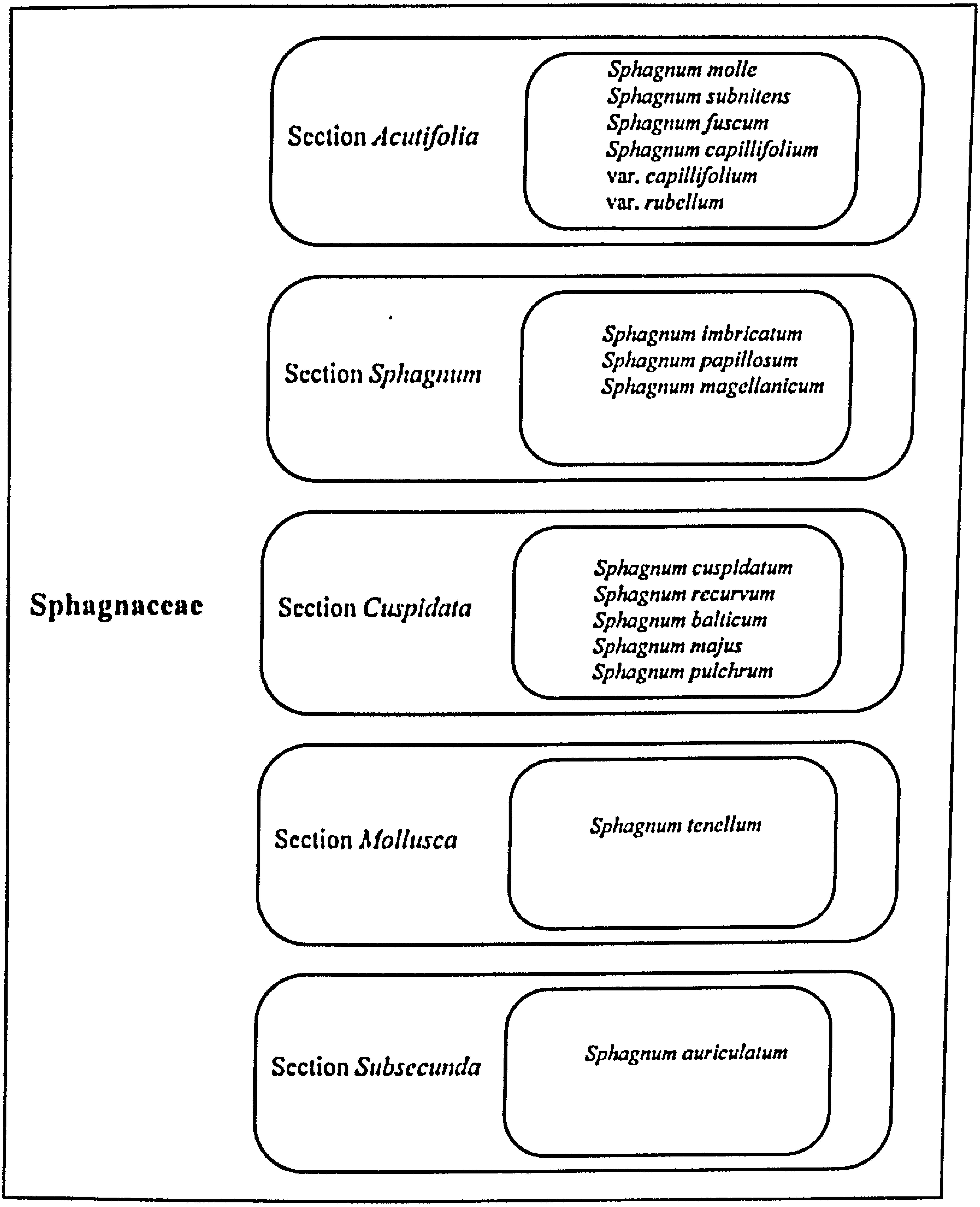
The differences in subfossil Sphagna assemblages can thus be interpreted climatically as well as ecologically. Daniels and Eddy (1990) define 14 different species which can be readily found on British mires (Figure 6.1). The lowest taxonomic levels to which many subfossil leaves can be identified is to section level (e.g. Smith 1985; Wimble, 1986) although section *Sphagnum* is easily identified to species level (e.g. Mauquoy and Barber, 1999b).

6.1.19 *Sphagnum* section *Acutifolia*

Sphagnum section *Acutifolia* contains hummock and lawn species, and thus has a bimodal distribution, or a wide ecological niche centering on a low hummock position along a moisture gradient. Identification was based on the leaves of the section species being generally small to medium size (0.8-1.9mm), ovate to lanceolate, and possessing a narrow border of 1-2 cells (Daniels and Eddy, 1990).

Five main species dominate section *Acutifolia* on ombrotrophic mires in Britain (Figure 6.1) and include the hummock species *S. capillifolium* var. *capillifolium*, *S. capillifolium* var. *rubellum*, and *S. fuscum*, and the lawn species *S. subnitens* and *S. molle*. *Sphagnum molle* is possible to identify due to its larger size (1.4-2.3mm) and the presence of a resorption furrow on the outer wall of the outer marginal cells (Daniels and Eddy, 1990). However, no subfossil leaves of this species were found, possibly due to the resorption furrow having been eroded during peat decomposition. The other species are difficult to differentiate in the subfossil form (Stoneman, 1993; Mauquoy, 1997), although recently a key has been developed by McMullen (in prep). Species have been identified to section level in this study.

Figure 6.1 The different *Sphagnum* species found on British ombrotrophic mires (after Daniels and Eddy, 1990), cited in Mauquoy, 1997.



Subfossil leaves of *Sphagnum tenellum* (section *Mollusca*) may also be misidentified as section *Acutifolia*, as they are small in size (0.8-1.5mm). *Sphagnum tenellum* generally grows with 'unaggressive' species, often associated with *S. subnitens*, *S. papillosum*, or *S. capillifolium* on raised bogs (Daniels and Eddy, 1990), and hence the ecology would not differ greatly from that of section *Acutifolia*. An abundance of *S. tenellum* has been documented growing amongst *Scirpus cespitosus*, as well as lining the pools between tussocks at Tregaron Bog, Wales (Godwin and Conway, 1939). Rydin (1985) suggests that as *S. tenellum* is a species which dries out quickly, it would be unable to survive at very high levels above the water table, even if supported by surrounding hummock mosses.

The most common species from this section to dominate the raised mires in Scotland is *Sphagnum capillifolium* var. *rubellum* (personal observation), which can form dense mound like cushions 60cm above the water table, although lower hummocks of about half the height are more common (Ratcliffe and Walker, 1958; Rydin and McDonald, 1985; Rydin, 1993). *S. capillifolium* var. *rubellum* also has broad niche overlaps with all *Sphagnum* species studied by Bragazza (1997), allowing it to colonise most microforms. Barber (1981) suggested that the section *Acutifolia* found in the stratigraphy of Bolton Fell Moss was predominantly *Sphagnum capillifolium* var. *rubellum*, and Store Mosse in southern Sweden has been shown to be stratigraphically dominated by *Sphagnum fuscum* and *Sphagnum rubellum* (Svensson, 1988a; 1988b).

Palaeoecological evidence suggests that within peat sequences section *Acutifolia* is generally dominated by hummock forming species, although this should be qualified with other macrofossil/microfossil evidence, stressing the need for multiproxy data. Colorimetric humification and testate amoebae analyses have been used by Mauquoy (1997), Chiverrell (1998), and Chiverrell and Atherden (1999) to aid interpretation of the climatic conditions under which section *Acutifolia* species grew, and these analyses have also been employed in this study.

6.1.20 *Sphagnum imbricatum* spp. *austinii*

Sphagnum imbricatum has occasioned more comment from British ecologists than any other bryophyte, predominantly due to its abundance of remains from mid-late Holocene ombrotrophic peats compared with its present scarcity or absence from mires (Dickson, 1973; Mauquoy and Barber, 1999b). This species is easily identified due to the presence of comb fibrils on the internal commissural leaf walls (Daniels and Eddy, 1990).

Two distinct European subspecies of *Sphagnum imbricatum* have been identified, namely *Sphagnum imbricatum* spp. *affine*, a minerotrophic subspecies, and *Sphagnum imbricatum* spp. *austinii*, an ombrotrophic subspecies (Flatberg, 1986). This division is supported by Hill (1988), although two separate ecads of *Sphagnum imbricatum* spp. *austinii* exist in British mires, allowing it to grow in both hummocks and lawns. A semi-aquatic ecad, occurring in *Sphagnum* lawn communities, has a lax growth form (Tallis, 1961), which has been experimentally induced by Green (1968). A relatively slow-growing, desiccation-resistant, compact hummock form also exists (Green, 1968). *Sphagnum imbricatum* is a very slow growing species (although the lax form is slightly faster than the compact form), at its best unable even to match the rate of *S. papillosum* (Green, 1968), suggesting that it is ousted from lawns by the latter species, unless competition from other *Sphagnum* species is low (Flatberg, 1986). The separate ecads of the species give to a bimodal distribution over moisture gradients (similar to the section *Acutifolia*), creating problems with palaeoenvironmental interpretations, and hence multiproxy data is necessary to overcome such difficulties.

Sphagnum imbricatum is generally regarded as an oceanic species (Flatberg, 1986; Daniels and Eddy, 1990) although Birks *et al.* (1976) consider it as having Northern-Continental rather than Atlantic affinities. This has also been demonstrated by Overbeck (1975, p. 589), who produced a detailed map of *Sphagnum* peat from the layers above the *Grenzhorizont* from over 50 bogs around northern Germany. The map illustrated a 'hinge line' between more oceanic areas to the north and northwest, where *Sphagnum imbricatum* dominated the stratigraphy, and more continental areas further south and southeast where *Sphagnum* section *Acutifolia* dominated, and *S. imbricatum* was absent from the stratigraphy.

Sphagnum imbricatum is found presently at few sites in Britain (see Birks *et al.*, 1976), but has been documented in the Silver Flowe in both growth forms (Ratcliffe and Walker, 1958) and also on Clara Bog, Co. Offaly (van der Molen and Wijmstra, 1994) in the compact hummock form. *Sphagnum imbricatum* is also restricted to hummocks at Borth Bog in Wales (Slater and Slater, 1978), at Claish Moss in southwest Scotland (Moore, 1977), and at Coom Rigg Moss in Cumbria, where the hummocks reach 50cm in height (Chapman, 1964). In Scandinavia, *Sphagnum imbricatum* has been documented at Åkhult mire in southwest Sweden where it is not restricted only to hummocks (Malmer, 1962).

Stratigraphically *Sphagnum imbricatum* was found to dominate on the hummock peats at Emmen in the Netherlands, although it was also found with *Sphagnum papillosum* in the hollow peats (Casparie, 1972). Further palaeoecological evidence for the lax form of *Sphagnum imbricatum* can be found at Bolton Fell Moss, where *Sphagnum imbricatum* leaves were found with frequent remains of *Sphagnum cuspidatum* in a minor pool layer (Barber, 1981). The species has also been recorded from hummock microforms in peat stratigraphy. Plant macrofossil and testate amoebae analysis from Engbertsdijksveen suggested that *Sphagnum imbricatum* was found to mainly grow over hummocks from *ca.* 1000 BC to *ca.* AD 960, as xerophilous testate amoebae species were found associated with leaves of *Sphagnum imbricatum* (van der Molen and Hoekstra, 1988).

Sphagnum imbricatum has been suggested to be sensitive to burning and changes in the water table, particularly in the compact hummock form (Birks *et al.*, 1976; Boatman, 1983). Slater and Slater (1978) note that *S. imbricatum* had often been burnt on Borth Bog in Wales, although a few shoots on hummocks were able to survive. They suggest that *S. imbricatum* shows some ability to recover after burning events when left undisturbed for at least a decade, but repeated burning will ultimately lead to the plants extinction.

The fact that *Sphagnum imbricatum* no longer grows on most raised bogs in Britain argues for some sharp discontinuity between past and present conditions and former history of the bog (Godwin and Conway,1939), and raises important questions as to the nature of the extinction at specific sites. The extinction of *S. imbricatum* occurs on most British mires within the last millennia and has been discussed extensively in the literature. The most common hypotheses suggest either climatic change (e.g. Godwin, 1956; Barber, 1981; Stoneman *et al.*, 1993; Mauquoy and Barber, 1999b), direct anthropogenic disturbances through burning and drainage (e.g. Pearsall, 1956; Pigott and Pigott, 1963) or indirect anthropogenic disturbance through airbourne pollution (e.g. van Geel and Middeldorp, 1988) as an extinction mechanism. This is discussed fully in the light of the results from this study in Chapter 7.

6.1.21 *Sphagnum magellanicum*

Sphagnum magellanicum can be differentiated from other species by focusing on the branch leaf photosynthetic cells under a microscope. These cells are oval, and generally completely enclosed by hyaline cells (Daniels and Eddy, 1990), and thus focusing down through the photosynthetic cell will reveal a 'thin-thick-thin' pattern, diagnostic of *S. magellanicum.*

Ratcliffe and Walker (1958) showed *S. magellanicum* to have a narrow range of water level tolerance, extending only a short distance above the water table. Økland (1990), however, found that *S. magellanicum* had the broadest habitat niche out of any *Sphagnum* species when analysed using various DCA techniques, suggesting that it demonstrates an excellent competitive ability, and is therefore not an opportunist as claimed by Andrus *et al.* (1983). Økland (1990) finds *S. magellanicum* to be 'almost ubiquitous', mainly preferring the upper lawn microform, but also being present on lower lawns, and in hummocks. This bimodal distribution has also been indicated by Clymo and Hayward (1982).

S. magellanicum appears to benefit *S. papillosum* in hummocks through lateral water transport (as *S. magellanicum* is a superior competitor for water under dry conditions), suggesting protocooperation when the species grow together (Li *et al.*, 1992). This cooperative characteristic may have aided *S. magellanicum* in colonising new areas, as it has been found to have broad niche overlaps with all *Sphagnum* species studied by Bragazza (1997), and therefore could explain its present day dominance, as seen from macrofossil diagrams (e.g. Stoneman, 1993; Mauquoy and Barber, 1999a; 1999b). Present day aerosol derived nutrients may also be responsible for the increase in abundance of *S. magellanicum* (van Geel and Middeldorp, 1988). Some evidence is available to corroborate this, as low levels of nitrogen addition to *S. magellanicum* show an increased growth response, although an increase in addition showed no significant effect. It was suggested from N:P ratios that phosphorus was lacking, thus probably limiting the growth response of *S. magellanicum* (Willams and Silcock, 1997).

Even though *S. magellanicum* is only a recent species addition to mires in Britain, it has existed in mainland Europe for much longer. Svensson (1988a) and Malmer *et al.* (1997) describe a 'Magellanicum bog stage' covering the last *ca.* 1000 years from Store Mosse, South Sweden, where the hollow peat is dominated by comparatively unhumified *Sphagnum magellanicum*. Janssens *et al.* (1992) document *S. magellanicum* from at least 2500 cal. BP from the peatlands of Minnesota.

Although *S. magellanicum* would appear to have a large moisture range, changes in water table depths have been shown to disturb *Sphagnum magellanicum* communities, although it was reported that fires rarely burn in these communities (Jasieniuk and Johnson, 1982). It therefore appears that it tends to avoid the driest and wettest parts as a single species of hummock/hollow microtransects, thus being a useful palaeoecological indicator (Mauquoy, 1997).

6.1.22 *Sphagnum papillosum*

The internal commissural walls of the branch leaves of *Sphagnum papillosum* are easily identified under the microscope by the diagnostic papillae they possess (Daniels and Eddy, 1990). The species tends to grow in extensive carpets relatively near to the water table, but can also form low hummocks. Ratcliffe and Walker (1958) report that *S. papillosum* appears well established a few centimetres above the water level, and has a maximum height of between 15-20cm above the water table as low hummocks or uneven lawns. Økland (1990) suggests that *S. papillosum* occurs only rarely in hummocks, mainly being established on lawn microforms, which is also documented by Bragazza (1997) who notes that *S. papillosum* is found associated with all *Sphagnum* species except the top hummock former, *S. fuscum*. The greatest niche overlaps were found with *S. capillifolium* var. *rubellum*, which can colonise lawns or low hummocks (Bragazza, 1997).

Tansley (1939), however, suggests that *Sphagnum papillosum* (along with *Sphagnum rubellum*) are the most important hummock formers in the British Isles, although he also notes its occurrence in wet hollows, indicating the relatively large range of moisture tolerance of this species. The water holding ability of *S. papillosum* has shown it to be a drought tolerator in low hummock positions, although in higher hummock positions it is benefited through lateral water transport by the presence of *S. magellanicum*, which exhibits superior competition for water (Li *et al.*, 1992). It would therefore appear that *S. papillosum* only tends to grow in hummocks when mixed with *S. magellanicum*, or if the water table is high. The balance of evidence would therefore suggest that this moss tends to occur in environments relatively close to the water table.

6.1.23 *Sphagnum* section *Cuspidata*

Elongate narrow hyaline cells, with triangular to trapezoidal shaped photosythetic cells make species within section *Cuspidata* clearly recognisable. The leaves are generally ovate to lanceolate in shape, with the most distinctive being *Sphagnum cuspidatum*, having leaves of up to 3.5mm in length, being at least 4-5 times as wide as long.

Species from section *Cuspidata* often dominate the lower parts of the moisture gradient (Rydin, 1993), and in Britain include *Sphagnum cuspidatum*, *Sphagnum recurvum*, *Sphagnum balticum*, *Sphagnum majus*, and *Sphagnum pulchrum* (Daniels and Eddy, 1990). However, all of these species are now rare in Britain with the exception of *Sphagnum cuspidatum* and *Sphagnum recurvum*, which tends to prefer slightly more eutrophic conditions than most oligotrophic Sphagna (Daniels and Eddy, 1990). This does not mean to say that these other species from section *Cuspidata* were rare in the past, but with subfossil *Sphagnum cuspidatum* leaves being found in most other northwest European peats (e.g. Casparie, 1972; van Geel, 1978; Barber, 1981; Mauquoy, 1997) it is sensible to suggest they form the bulk of subfossil section *Cuspidata* leaves identified in this project.

Malmer (1962) notes how during dry periods on Åkhult mire, *S. cuspidatum* carpets totally dry up, whereas the Sphagna on the highest parts of the hummocks remain wet. This has also been observed on Walton Moss, where extensive carpets of *Sphagnum cuspidatum* in April 1995 had disappeared by September after a hot/dry summer (Barber, pers. comm.). Økland (1990) found that *Sphagnum cuspidatum* had the most narrow habitat niche of all Sphagna on the mire surface at Åkhult, southwest Sweden. However, it has also been suggested that ombrotrophic conditions are sub-optimal for *S. cuspidatum*. Boatman (1977) suggests that smothering of *S. cuspidatum* shoots in winter on pool bottoms, and a low rate of innovation production in summer, caused by wind heaping shoots at end of pools, limits *Sphagnum cuspidatum* to fully occupy the pools on the Silver Flowe bogs.

Although clearly a pool species, compact forms can be found growing a short distance above the water level (Ratcliffe and Walker, 1958). Species from section *Cuspidata* (*Sphagnum balticum*) have been found growing at quite high levels above the water table, as they are able to obtain water through lateral transport via the contact of spreading branches of different plants (Rydin, 1985). However, only a few individuals are able to survive in this way. Thus small proportions of *Sphagnum* section *Cuspidata* in the stratigraphy do not necessarily imply a wet phase, and should be treated cautiously. Large abundances of *Sphagnum* section *Cuspidata*, however, indicate a former pool environment.

6.1.24 *Cenococcum grandiforme*

The distinctive black spherical sclerotia of the fungus *Cenococcum grandiforme* are frequently found in upper soil layers, as well as in surface vegetation, and seem to mainly be associated with woody peats, although they can thrive under a range of moisture conditions (Ferdinandsen and Winge, 1925). In a study of fen/bog transitional peats, Hughes (1997) found *C. grandiforme* to be mainly associated with *Eriophorum vaginatum* and *Calluna vulgaris* (and often charcoal), which he interpreted as a dry ombrotrophic bog community. *Cenococcum grandiforme* can therefore be viewed as a dry indicator within the stratigraphy.

6.1.25 Charcoal

Macroscopic charcoal can be indicative of burning events. Boatman (1983) showed how documentary evidence from the end of the 18th century indicated that muir-burning was common place in parts of Scotland, often carried out with a lack of care and discrimination. However, long term vegetation development may not be significantly influenced by peat surface fires, with changes in moss cover only limited to a few decades after the fire event (Kuhry, 1994).

In general fire only kills the surface vegetation, and often selectively burns the Ericaceaeous shrubs, having less effect on the *Sphagnum* (Jasieniuk and Johnson, 1982). Postfire assemblages in western subartic bogs have been overgrown initially by *Polytrichum* and *Pohlia*, followed by *Sphagnum* within 20-40 years (Jasieniuk and Johnson, 1982). However, fire does effect changes in species abundance within the stand. Sometimes the ash remains from fires can block up the pores of the underlying substrate, increasing localised water retention, and hence potential colonisation by hydrophiles (Mallik *et al.*, 1984).

Regional climate has been cited as the most useful variable in defining fire-frequency landscape units, for a watershed in the southern Canadian Rockies, with warmer and drier periods in postglacial times being highly correlated with increased fire frequency (Johnson, and Larsen, 1991). A decrease in mire water table, and hence drier climatic conditions have been tentatively correlated with fire abundance by Pitkänen *et al.* (1999), although Tolonen (1985) suggests a strong human influence on peatland fires in southwest Finland due to slash-and-burn cultivation.

It has also been suggested that fires have been a major factor in reducing peat accumulation in some areas. Pitkänen *et al.* (1999) were able to demonstrate this, by assuming constant accumulation rates for peat growth, and correlating the mean interval of charcoal layers and total carbon and peat accumulation, suggesting a reduction in peat accumulation. The amount of peat lost in fires is suggested to have varied from surficial burnings to the burning of thick layers of peat, with more discrete charcoal layers indicative of superficial burning.

Macroscopic charcoal remains were recorded on a 1-5 scale. A few charred plant remains usually represent only 1-5% of macrofossil assemblages and most likely represent local peatland fires, and would score 1-2 on the 5 point scale. However, macroscopic charcoal remains representing over 10% of the macrofossil assemblage are indicative of more severe fires, and would score 3-5, and where charred Ericaceae roots are found could indicate the burning of the local peat surface (Kuhry, 1994).

Table 6.1 Mean water depths of the species identified in this study on ombrotrophic bogs, from selected references.

Species	Mean depth to water table in cm (and range)	Country/Region of study	Author
Calluna vulgaris	28	Ireland	van der Molen 1986
	30 (10-52)	Ireland	Hammond *et al.* 1990
Erica tetralix	25	Ireland	van der Molen 1986
	35 (5-52)	Ireland	Hammond *et al.* 1990
Eriophorum vaginatum	28	Ireland	van der Molen 1986
	24 (0-28)	Ireland	Hammond *et al.* 1990
Eriophorum angustifolium	8-10	Ireland	van der Molen 1986
	24 (0-28)	Ireland	Hammond *et al.* 1990
Rhynchospora alba	10-15	Wales	Godwin and Conway 1939
	<15-20	S Sweden	Malmer 1962
	12-14	Ireland	van der Molen 1986
	8	Ireland	Hammond *et al.* 1990
	5-7	SE Norway	Økland 1990
	14	The Netherlands	Barkman 1992
Scirpus cespitosus	25	Ireland	Hammond *et al.* 1990
Aulacomnium palustre	31.4 ± 7.2	NE America	Andrus *et al.* 1983
	30.0 ± 8.1	NE America	Andrus *et al.* 1983
	19.5 ± 8.5	Minnesota	Janssens *et al.* 1992
Drepanocladus fluitans	16.7 ± 9.3	NE America	Andrus *et al.* 1983
	3.7 ± 7.0	Minnesota	Janssens *et al.* 1992
Pleurozium schreberi	28.5 ± 9.2	NE America	Andrus *et al.* 1983
	29.6 ± 7.1	NE America	Andrus *et al.* 1983
	>12cm	SE Norway	Økland 1990
	26.9 ± 9.7	Minnesota	Janssens *et al.* 1992
Pohlia nutans	31.1 ± 13.7	NE America	Andrus *et al.* 1983
	34.9 ± 11.1	NE America	Andrus *et al.* 1983
Polytrichum juniperum/alpestre type	30.8 ± 9.5 (*P. commune*)	NE America	Andrus *et al.* 1983
Sphagnum capillifolium var. *rubellum*	24.8 ± 9.4	NE America	Andrus *et al.* 1983
	28.8 ± 9.4	NE America	Andrus *et al.* 1983
	10-20	E Sweden	Rydin and McDonald 1985
	25	Aggregate data	van der Molen and Hoekstra 1988
	18.4 ± 11.6	Italy	Bragazza 1997
Sphagnum capillifolium var. *capillifolium*	19.3 ± 7.4	Minnesota	Janssens *et al.* 1992
	24-38	Italy	Gerdol 1995
	33.7 ± 17.8	Italy	Bragazza 1997
Sphagnum fuscum	30.9 ± 10.6	NE America	Andrus *et al.* 1983
	15-30	E Sweden	Rydin and McDonald 1985
	27.5 ± 9.5	Minnesota	Janssens *et al. 1992*
Sphagnum imbricatum	35	Aggregate data	van der Molen and Hoekstra 1988
Sphagnum magellanicum	25.8 ± 9.5	NE America	Andrus *et al.* 1983
	37.5 ± 12.1	NE America	Andrus *et al.* 1983
	15	Aggregate data	van der Molen and Hoekstra 1988
	21	The Netherlands	Barkman 1992
	19.4 ± 8.3	Minnesota	Janssens *et al.* 1992
	19-34	Italy	Gerdol 1995
	27.4 ± 15.4	Italy	Bragazza 1997
	1.5-8.5	NE Scotland	Willams and Silcock 1997
Sphagnum papillosum	max 15-20	Scotland	Ratcliffe and Walker 1958
	4	Aggregate data	van der Molen and Hoekstra 1988
	12.0 ± 6.8	Minnesota	Janssens *et al. 1992*
	11.5 ± 5.4	Italy	Bragazza 1997
Sphagnum cuspidatum	12.1 ± 5.4	NE America	Andrus *et al.* 1983
	1	Aggregate data	van der Molen and Hoekstra 1988
	1.2	The Netherlands	Barkman 1992

6.2 Peat humification

The degree of peat humification is related to the decomposition of the subfossil peat matrix (see section 3.5 for a full review). Higher values of absorbance are related to more humified peats, as these contain a greater amount of decomposed humic acids. The decomposition state is governed by the time taken for vegetative remains to pass through the biologically active acrotelm to the relatively inert catotelm. This is controlled by the average water table depth within the peat, high water tables resulting from a high effective precipitation regime, leaving less time for the peat to decompose before it is passed through to the catotelm, where rates of decomposition are extremely slow (Ingram, 1982; Clymo, 1984).

The humification value obtained from colorimetic analysis is an average value of the time taken for the peat to pass through the acrotelm to the catotelm, and may therefore be out of phase with other mire surface wetness proxies (Stoneman, 1993). Peat can reside for long periods in the acrotelm, up to 100 years, and the range of variation for acrotelm decay losses has been shown to be between 25-80% (Malmer *et al.*, 1997). However, under regimes of increased effective precipitation decay losses in the acrotelm could be reduced to *ca.* 40-50%, shortening the residence time by >50% (Malmer *et al.*, 1997), thus reducing any lag in the humification signal for a wet phase.

Humification records from blanket mires have been shown to be independent of major vegetational changes, suggesting that they can be solely attributed to climatic changes (Chambers *et al.*, 1997a). Changes in humification from raised mires have also been shown to be broadly synchronous with changes in plant macrofossil records, suggesting they can be used in tandem for reconstructing mire water-levels (Barber *et al.*, 1994c; Mauquoy and Barber, 1999a).

6.3 Testate amoebae

The different taxa found in this study are documented in Table 6.3, along with qualitative and semi-quantitative descriptions. The data from Woodland (1996) is more quantitative, based on hydrological reconstructions of depth to the water table from British mires. Table 6.3 is not exhaustive, but documents a reasonable proportion (including the most recent) of the critical work undertaken in this field. Semi-quantitative moisture classes have been attributed to testates by Jung (1936) along a hydrophilous hummock-hollow gradient. The average water content of these groups has been characterised by Meisterfield, and is shown along with other data in Table 6.2.

Table 6.2 Moisture classes of testate amoebae (cited in Tolonen, 1986; Heathwaite *et al.*, 1993; Hendon, 1998).

Rhizopod group de Graff (1956)	**Class** Jung (1936)	**Average water content** Meisterfield (1977)
Hygrophilous	I	Open water or submerged vegetation average water content >95%
	II	floating vegetation, partly submerged partly at the surface; average water content >95%
	III	emerged vegetation, very wet, water drops without pressure >95% average water content
α-Hydrophilous	IV	wet, water drops with weak pressure, *ca.* 95% average water content
	V	half-wet, water drops out with moderate pressure, 95-85% water content
β-Hydrophilous	VI	moist, water drops out with strong pressure, average water content 80%
Xerophilous	VII	half-dry, a few drops with strong pressure <80% water content
	VIII	dry, no water drops at strong pressure, <50% average water content

Corbett (1973) defines xerophiles, hygrophiles and hydrophiles as follows:

Xerophiles: testate amoebae living in relatively dry habitats and able to withstand desiccation, usually by encysting. Typical species are able to survive in a thin film of water on drying moss, and include *Hyalosphenia subflava*, *Trigonopyxis arcula*, and *Bullinularia indica.* Xerophiles were considered to indicate areas of disturbance on bogs (Tolonen, 1986; Warner, 1987) but Woodland (1996) and Hendon (1998) have shown them to also inhabit dry areas of intact bogs.

Hygrophiles: testate amoebae living in moist plants and subject to desiccation less frequently than xerophiles. Typical species will normally inhabit an ample water film in which the tests can be spiny and can be carried upright without distorting the meniscus.

Hydrophiles: testate amoebae inhabiting plants submerged in water and are therefore not limited to the water film on *Sphagnum* leaves. Encystment is less common and the species can therefore afford to have large mouths.

Table 6.3 Taxa found in this study based on the taxonomy of Charman *et al.* (in press) and their hydrological requirements (modified from Hendon, 1998).

Taxonomic Group	Taxa included*	Peat water content/habitat	Author
Amphitrema flavum		wet conditions	Tolonen 1966
		87.9-95.1% water content	Tolonen *et al.* 1985
		>95% water content	Tolonen *et al.* 1992
		bog pools - wet	Warner 1987
		water table optimum 15.04cm	Warner & Charman 1994
		water table depth 4.51cm	Woodland, 1996
		water table optimum 10.27cm	Charman & Warner 1997
Amphitrema stenostoma		found with *A. wrightianum*	Corbet 1973
		bog pools	Tolonen 1986
		water table optimum 6.54cm	Charman & Warner 1997
Amphitrema wrightianum		bog pools - wet	Corbet 1973
		95% water content	Tolonen *et al.* 1992
		water table optimum 4.07cm	Charman & Warner 1997
Arcella artrocrea		hydrophilous	Tolonen 1986
Arcella catinus type	*A. catinus*	uncertain status	de Graaf 1956
		water table optimum 9.06cm	Warner & Charman 1994
		water table optimum 9.81cm	Charman & Warner 1997
	A. arenaria	xerophilous	Tolonen 1986
Arcella discoides type		submerged or very wet *Sphagnum*	Tolonen 1986
		>95% water content	Tolonen *et al.* 1992
		very wet	Warner 1987
		water table optimum 8.98cm	Warner & Charman 1994
		water table depth 3.11cm	Woodland 1996
		water table optimum 0.91cm	Charman & Warner 1997
	A. rotunda var. *aplanata*	hydrophilous	de Graaf 1956
Arcella vulgaris			
Assulina muscorum		common - both bog and fen	Corbet 1973
		Sphagnum	de Graff 1956
		xerophilous tendency	Tolonen *et al.* 1992
		85-90% moisture content	Warner 1987
		hydro-xerophilous taxon	Warner 1990
		water table optimum 42.53cm	Warner & Charman 1994
		water table depth 7.77cm	Woodland 1996
		water table optimum 17.44	Charman & Warner 1997
Assulina seminulum		hygrophilous	de Graff 1956
		bog hummocks	Corbet 1973
		water table optimum 39.29cm	Warner & Charman 1994
		water table depth 6.41cm	Woodland 1996
		water table optimum 13.95cm	Charman & Warner 1997
Bullinularia indica		xerophilous	de Graff 1956
		bog hummocks	Heal 1964
		xerophilous taxa	Tolonen 1966
		hummock *Sphagna*	Tolonen 1986
		xerophilous taxa <85% water content	Tolonen *et al.* 1992
		water table depth 12.9cm	Woodland 1996

Centropyxis aculeata type		aquatic habitats	de Graaf 1956
		aquatic habitats	Schönborn 1962
		hydrophilous taxa	Tolonen 1986
		very wet conditions	Warner 1987
		water table optimum 24.35cm	Warner & Charman 1994
		water table depth 4.44cm	Woodland 1996
		water table optimum 6.18cm	Charman & Warner 1997
Centropyxis cassis type	*C. cassis*	hygrophilous, submerged and wet mosses	de Graaf 1956
	C. aerophila type	water table optimum 20.53cm	Warner & Charman 1994
		water table optimum 5.16cm	Charman & Warner 1997
	C. aerophila var. *sphagnicola*		
	C. aerophila var. *sylvatica*		
Centropyxis arcelloides type	*C. arcelloides*	moderately dry, 78-89% water content	Warner 1987
		shallow peatland pools and very wet *Sphagnum* soils	Warner 1990
		water table depth 20.6cm	Tolonen *et al.* 1992
		water table optimum 32.01cm	Warner & Charman 1994
		water table depth 4.25cm	Woodland 1996
		water table optimum 4.69cm	Charman & Warner 1997
	C. eurystoma		
	C. minuta		
	Phryganella acropodia	moderately wet, 90=95% water content	Warner 1987; 1990
		dry mosses	Schönborn 1962
		moderately dry conditions	Meisterfield 1977
Difflugia acuminata		bog pools	Cash & Hopkinson 1909
		hydrophilous taxon	de Graaf 1956
Difflugia lucida type			
Difflugia oblonga type	*D. oblonga*	very wet *Sphagnum*	de Graaf 1956
		water table optimum 11.00cm	Warner & Charman 1994
		water table optimum 5.25cm	Charman & Warner 1997
Difflugia pristis type	*D. prisits*		
	D. angulostoma	aquatic	Cash & Hopkinson 1909
	Cryptodifflugia paludosa		
	D. fallax		
Difflugia pulex			
Difflugia rubescens		hydrophilous	de Graaf 1956
Euglypha ciliata			
Euglypha compressa			
Euglypha rotunda		water table optimum 38.91cm	Warner & Charman 1994
		water table depth 7.71cm	Woodland 1996
		water table optimum 18.41cm	Charman & Warner 1997
Euglypha strigosa		bog hummocks	Heal 1961
		water table depth 3.98cm	Woodland 1996
Euglypha tuberculata		water table optimum 36.55cm	Warner & Charman 1994
		water table depth 5.43cm	Woodland 1996
		water table optimum 14.43cm	Charman & Warner 1997
Heleopera petricola		very wet *Sphagnum*	de Graaf 1956
		ecology variable and disputed	Tolonen 1986
		ecology variable and disputed	Warner 1987
		95% water content	Tolonen *et al.* 1992
		water table optimum 31.08cm	Warner & Charman 1994
		water table depth 8.14	Woodland 1996
		water table optimum 7.27cm	Charman & Warner 1997
Heleopera rosea		bog hummocks and drier *Sphagnum*	Jung 1936
Heleopera sphagni		β-hygrophilous taxon	de Graaf 1956
		water table depth 7.19cm	Woodland 1996
Heleopera sylvatica		drier mosses	Tolonen 1986
		water table optimum 23.37cm	Warner & Charman 1994
		water table optimum 23.22cm	Charman & Warner 1997

Hyalosphenia elegans		α-hygrophilous	de Graaf 1956
		bog hummocks	Corbet 1973
		95% water content	Tolonen *et al.* 1992
		90-95% water content	Warner 1987
		water table optimum 26.26cm	Warner & Charman 1994
		water table depth 3.61cm	Woodland 1996
		water table optimum 21.59cm	Charman & Warner 1997
Hyalosphenia ovalis			
Hyalosphenia papilio		very wet *Sphagnum*	de Graaf 1956
		wet *Sphagnum* in bog hummocks	Heal 1961
		95% water content	Tolonen *et al.* 1992
		90-95% water content	Warner 1987
		>90% water content, moderately wet	Warner 1990
		water table optimum 9.87cm	Warner & Charman 1994
		water table depth 4.08cm	Woodland 1996
		water table optimum 17.74cm	Charman & Warner 1997
Hyalosphenia subflava		moderately dry 78-89% water content	Warner 1987
		water table optimum 49.92cm	Warner & Charman 1994
		water table optimum 22.81cm	Charman & Warner 1997
Nebela carinata		wet or very wet *Sphagnum*	Jung 1936
		bog pools	Heal 1964
		water table depth 3.9cm	Woodland 1996
		water table optimum 1.09cm	Charman & Warner 1997
Nebela collaris		moderately dry conditions	de Graaf 1956
		90-95% water content	Warner 1987
Nebela flabellulum		*Sphagnum* in bog hummocks	Corbett 1973
		water table optimum 30.04cm	Charman & Warner 1997
Nebela griseola		very wet to aquatic *Sphagnum*	Heal 1964
		water table depth 4.06cm	Woodland 1996
		water table optimum 11.14cm	Charman & Warner 1997
Nebela marginata		α-hygrophilous taxon	de Graaf 1956
		water table optimum 6.89cm	Charman & Warner 1997
Nebela militaris		drier mosses	de Graaf 1956
		wet *Sphagnum* of bog hummocks	Corbett 1973
		water table optimum 41.85cm	Warner & Charman 1994
		water table depth 7.68cm	Woodland 1996
		water table optimum 20.66cm	Charman & Warner 1997
Nebela minor			
Nebela parvula		very dry conditions	Warner 1987
		water table optimum 27.58cm	Warner & Charman 1994
		water table depth 9.17cm	Woodland 1996
		water table optimum 6.89cm	Charman & Warner 1997
Nebela tincta		xerophilous, <85% moisture content	Tolonen *et al.* 1992
		very wet	Warner 1987
		water table optimum 32.63cm	Warner & Charman 1994
		water table depth 6.18cm	Woodland 1996
		water table optimum 11.67cm	Charman & Warner 1997
Nebela tubulosa			
Nebela vitraea type			
Placosista spinosa		bog pools	Corbett 1973
		water table depth 5.36cm	Woodland 1996
		water table optimum 7.44cm	Charman & Warner 1997
Pseudodifflugia fasicularis		aquatic	Cash & Hopkinson 1909
Cryptodifflugia oviformis			
Sphenoderia lenta		α-hygrophilous - wet *Sphagna*	de Graaf 1956
		85-90% moisture content	Tolonen *et al.* 1992
		moderately dry 78-89% water content	Warner 1987: 1990
		water table optimum 5.86cm	Charman & Warner 1997
Trigonopyxis arcula		xerophilous taxa	de Graaf 1956
		water table optimum 57.68cm	Warner & Charman 1994
		water table depth 9.37cm	Woodland 1996
		water table optimum 15.58cm	Charman & Warner 1997
Corythion/Trinema type	*Trinema lineare*	hygrophilous taxa	de Graaf 1956
		water table depth 7.48cm	Woodland 1996

6.4 Detrended Correspondence Analysis (DCA)

Determining clear patterns in relatively complex ecological data sets is a difficult task, and merits statistical techniques. Detrended Correspondence Analysis (DCA) is an ordination technique, which is a collective term for methods which aim to order data along axes on the basis of species composition (ter Braak, 1995). The result of ordination is a two dimensional diagram, on which sites (samples in this case) which are similar in species composition are arranged close together, whereas points which are far apart correspond to sites which are dissimilar in species composition. An excellent review of ordination techniques is given by ter Braak (1995). The aim of this thesis is to use water table fluctuations as a proxy for climatic change. Hence species which are abundant under high water tables should be clustered together at one end of the diagram, at the opposite end from species which thrive under a low water table, creating a species gradient, from wet to dry.

The ordination technique used in this thesis (DCA) was developed by Hill and Gauch (1980), and is based on correspondence analysis (CA), which is probably the most popular ordination technique amongst ecologists (Digby and Kempton, 1987). It has been specifically developed to deal with data in the form of contingency tables, and is thus well suited to the data produced in Quaternary studies (Kovach, 1995), meriting its use in this thesis.

DCA, and most other ordination techniques, are based on *eigenanalysis*. This enables axes to be extracted from linear transformations of multidimensional data, which explain as much of the data as possible. Each axis is termed an *eigenvector* (a set of values denoting a point in multidimensional space) and has a corresponding *eigenvalue*. The eigenvalue is equal to the maximum dispersion of the species scores on the ordination axis, and is thus a relative measure of the importance of the axis, with the first axis having the largest eigenvalue, the second axis the second largest eigenvalue and so on (ter Braak, 1995).

DCA is an ordination technique which assumes an underlying structure within the data, suggested by the occurrences of species that are determined by a few unknown environmental variables, according to a simple response model (ter Braak, 1995). DCA assumes the relationships between the taxa and the environment are non-linear and *unimodal* (Gaussian-like), which commonly show bell-shaped response curves with respect to environmental gradients.

In order to understand how DCA works, it is necessary to comprehend how CA works. CA is based on reciprocal averaging (Hill, 1973) which is essentially a double weighted averaging procedure or two-way weighted averaging, which is also reviewed in detail by Pielou (1984). Weighted averaging (WA) is a highly regarded technique for reconstructing environmental variables from species assemblages (ter Braak and Juggins, 1993).

The fundamental algorithm in CA (which DCA also uses) is described by ter Braak (1995, p97-101). Initial arbitrary site scores are selected, and species scores are derived by calculating the weighted average of the site scores for each species. These species scores are then used to derive new site scores by calculating for each site the weighted average of the species scores. Then, new species scores are recalculated from the preceding site scores until the values of the species and site scores stabilise. This produces the first axis (eigenvector) which has the maximum eigenvalue and dispersion for each site. The second axis is created using the same algorithm with one extra step in which the trial scores for the second axis are made uncorrelated with the scores of the first axis, and so on for the subsequent axes.

However, CA is susceptible to two faults which are common in many ordination methods, the main one of which is referred to as the arch effect (or horseshoe effect). When analysing samples from diverse habitats along a single environmental gradient, the ordination on the principal axes does not always produce a linear sequence of species (Digby and Kempton, 1987). The curvilinearity may produce an arching effect, and when diversity is large, the ordination may produce a horseshoe shape, where the end points are involuted, which is simply a mathematical artifact corresponding to no real structure in the data (Hill and Gauch, 1980). The second fault, which is a product of the first, is the compression of data points at the ends of the axes, thus not preserving ecological distances (Kovach, 1995).

A complex pattern of abundances, or a large noise component within the data (the presence of curvilinearity), may obscure a gradient, or create a secondary gradient where none exists (Digby and Kempton, 1987). The curvilinearity exists because the second axis is in no way constrained to be independent of the first axis, and thus cannot be interpreted separately (Hill and Gauch, 1980). The method of reducing this curvilinearity is described in detail by Hill and Gauch (1980) as detrending the data. Detrending intends to ensure that at any point along the first axis the mean value of the site scores on the subsequent axes is about zero. To enable this, the first axis is divided into a number of segments, within which the site scores of axis 2 are adjusted by subtracting their mean. Subsequent axes are derived similarly by detrending with respect to each of the previous axes (ter Braak, 1995). When the data from DCA has only one underlying gradient of variation, they are said to form a 'coenocline': two underlying gradients form a 'coenoplane' (Hill and Gauch, 1980).

The range of the site scores along an ordination axis obtained from a DCA analysis are expressed as standard deviation (s.d.) units. Site scores are standardised so that within-site variance equals 1, and hence the tolerances of the curves of the species (maximum value of a bell shaped curve - Gaussian response model) will approach 1. A Gaussian response curve with tolerance 1 rises and falls over an interval of about 4 s.d. (ter Braak and Looman, 1995). Thus, a first axis which has a length approaching 4 s.d. tends to have few species in common at opposite ends, and could be said to represent a complete turnover in taxonomic composition of samples. A change of 50% in sample composition occurs in about 1 s.d. unit or slightly more. This can be seen in Figure 6.2.

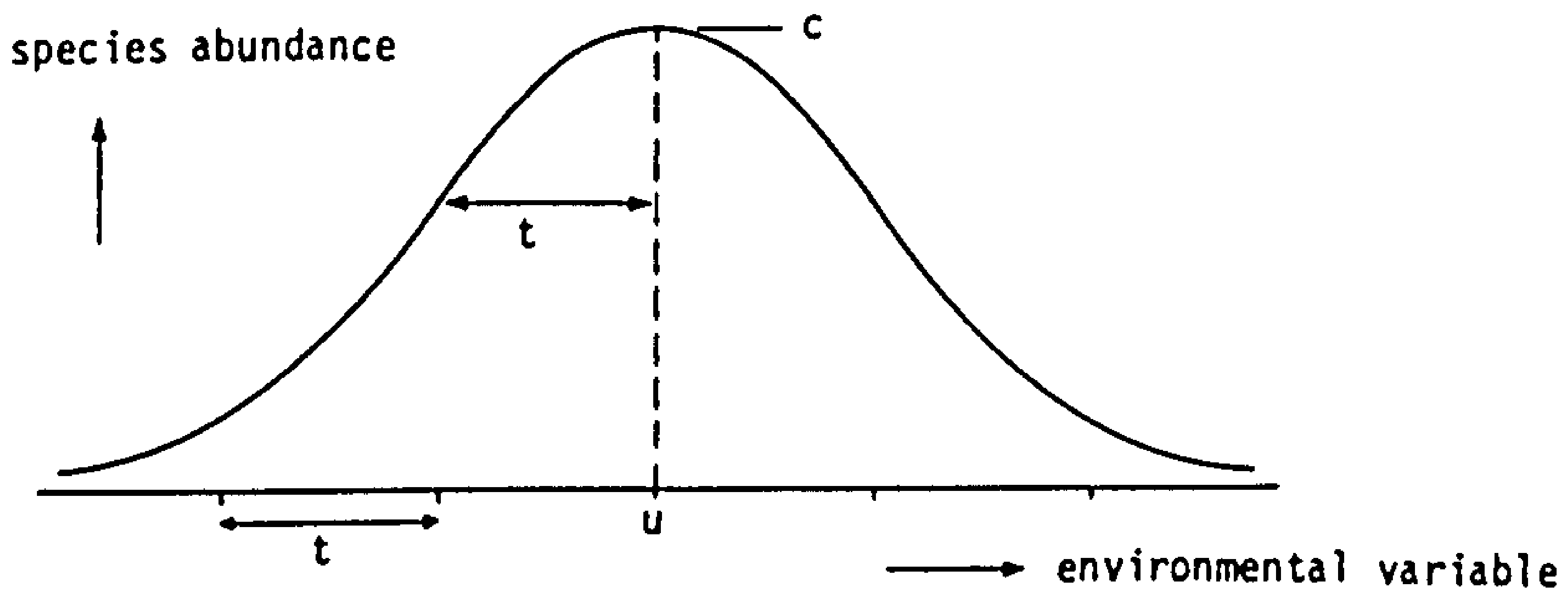


Figure 6.2 Taxon-environment response model. Gaussian unimodal relationship with its three ecologically important parameters: maximum (c), optimum (u) and tolerance (t). The range of occurrence of the species is seen to be about 4t (after ter Braak and Looman, 1995, p42).

DCA thus appears to be an excellent tool for extracting patterns from relatively complex data sets, and Hill and Gauch (1980, p.56) comment on the good attributes of DCA, in that:

> "It provides an interpretable species ordination as well as a sample ordination. The axes are scaled in units (sd) that have definite meaning. The arch effect is avoided."

However, Hill and Gauch (1980) also document the bad features of DCA. The most persistent difficulties are coping with outliers and discontinuities, and it is suggested the only way to deal with extreme outliers is to remove them. Interpreting the axes is a problem which can never be eliminated, although in practice environmental data may permit the interpretation of the axes with a useful meaning. Another problem with DCA (and multivariate analyses in general) is that optimum results are often achieved when a large number of species are represented in the data sets as opposed to one species dominating. The potential problem with this study is that the data sets are often dominated by one species, usually *Sphagnum imbricatum* or *Sphagnum* section *Acutifolia*. Hill and Gauch (1980) suggest this problem can be overcome by transforming the data to reduce the effect of the dominant species, although they also suggest trying a first ordination without transformed data. Hill and Gauch (1980) also suggest downweighting the rare species and removing species occurring in fewer than a threshold number of samples (e.g. 5% of the total sample number).

One of the major problems with the use of DCA in this study is that it was designed for species ordinations using a Gaussian approach. Hence, the species plots are unimodal, and may therefore, at the variable end of the gradient, become excessively polarized (Hill and Gauch, 1980). Some of the taxa analysed in this project have bimodal distributions, notably *Sphagnum* section *Acutifolia*, *Sphagnum imbricatum*, and to a certain extent UOM. Thus, if DCA plots one of these in an extreme position along a moisture gradient, care must be taken when interpreting the data.

DCA has been notably used to reconstruct latent water level gradients by Barber *et al.* (1994b) and Mauquoy and Barber (1999a) from Cumbrian mires, as the technique is relatively simple, has good predictive power, and the underlying assumption of a unimodal species response carries ecological plausibility. Mauquoy and Barber (1999a) suggest that eigenvalues below 150 (1.5 s.d. units) in the DCA axis 1 scores can be identified as a wet-shift, assuming the axis length is in the order of 4 s.d. units, simulating a Gaussian response. A more cautious approach is taken here, as changes must be viewed with respect to the original data, as when the data are dominated by one species DCA can perform sub-optimally. Hence, with the problems noted above, a change of 1 s.d. unit (representing a change of 50% in sample composition) may be significant in terms of a climatic shift.

The DCA construction involved TILIA data being entered into TRAN, which enabled the data to be transformed into Condensed Cornell format, so it could be incorporated into CANOCO (ter Braak, 1987) which was used for the analyses. It has been suggested that detrending-by-segments (Hill and Gauch, 1980) can result in the loss of some ecological information (Pielou, 1984; Minchin, 1987), and a more stable method is recommended, namely detrending-by-polynomials (ter Braak and Prentice, 1988; ter Braak, 1995). This option is available in CANOCO, although when detrending-by-polynomials was attempted the species scores tended to cluster around the origin, with the exception of the dominant species, suggesting no discernible gradient was present.

It thus appears that the dominant species within the stratigraphy dictates the response of the DCA using this approach, as indicated by Hill and Gauch (1980). Transforming the data (as suggested by Hill and Gauch) appeared to have little affect on the ordination. However, when undertaking detrending-by-segments on the data in this thesis (which Hill and Gauch indicate should be done first with any data set), significant moisture gradients could be recognised from the ordinations from each site, indicating that the caveats identified by Pielou (1984) and Minchin (1987) are not applicable to these data.

The data obtained from the DCA procedures are presented as biplots, with sample scores being derived from the WA analysis of species scores. Hence, when sample scores are plotted near to species scores, the sample must have high values of that species. Thus, species which are poorly represented occur on the periphery of the axes. All rare species were downweighted in the DCA experiments, and in cases where species were represented by only one sample, they were removed, as suggested by Hill and Gauch (1980).

The extinction of *Sphagnum imbricatum*, and subsequent replacement by *Sphagnum magellanicum* has been picked up as a large change in species composition in DCA by Stoneman (1993), Barber *et al.* (1994b) and Mauquoy and Barber (1999a). As this species replacement is not related to a moisture gradient, it can potentially mask the detection of latent changes in mire water table depths. Thus, following the advice of Stoneman (1993), who considered both species to inhabit the same niche, *Sphagnum imbricatum* and *Sphagnum magellanicum* have been grouped together as one species unit (*S.imb/mag*), in order to aid the chance of detecting a mire surface wetness gradient.

Weighted Averages Ordination (WAO) is another simpler method of locating patterns in the data, and follows the indicator values approach of Dupont (1986), often referred to as Dupont indices. A single value is given to each species, and the results plotted down core (e.g. see Barber *et al.*, 1994c; Mauquoy, 1997). However, Dupont indices have not been adopted in this study, as although there is never any difficulty in interpretation with this method, assigning a single weight to each taxon simplifies the data too much (Mauquoy, 1997), especially as some of the taxa that can dominate the assemblages have bimodal response curves, resulting in a complacent proxy climate signal.

6.5 Testate amoebae transfer function

The transfer function analyses on the testate amoebae data were undertaken by D. Charman (Plymouth University). Weighted averaging (WA) analysis was undertaken using the same training set as discussed in Woodland *et al.* (1998).

Woodland *et al.* (1998) were able to show the strong relationship between testate amoebae assemblages and hydrological parameters, and used four methods in an attempt to evaluate these relationships using the program CALIBRATE (Juggins, unpublished). After a screening process whereby 'rogue' samples are identified and removed from the data (Birks *et al.*, 1990) it was found that WA and WA-PLS (weighted averaging partial least squares) produced the smallest prediction errors for water table depths. However, WA was deemed to be advantageous as it is relatively simple computationally and easily interpreted ecologically. WA regression and calibration has been summarised by ter Braak and Prentice (1988).

The prediction errors of the training sets, as well as fossil samples, are important when deciding which method best suits the data. The root mean square error of prediction (RMSEP) is a measure of the overall predictive abilities of the training set, which can be used to compare methods. RMSEP for individual fossil samples (as well as for all individual modern training samples) can be derived using a technique called *bootstrapping*, which is a computer-intensive resampling procedure (Birks *et al.*, 1990).

Bootstrapping for the estimation of sample-specific errors has been implemented for WA regression and calibration (Birks *et al.*, 1990; Line *et al.*, 1994). The reconstructed values are presented +/- the sample specific errors of prediction from 1000 bootstrap cycles using WACALIB version 3.3 (Line *et al.*, 1994).

A descriptive method of bootstrapping is summarised by Line *et al.* (1994) and Birks (1995), as follows. A bootstrap cycle randomly selects a subset of training samples but with replacement from the original set, creating a bootstrap training set of the same size as the original training set, and thus mimicking sampling variation in the training set. Some samples may be selected more than once in a cycle as sampling is with replacement, but any samples which are not selected for the training set form a bootstrap *test* set for that cycle. WA regression and calibration are then carried out with the bootstrap training set to infer the environmental variable of interest (x_i) for the modern samples (which is known) in the bootstrap test set. WA calibration is also used to infer the environmental value (x_0) for each fossil sample in each cycle, and the standard deviation values for both modern and fossil samples are calculated. This constitutes one component of the prediction error (s_1), being the estimation error in the optima and tolerance of the taxa. The error due to variation of abundance of taxa at a given environmental value (the second component - s_2) is estimated from the training set by the root mean square (across all training samples) of the difference between observed x_i and the mean bootstrap estimate in all bootstrap cycles when that modern sample is in the bootstrap test set. The composition of the fossil assemblage will affect the variability of s_1 between fossil samples, whereas s_2 remains constant for fossil samples. The estimated RMSEP for a fossil sample is calculated as the square root of the sum of squares for the two components. The underlying theory of bootstrapping is presented by Birks *et al.* (1990).

6.6 Longbridge Moss

6.6.1 Stratigraphy

Three profiles covering the top 3 metres of ombrotrophic peat were examined in a northwest-southeast transect on Longbridge Moss, to compliment the master core taken from near the centre of the mire, and are shown in Figure 6.3 (plotted using the TSPPlus program, Waller *et al.*, 1995). The Troels-Smith (1955) stratigraphic system and von Post (1924) humification scale were used in the field as they provide a quick and reliable method of obtaining a first order interpretation of changes in stratigraphy.

Stratigraphic profiles were taken in order to show that the wet and dry phases of peat growth can be detected laterally throughout the bog (following Walker and Walker, 1961; Barber, 1981; Barber *et al.*, 1998). Species changes from ombrotrophic bogs, which have reacted synchronously to climatic forcing have been demonstrated by Barber (1981), Stoneman (1993), Svensson (1988a), Barber *et al.* (1998), and Hendon (1998). The stratigraphic data from each site analysed is presented in Appendix 1.

Examination of the gross stratigraphy shows periodic wet and dry phases throughout the last *ca.* 4000 years. The wet phases within the stratigraphy are represented by the bands of Sh (*Substantia humosa*). In the field these layers are recognised as algal muds, often with layers of fresh (H2-3) *Sphagnum cuspidatum* leaves and *Rhynchospora alba* remains indicating pool conditions. For simplification in the Figures in this study the pool layers are represented by Sh, and not *Sphagnum* (Tb Spag - *Turfa bryophytica*) or sedges (Th - *Turfa herbacea*) in order to differentiate them from the rest of the stratigraphy (see Aaby and Berglund (1986) for a full description of peat sediments). Many of these pool layers can be seen to correlate well between cores. Occasionally, there is an offset of a few centimetres between cores, probably due to microtopographic differences and slight changes in rates of growth across the bog (see also Chapter 5, section 5.2.2). One of the layers from the master core around 14.30m O.D. does not appear to be represented on the other diagrams. The macrofossil diagram (Figure 6.4) shows this level to be dominated by *Sphagnum papillosum*, indicating a return to a more wet lawn microtype. However, in the field it is difficult to differentiate between the

Sphagnum section *Sphagnum* species, and these changes are therefore not detected in cores LBM1-3.

There is an increase in *Turfa herbacea* (Th, especially *Eriophorum vaginatum*) below 13m O.D. in the stratigraphy. This is shown particularly in cores LBM1, LBM3, and the master core. Only a slight amount of monocots (Th+) is found in LBM2, suggesting that this was possibly a pure *Sphagnum* hummock at this time.

6.6.2 Plant macrofossils

The plant macrofossils for Longbridge Moss are shown in Figure 6.4. Three main zones are described, the pre-*Sphagnum imbricatum* zone (LBM-a), the *Sphagnum imbricatum* zone itself (LBM-b) and the post *Sphagnum imbricatum* zone (LMB-c).

Zone LBM-a (300-254cm): The Sphagna are dominated by *Sphagnum* section *Acutifolia*, apart from a major decline between 288-276cm. This decline in Sphagna corresponds with an increase and subsequent decline in Ericaceae, which lasts for *ca.* 200 years according to the accumulation rate, followed by peaks in *Aulacomnium palustre* and then *Dicranum scoparium* around the time of the Hekla-4 ash layer. The decline in Ericaceae and subsequent rise in pleurocarps correlates with relatively large amounts of macroscopic charcoal, an indication that fires may have occurred. Kuhry (1994) reports a postfire sequence in which *Sphagnum* is replaced by brown mosses, notably *Aulacomnium palustre*, *Polytrichum strictum*, and *Pohlia*. At Longbridge it appears that a possible hummock microform had suffered, as the Ericaceae appears to have been burnt, Sphagna declined, and *Aulacomnium palustre*, which was already present in smaller amounts, appears to have established itself on the hummock before the other vegetation could recolonise. After the fire, *Sphagnum* section *Acutifolia* was able to recolonise, possibly due to a climatic deterioration, although *Eriophorum vaginatum* soon established itself, before Ericaceae eventually dominated again. The presence of flora which can withstand low water tables in this zone suggests a dry climatic phase, although the rise in Sphagna after the fire may indicate a rise in the water table. These climatic inferences are explored further when the plant macrofossil record is compared with the other proxy reconstructions.

Zone LBM-b (254-46cm): *Sphagnum imbricatum* appeared on the mire surface before the start of this zone, having been counted at very low levels between 288-280cm, and it may be possible that it could not establish itself on the mire due to its intolerance of fires (Slater and Slater, 1978; Boatman, 1983). It becomes established between 256-252cm (dated to 2130-1780 cal. BC: SRR-6330) as Ericaceae decline, and there is a small peak of *Sphagnum* section *Cuspidata* just before *Sphagnum imbricatum* becomes established, suggesting a climatic wet shift. The decline in Ericaceae is probably related to the raised water tables, which would have created the niche for *Sphagnum imbricatum* to exploit. Although the levels of *S.s. Cuspidata* appear low at 256cm (Figure 6.4), due to low levels of Sphagna being present (18%), 50% of the 100 leaves counted were of section *Cuspidata.*

Sphagnum imbricatum is the dominant species within the zone, as with so many other bogs in Cumbria and the Scottish Borders (Barber, 1993; Stoneman, 1993; Mauquoy and Barber, 1999a), but is occasionally replaced by other taxa throughout the stratigraphy. A notable decline in *Sphagnum imbricatum* occurs between 216-196cm, and it is replaced by *S.s. Acutifolia* and monocots undiff., with peaks in *S.s. Cuspidata* at the start and end of the *S. imbricatum* decline. A small amount of *Eriophorum vaginatum* is present at the same levels as the section *Acutifolia* peak, but it is suggested that the decline in *Sphagnum imbricatum* is due to a climatic wet phase, enabling other species to invade where the *S. imbricatum* could no longer survive.

Sphagnum imbricatum became re-established at 196cm, accompanied with high levels of Ericaceae, indicating lowered water tables and a possible hummock microform. From 166cm to the top of the stratigraphy Ericaceae levels are negligible, and the decline at 166cm may represent a raising of the water table, indicating a climatic deterioration. Within the rest of zone LBM-b *Sphagnum imbricatum* dominates the stratigraphy, being replaced by more hygrophilous *Sphagnum* species at 144cm (*Sphagnum papillosum,* dated to 400-200 cal. BC: SRR-6328), 112cm (*Sphagnum* section *Cuspidata,* dated to cal. AD 70-330: SRR-6327), and 68-64cm (*Sphagnum papillosum,* dated to cal. AD 950-1170: SRR-6326), suggesting climatic deteriorations at these times. It is possible that the *Sphagnum imbricatum* at this time was growing over a lawn microform, with pools expanding at times of higher effective precipitation, being represented by the more hygrophilous species in Figure 6.4. At 96cm, however, the high abundance of *Racomitrium lanuginosum* would suggest a hummock was being colonised, with a correspondingly low water table. Tallis (1995) suggests that *Racomitrium lanuginosum* macrofossils from blanket peats can signify a shift from a dry climate to a wetter one, as the moss is predominately oceanic, but can only compete and grow with Sphagna when the water table is low. However, there is no macrofossil evidence for a shift in climate immediately after the decline of *Racomitrium.*

Zone LBM-c (46-0cm): The extinction of *Sphagnum imbricatum* occurs between 48-44cm, and is associated with an increase in *Sphagnum papillosum*, with very small amounts of *Sphagnum* section *Cuspidata*, and is dated at cal. AD 1030-1230 (SRR-6325). This change in species assemblage is common in raised bogs (Mauquoy and Barber, 1999b), and could be associated with a shift to a lawn microform, being associated with a higher effective precipitation regime. *Sphagnum papillosum* then becomes superseded by *Sphagnum magellanicum*, which is dated at cal. AD 1280-1400 (SRR-6324). A tussock of *Eriophorum vaginatum* corresponds with a drop in Sphagna percentages between 24-16cm, possibly indicating a lowering of the water table. The *Eriophorum* tussock is overgrown by *Sphagnum magellanicum* between 20-16cm, possibly representing a return to wetter conditions, as small quantities of *Sphagnum papillosum* are also present, with low amounts of UOM and Ericaceae.

6.6.3 Humification

A clear increase in absorbance values can be seen towards the base of the core (Figure 6.5). This is expected as Clymo (1984) has shown that the peat continues to decay at depth, although once incorporated into the catotelm it decays at a much slower rate compared to the aerated catotelm. This increase in absorbance values is also in accordance with the findings of Blackford (1990), Blackford and Chambers (1995), Chambers *et al.*, (1997a) and Mauquoy and Barber (1999a). The curve was detrended by fitting a linear regression model to the data and then plotting the derived residuals, as shown in Figure 6.5. Using a time series created from the age/depth model (from interpolating the radiocarbon and tephra based chronology, see section 5.3.1) the changes in humification can be plotted against time (Figure 6.6).

Defining exact phases or shifts of climatic change from proxy reconstructions can be problematic. Major changes in humification were defined by Blackford (1990) as those that mark points beyond which the curve did not return, and were suggested as being a significant climatic change. In this thesis, where a humification curve, or other proxy reconstruction, displays a sharp deviation, the extreme point(s) will be taken as a climatic phase, as it is considered that the climate will actually be within that phase (wet or dry) at these points, and not midway through changing. As the climatic change may be a rapid response, it was considered dangerous to take the phase change from the halfway point on a curve, especially as this may be between samples. When a phase is represented by only one sample, the date is assumed to have a possible error range of ±50 years to compensate for sampling intervals. Where a more gradual change takes place, a range of dates may be quoted when the climate is considered to be within a specific phase. An example of this is discussed in Chapter 7. The greyscale lines in the site synthesis sections are *not intended as the definitive phases of change*, but to suggest that major changes can be identified at these times, and are intended to aid discussion. Minor changes of a less prominent nature are considered as noise within the structure of the data.

Clear phases of lower humification (Figure 6.6) can be distinguished around the interpolated dates of *ca.* cal. 1820 BC, 1630 BC, 1440 BC, 1000 BC, 570 BC, and less markedly around *ca.* cal. AD 750, AD 1300, and *ca.* cal. AD 1680. These are interpreted as times when peat passed quickly through a shallow acrotelm, indicating a raised water table and therefore a high effective precipitation regime. High levels of humification are thus interpreted as times of drier climate, and can be seen clearly around *ca.* cal. 2310-2000 BC, 1320-1200 BC, 750 BC, AD 500-620, and *ca.* cal. AD 1500-1620.

6.6.4 Testate amoebae

The same zones have been used for the testate amoebae data (Figure 6.7) as for the macrofossil data, allowing a direct comparison between the two techniques. This approach will be used at each site. The stratigraphy is dominated by *Difflugia pulex*, a small testate (15-30μm length, this study; Hendon, 1998; Charman, *et al.*, in press), for which there is no hydrological data in the literature (Penard, 1902 describes the taxon, but give no details of habitat requirements). The taxon was not found in samples used to construct the modern analogue transfer function (Woodland, 1996; Woodland *et al.*, 1998). In order to gain an initial idea about the hydrological requirements of this taxon, Hendon (1998) plotted the abundance of *Difflugia pulex* against the reconstructed water table depth derived from the analogue taxa in each horizon from which *Difflugia pulex* was found. There was, however, a circularity with this approach and the systematic error arising from the data, but Hendon stressed that it was the best available method at the time. These plots revealed *Difflugia pulex* to have a cosmopolitan distribution with depths to the water table ranging between 1.3cm to 16.5cm. Hendon (1998) noted that although most samples of *Difflugia pulex* were found in the 4-6cm water table class, this value is probably wetter than their actual value, as most of the 'drier taxa' are poorly represented in the transfer function. Analysis of the ordination plots carried out by Hendon (1998) showed *Difflugia pulex* to be associated with *Hyalosphenia subflava* in seven out of eight species ordination plots, as well as being associated with *Assulina muscorum*, *Bullinaria indica*, and *Difflugia pristis* type, suggesting it is representative of the 'drier' end of the hydrological scale. *Difflugia pulex* will therefore be interpreted with caution, but when it occurs in abundance, a drier hydrological regime will be inferred.

Zone LBM-a (300-254cm): The initial conditions at 296cm are of high levels of *Amphitrema flavum* (>45%) with levels of *Difflugia pulex* at *ca.* 20%, possibly indicating intermediate water table levels. However, high test concentrations, notably of *Difflugia pristis* type and *Assulina muscorum* were found at 288cm as *Amphitrema flavum* fall to near zero, implying a lowering of the water table. This level coincides with a peak in Ericaceae macrofossils, supporting the testate evidence. From this level to the end of the zone there are decreasing concentrations of testates to the lowest levels for the whole of the profile, probably as a result of the fires which were inferred from the macroscopic charcoal in Section 6.6.2. Most species decline in abundance, with the exception of *Amphitrema flavum*, which increases from 288-264cm. This may be due to rising water levels, or be a function of test resistance to burning. Lousier and Parkinson (1981) suggest that chitinous tests (e.g. *Amphitrema* species) appear to be the most resistant to decay in soils, hence possibly influencing their dominance at these depths in peats.

There is a notable increase in the hydrophilous taxa *Arcella discoides* type and *Centropyxis aculeata* type around 272cm, when *Difflugia pulex* values are less than 5%, indicating a rise in water tables (and coincident with an increase in Sphagna, Figure 6.4), either due to a climatic deterioration, or as a result of an impermeable layer of charcoal on the mire surface (Mallik *et al.*, 1984). This would have enabled water to pond up, and initiated *Sphagnum* growth, as seen in Figure 6.4.

Zone LBM-b (254-46cm): This zone shows four major phases of high *Amphitrema wrightianum* abundance, suggesting high mire water tables, interspersed with four phases of *Difflugia pulex* dominance, suggesting lower water tables. A significant increase in the hydrophilous *Amphitrema wrightianum* at the start of this zone suggests that water tables were high, implying an increase in effective precipitation. This climatic phase coincides with the expansion of *Sphagnum imbricatum*, as seen in Figure 6.4. Wet conditions, as indicated by *Amphitrema wrightianum*, seem to prevail until *Difflugia pulex* increases in abundance between 240-232cm, the first of the four major expansions of this taxon within this zone. A lowering of *Difflugia pulex*, and corresponding increase in abundance of *Amphitrema wrightianum* and *Amphitrema flavum* between 224-208cm suggests a high water table, until *Difflugia pulex* increases in abundance for the second time, up until 168cm. This second stage of *Difflugia pulex* domination is associated with the occurrence of the xerophilous *Trigonopyxis arcula*.

At 160cm *Difflugia pulex* is replaced by *Amphitrema flavum*, although higher than average levels of *Assulina muscorum* are also present, suggesting a more intermediate water table depth. From 152-144cm to 112-104cm high levels of *Amphitrema wrightianum* fluctuate with high levels of *Difflugia pulex* and *Amphitrema flavum*. This is therefore possibly a time of rapidly fluctuating water tables, dated to between *ca.* cal. 400-200 BC (SRR-6328) to *ca.* cal. AD 70-330 (SRR-6327).

After 104cm and until 80-72cm, *Difflugia pulex* dominates the stratigraphy once again, peaking with an abundance of just below 80% around 88cm. The water table is thus suggested to be low during this time, inferring a dry climatic regime. At around 72cm *Difflugia pulex* is replaced by *Amphitrema wrightianum* which are most common in pools (Heal, 1961; Tolonen *et al.*, 1992), thus inferring wet conditions. This peak in *Amphitrema wrightianum* corresponds with the only major peak in UOM, suggesting that the UOM is from an algal mud. The peak in *Amphitrema wrightianum* is short lived, and *Difflugia pulex* becomes re-established once more from 64cm. However, the abundance of *Difflugia pulex* decreases from 64cm to the top of the core, from which point onwards the taxon never rises above 20% occurrence. An increase in *Arcella discoides* type around 56cm, corresponding with a comparative decrease in *Difflugia pulex* may indicate a significant rise in the water table.

Zone LBM-c (46-0cm): At the start of this zone there are significant increases in *Amphitrema wrightianum* and *Amphitrema flavum* testacea, suggesting a rise in the water table, and inferring a climatic deterioration. However, these taxa both decline by 32cm, being replaced mainly by *Cyclopyxis arcelloides* type, until 16cm. The literature suggests that this group of testacea live in more intermediate water table depths, although the macrofossils show an *Eriophorum vaginatum* tussock present in the stratigraphy between these depths, indicating that water table depths were probably lowered or fluctuated rapidly. The top of the stratigraphy records the largest variety of taxa for the whole profile with comparatively numerous *Nebela militaris* testacea indicating low water table depths. Hendon (1998) found that the top 1m to 1.5m of each core from Coom Rigg Moss had a better diversity of taxa than the lower peats, and hence produced better hydrological reconstructions, as there were more species with modern analogues.

6.6.5 DCA

A clear differentiation between the mire hygrophilous taxa (*Sphagnum papillosum* and *Sphagnum* section *Cuspidata*), and the more xerophilous components (Ericaceae and UOM) can be seen on the DCA ordination (Figure 6.8). The ordination places *Sphagnum imbricatum/magellanicum* (*S.imb/mag*) to the far right of axis 1, amongst the drier taxa suggesting that it represents a comparatively drier community. Evidence from the testate amoebae is in agreement, as LBM-b (the phase of *Sphagnum imbricatum* growth) is dominated by relatively high abundances of the xerophile *Difflugia pulex*. The placement of *Sphagnum* section *Acutifolia* at the opposite end of axis 1 from *Sphagnum imbricatum/magellanicum* reflects the dominance of *Sphagnum imbricatum* within the stratigraphy, and suggests that the two species grow in mutual exclusion, which can be seen in the macrofossil diagram (Figure 6.4). The replacement of *Sphagnum imbricatum* in LBM-b by the more hygrophilous *Sphagnum papillosum* and *Sphagnum* section *Cuspidata* is thus represented as wet shifts in the DCA axis 1 scores. The axis 1 scores have been plotted against time (Figure 6.9) from the interpolated dates, and as axis 1 represents a moisture gradient, the plot can be interpreted as a proxy effective precipitation curve.

Wet phases are reconstructed as occurring around *ca.* cal. 2630 BC, 2250-2130 BC, 1380-1250 BC, 250 BC, AD 240, AD 930-1000, AD 1310-1620, and *ca.* cal. AD 1870-1995. Comparatively drier phases are reconstructed around *ca.* cal. 2500 BC, 1940-1820 BC, 1690-1440 BC, 1070-1010 BC, 690 BC, 570-320 BC, 130 BC – AD 180, AD 370-810, AD 1060-1240, and *ca.* cal. AD 1810.

6.6.6 Testate amoebae transfer function

The mean reconstructed water table depths with lower and upper limits to the sample prediction errors has been plotted against the age/depth model with interpolated dates for each sample depth (Figure 6.10). The first thing to note about the reconstructed water table levels is the relatively small amount of change within the error margins, the largest shift being from a relatively low water table at *ca.* cal. AD 500 (mean depth -7.5cm) to a relatively high water table depth *ca.* cal. AD 870 (mean depth 0cm). This change follows the decline in *Difflugia pulex* after 88cm, and subsequent replacement by the more hydrophilous *Amphitrema wrightianum* (Figure 6.7). The fact that this large change in species assemblage is only recognised as a 7cm change in water table depth, which is relatively small compared to the other sites, may be an artifact of the lack of a modern analogues for *Difflugia pulex*, which is discussed further in Chapter 7.

Other notable wet shifts occur at *ca.* cal. 2250 BC, 880 BC, 250 BC, AD 250, and *ca.* cal. AD 1370. Shifts to a lower water table depth, and hence drier climate can be identified at *ca.* cal. 2500 BC, 1000 BC, 750 BC, AD 120, AD 500, and *ca.* cal. AD 1870, which correspond with peaks in *Difflugia pulex* (Figure 6.7).

6.5.7 Comparison of the three proxy techniques

The three proxy records of climatic changes have been drawn together in a composite figure, all being plotted against time from the interpolated age/depth model (Figure 6.11). The greyscale lines indicate periods when each record is showing a tendency towards increased wetness, and probably a climatic deterioration (or two of the three records indicate strongly there has been a climatic deterioration). Wet phases can be distinguished around *ca.* cal. 2250 BC, 1380 BC, 940 BC, 250 BC, AD 870, and *ca.* cal. AD 1350. All three proxies agree remarkably well for these dates, with the major exception of the humification record, which does not agree with the wet phases at *ca.* cal. 2250 BC and *ca.* cal. 1380 BC, and suggests a drier phase at these times. However, the presence of relatively high levels of *Centropyxis aculeata* type and *Arcella discoides* type (Figure 6.7), along with an increase in *Sphagnum* abundance (Figure 6.4) at *ca.* cal 2250 BC indicates that the water tables were relatively high. At *ca.* cal. 1380 BC there was a change in species assemblage from *S. imbricatum* to *S.*s. *Cuspidata* and high levels of *S.*s. *Acutifolia* (Figure 6.4), coupled with a replacement of the xeric *Difflugia pulex* with the more hygrophilous *Amphitrema flavum* and *Amphitrema wrightianum* (Figure 6.7), indicative of a wetter climate. There is a possibility that the records could be out of phase, but the evidence presented by Mauquoy and Barber (1999a) suggest humification and plant macrofossil records agree very well. A more likely explanation is that the humification curve is responding to specific macrofossils, namely the monocotyledons. Both peaks in humification at these times correspond with relatively high levels of monocots, the peak around *ca.* cal. 2250 BC corresponding with *Eriophorum vaginatum* (which can grow over a wide range of moisture gradients - see section 6.1.4), and the second peak around *ca.* cal. 1380 BC corresponds with a peak in monocots undiff., which could relate to the more hygrophilous *Rhynchospora alba* or *Eriophorum angustifolium*. Mauquoy (1997) also noted high levels of humification corresponding with the presence of monocotyledons.

Identifying dry climatic shifts is difficult in peat records, due to the potential of a cessation in peat growth or erosion associated with the dry climate. However, phases when the climate was relatively dry can be identified from the three proxy records in Figure 6.11, and occur around *ca.* cal. 1570 BC, 750 BC, AD 430-680, and *ca.* cal. AD 1250.

The following applies to all TILIA diagrams (plant macrofossil and testate amoebae): Depth is measured in centimetres; the dates on the left hand side are calibrated radiocarbon dates, either from the dated peat samples, or from the dated tephras; the curves on the plant macrofossil diagrams are all percentages of the peat samples; the bar graphs of the other components are on a 1-5 scale as described in Chapter 3; the curves on the testate amoebae diagrams are all percentages of the total count; the pine pollen curves on these diagrams relate to the actual pine pollen grains counted and are independent of amoebae percentages.

Figure 6.3 Longbridge Moss: Stratigraphy

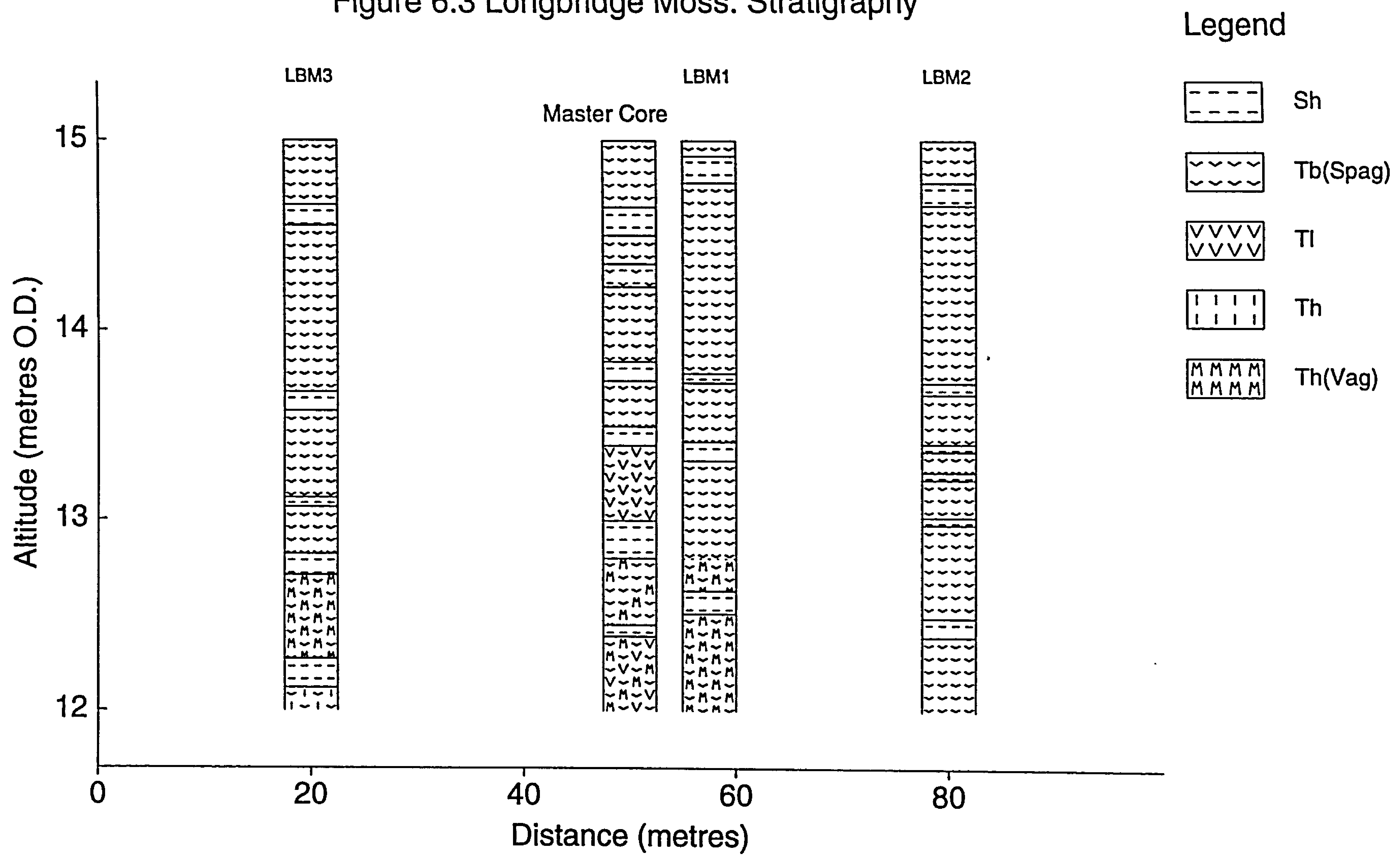
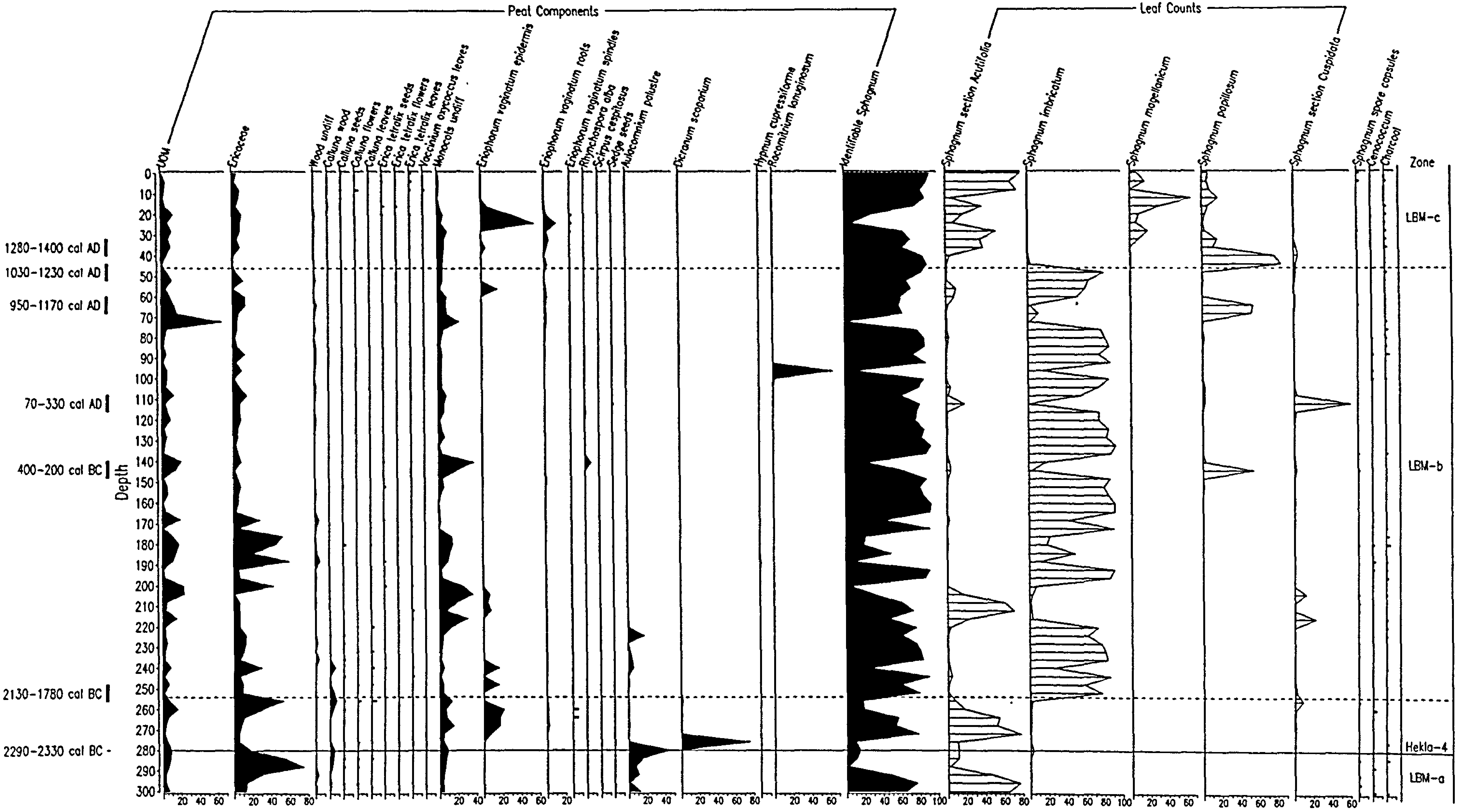


Figure 6.4 Longbridge Moss – Macrofossil Diagram



Analysis: Pete Langdon

Figure 6.5 Longbridge Moss - Humification data. The raw data are presented at the top, and have been detrended by linear regression, producing the lower data set.

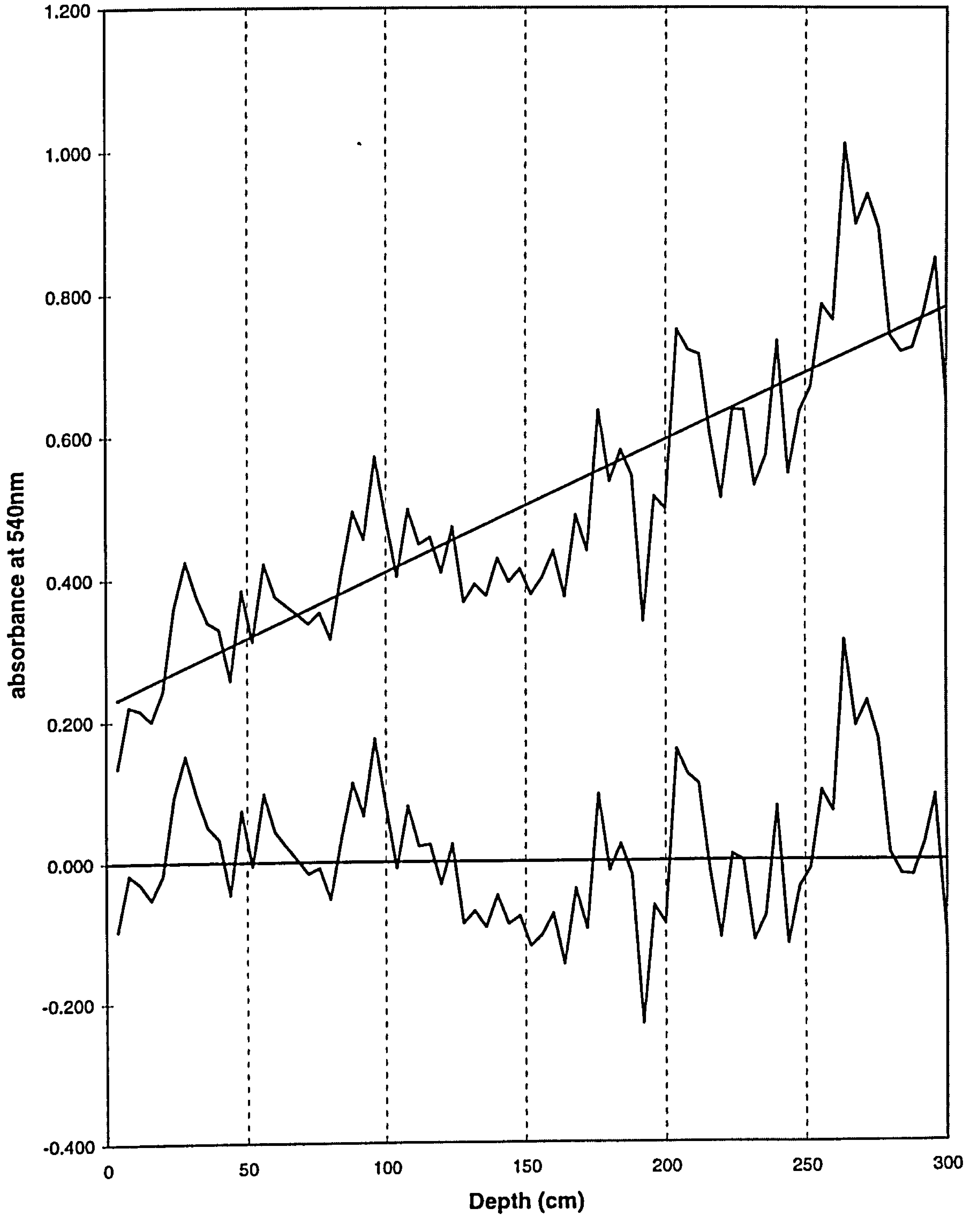


Figure 6.6 Longbridge Moss - Detrended humification vs time.

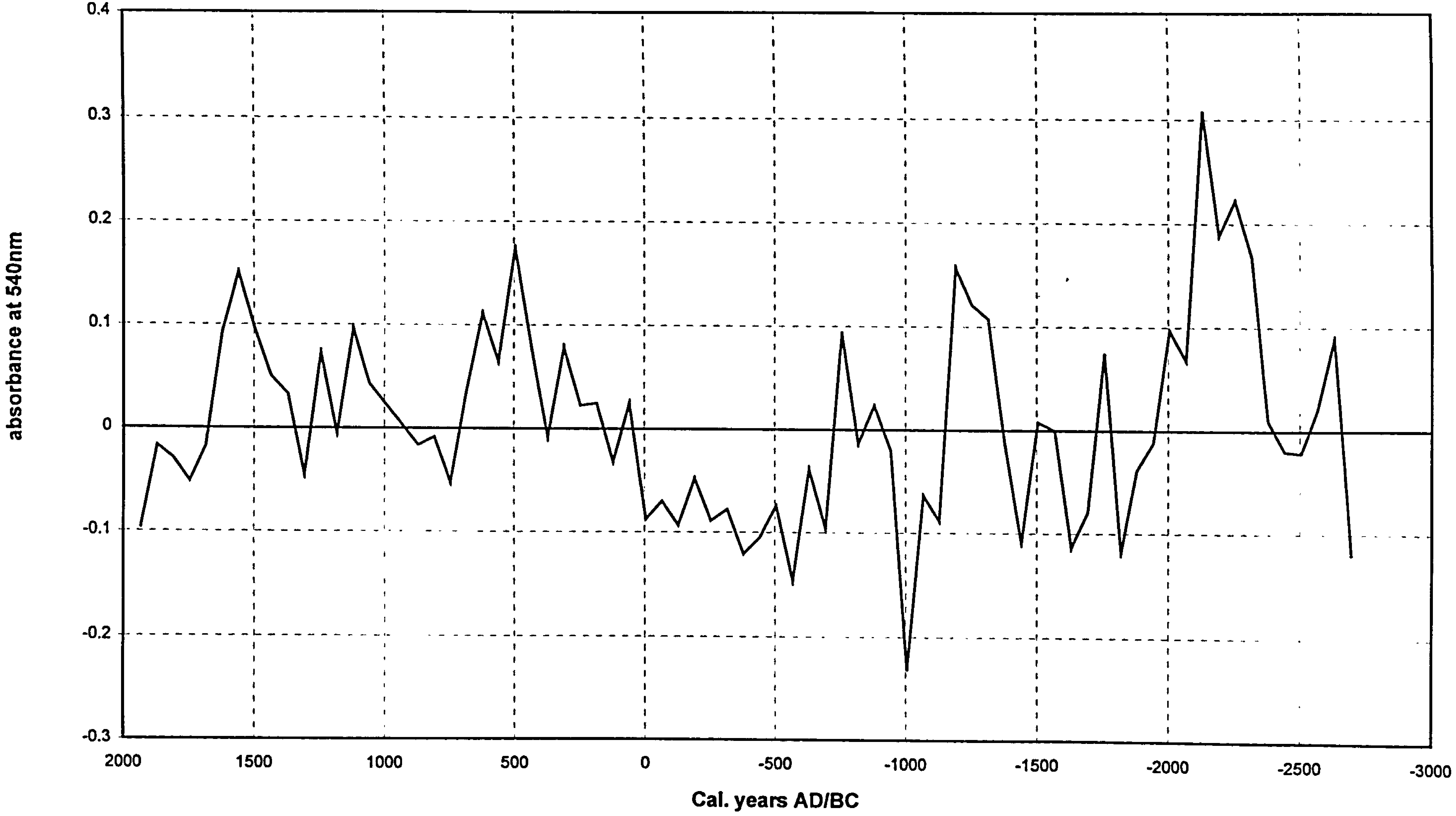


Figure 6.7 Longbridge Moss – Testate amoebae

Analysis: Pete Langdon

Figure 6.8 Longbridge Moss - Plant macrofossils: detrending by segments.

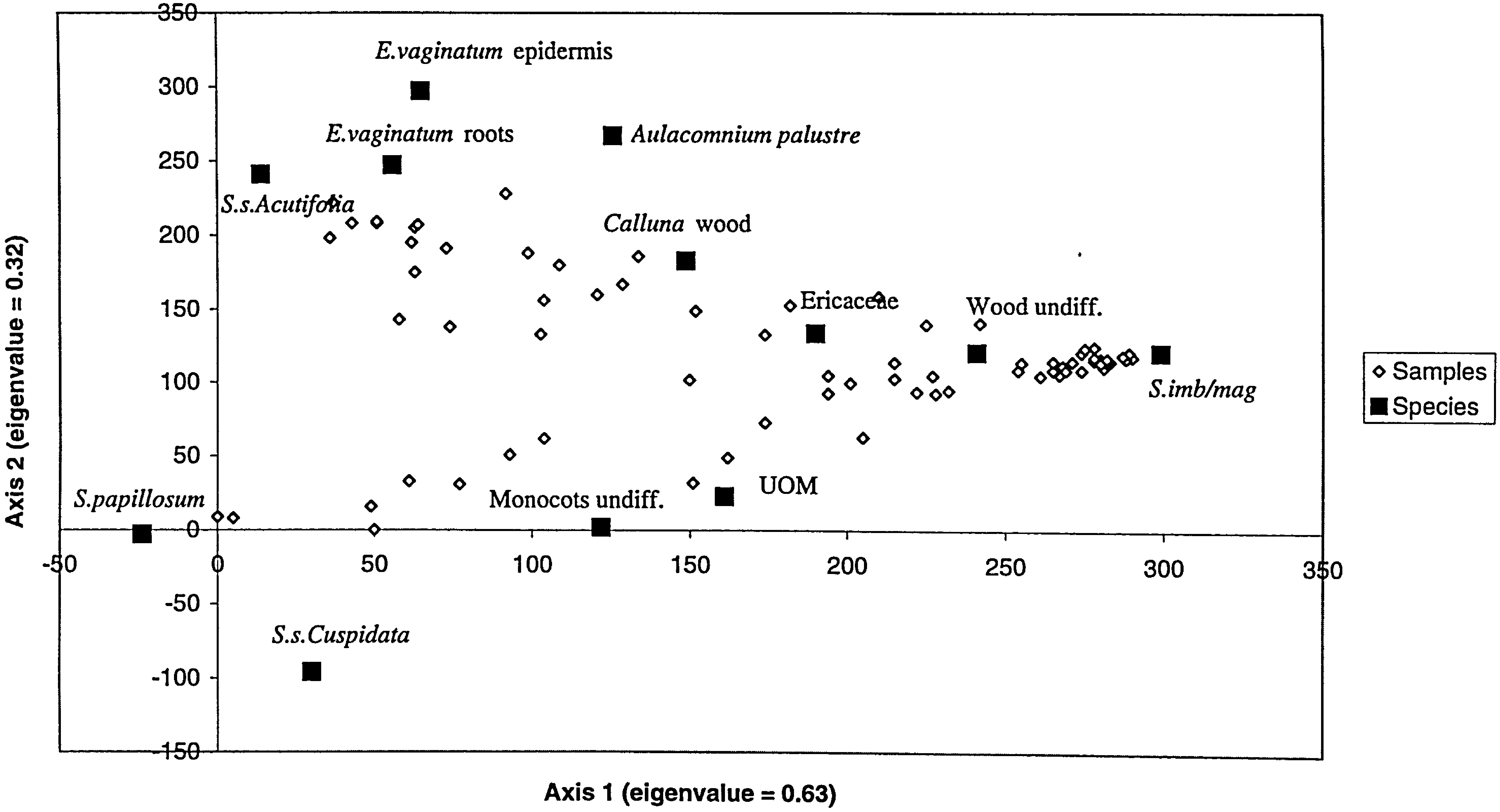


Figure 6.9 Longbridge Moss - Plant macrofossils DCA: axis 1 scores vs time.

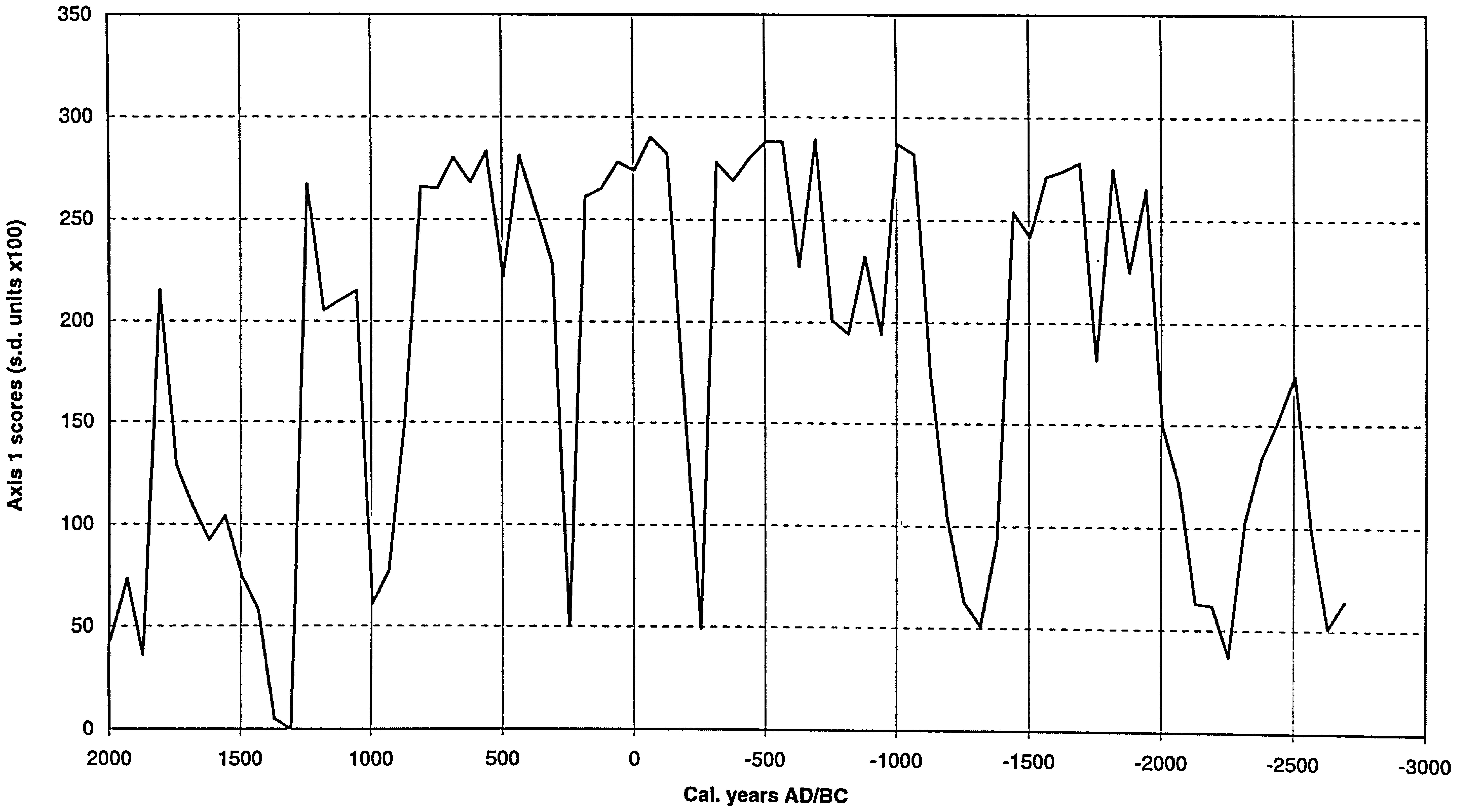


Figure 6.10 Longbridge Moss - Testate amoebae reconstructed water table depths +/- sample prediction errors vs time.

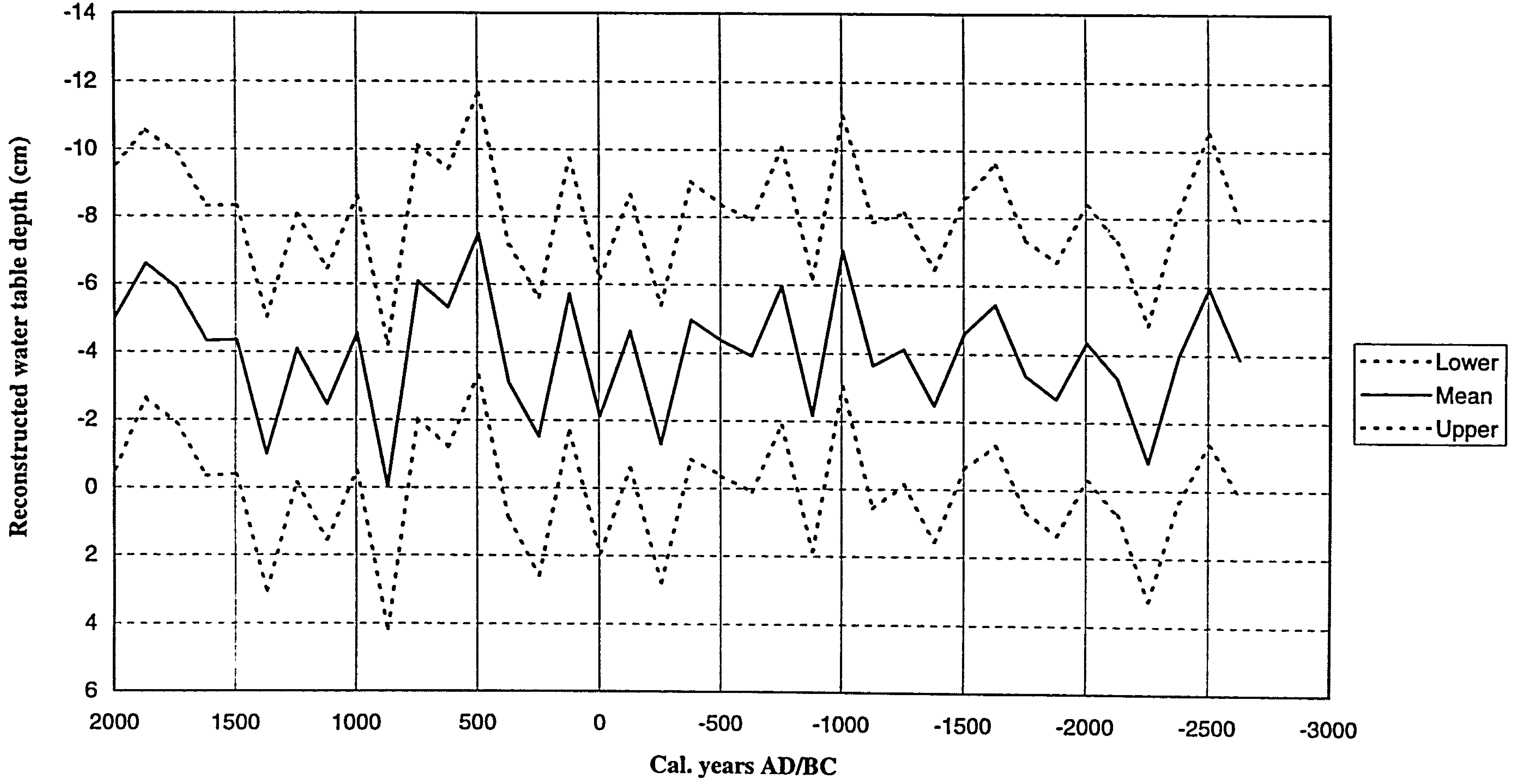
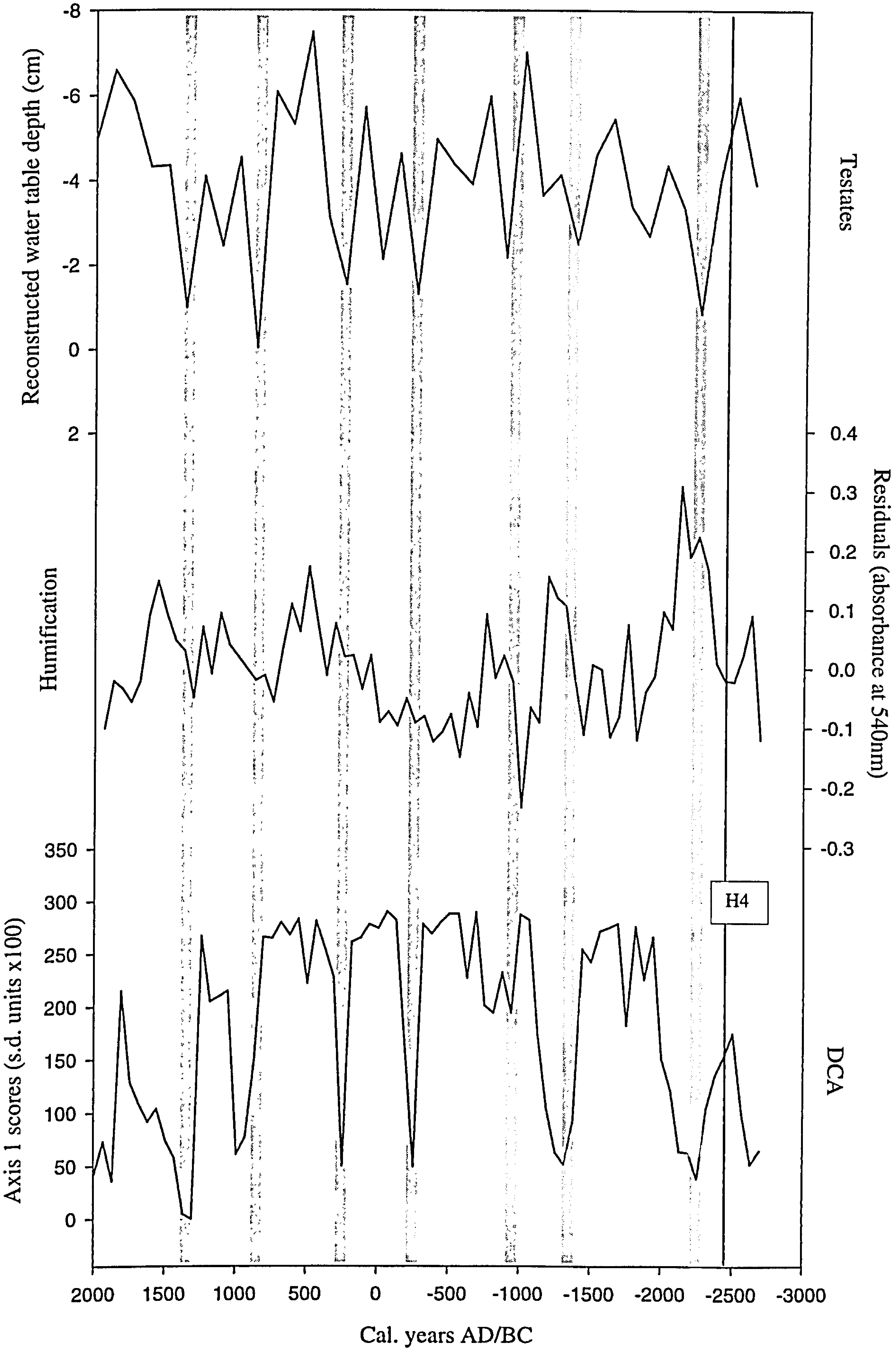


Figure 6.11 Longbridge Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae data plotted against time. H4 indicates the Hekla-4 tephra.



6.7 Langlands Moss

6.7.1 Stratigraphy

The stratigraphy from Langlands Moss (Figure 6.12) is different to that from Longbridge Moss, as it is difficult to recognise clear pool layers within the profiles. The majority of monocots and Ericaceae only occur as small fractions of a *Sphagnum* dominated stratigraphy. The dominance of *Sphagnum imbricatum* within the stratigraphy (data are presented in Appendix 1) may be responsible for this, potentially masking climatic deteriorations as it can adopt a compact or lax form according to height above the water table (Tallis, 1961; Green, 1968). This highlights the problems of using field stratigraphy alone to compare profiles from the same bog. Another profile from Langlands Moss has been analysed by A.Gent (in prep) at contiguous centimetre intervals to two metres depth. A.Gent's core was taken *ca.* 50 metres from the core used in this project, and the changing assemblage of species downcore is remarkably similar between the two profiles.

6.7.2 Plant macrofossils

The plant macrofossil diagram for Langlands Moss is shown in Figure 6.13, and has been divided into six zones on the basis of ecological (and hence climatic) changes.

Zone LAG-a (300-250cm): The flora at 300cm and above is composed of a large abundance of *Sphagnum* section *Acutifolia,* which appears only to be disrupted by a peak of UOM at 280cm, probably the result of a fire, as indicated by relatively large amounts of macroscopic charcoal at this level. The *Sphagnum* section *Acutifolia* then recovers, only to be colonised by an *Eriophorum vaginatum* tussock, indicating a relatively low water table. Boatman and Tomlinson (1977) note how a *Sphagnum capillifolium* var. *capillifolium* hummock can rise as much as 12cm above surrounding lawns in as little as 4 years, but then slow down as they grow to large hummocks, indicating that it may not take long for the hummock to grow and be colonised by sedges. The presence of brown mosses, such as *Hypnum cupressiforme*, *Pohlia nutans*, and *Polytrichum juniperum/alpestre* type associated with the *Sphagnum* section *Acutifolia* also suggests that the water table was relatively low. The *Eriophorum* tussock is suddenly replaced by a sharp peak of UOM at 256cm, possibly indicating a dry climate as high levels of *Cenococcum* are present.

Zone LAG-b (250-154cm): There is a sharp rise in *Sphagnum* section *Acutifolia* at the start of the zone, followed swiftly by an increase in abundance of *Sphagnum imbricatum*. These two communities are constantly replacing each other throughout this zone, in what appears to be a cyclic pattern. *Sphagnum imbricatum* increases dramatically in abundance around 244cm, dated to cal. 1670-1420 BC (SRR-6335). It is suggested that the fluctuations between *Sphagnum imbricatum* and *Sphagnum* section *Acutifolia* represent changes in water table depths, with *Sphagnum* section *Acutifolia* able to replace *Sphagnum imbricatum* in response to a fall in the water table, possibly due to the slow growth rate of *Sphagnum imbricatum* (Flatberg, 1986). A peak of UOM (>50%) is recorded at 224cm, associated with macroscopic charcoal and high levels of *Cenococcum*, thus indicating a fire had occurred. The charcoal and highly oxidised burnt peat may have formed an impermeable layer by blocking pore space and reducing soil porosity (Mallik *et al.*, 1984), allowing water to pond on the mire surface, encouraging the *Sphagnum imbricatum* to colonise the mire once again.

Two small expansions of *Eriophorum vaginatum* are recorded around 188cm and 176-172cm, although both times *Sphagnum imbricatum* appears to overgrow the *Eriophorum*, possibly due to raised water tables, although further evidence from another proxy is needed to confirm this.

Zone LAG-c (154-98cm): This zone is comprised almost totally of *Sphagnum imbricatum* as *Sphagnum* section *Acutifolia* is wholly replaced, apart from a peak in Ericaceae (43%) at 140cm. Very little UOM is present, and this may be indicative of a wetter climate.

Zone LAG-d (98-70cm): A decline in abundance of *Sphagnum imbricatum* is associated with an immediate rise in *Eriophorum vaginatum* (>20%). The *Eriophorum* is then replaced by peaks in Ericaceae and UOM, whilst *Sphagnum* drops to values near zero, apart from a brief resurgence of *Sphagnum imbricatum* at 80cm. However, the

Sphagnum at this level is very broken down, suggesting it spent a long period of time in the aerated acrotelm, and hence probably grew on a hummock. High levels of macroscopic charcoal exist within this zone, suggesting that fires were commonplace, hence the high levels of UOM. The Glen Garry tephra seems to have fallen onto a mire surface rich in Ericaceae and *Eriophorum vaginatum*, with very little *Sphagnum*, almost certainly being a hummock microform, indicating a dry climate.

Zone LAG-e (70-18cm): An increase in *Sphagnum* abundance at the start of this zone, firstly by *Sphagnum* section *Acutifolia*, dated at cal. AD 140-390 (SRR-6333), followed by *Sphagnum imbricatum,* which peaked at cal. AD 440-640 (SRR-6332), suggests a possible return to wetter conditions with a relatively high water table. However, large amounts of UOM (65%) between 52-48cm, in the absence of any charcoal, suggest that high levels of decay were taking place in the acrotelm, and may indicate a temporary lowering of the water table. *Sphagnum imbricatum* levels subsequently increase again until the ultimate demise at between 20-16cm. There is another peak of UOM at 24cm (>50%), again suggesting that decay levels within the mire were high. This would therefore imply that the last phase of *Sphagnum imbricatum* growth was of the compact hummock form, before its extinction, dated at cal. AD 890-1030 (SRR-6331). The 'AD 860' tephra layer was deposited on *Sphagnum imbricatum* between 30-31cm.

Zone LAG-f (20-0cm): The decline in *Sphagnum imbricatum* is associated with an increase in the hygrophilous *Sphagnum* section *Cuspidata,* and subsequent increase in *Sphagnum papillosum.* This increase in hygrophilous taxa associated with the decline in *Sphagnum imbricatum* has been discussed extensively by Stoneman *et al.* (1993) and Mauquoy and Barber (1999b). The date of the local *Sphagnum imbricatum* extinction around 20cm, coupled with the 'AD 860' tephra at 30cm suggest that a major hiatus exists in the top of the stratigraphy. This problem has been addressed in the chronology section 5.3.2. The date of the local extinction of *Sphagnum imbricatum* is in general agreement with dates from other sites (Stoneman et al., 1993; Mauquoy and Barber, 1999b) and therefore this part of the stratigraphy is accepted as being intact. It is suggested that the hiatus exists in the very top of the stratigraphy, indicated by the comparatively large amounts of UOM present between 0-15cm.

6.7.3 Humification

A slight trend can be detected within the raw humification data, so it has been detrended using the same methods as for Longbridge Moss (Figure 6.14). By using the time series of interpolated dates created from the age/depth model (Section 5.3.2), the detrended humification data can be plotted against time (Figure 6.15). Clear phases of lower absorbance, and suggested wetter climates can be seen around the interpolated dates of *ca.* cal. 1925 BC, 1550 BC, 1250 BC, 930 BC, 740 BC, and less markedly around *ca.* cal. AD 240, AD 600, and *ca.* cal. AD 1000. Phases of higher absorbance in relation to a higher humic content within the peat suggests times of lowered water tables and drier climates. The potentially drier times can be identified from the interpolated dates around *ca.* cal. 1760-1650 BC, 1040 BC, 700 BC, 310-110 BC, AD 110, AD 450-520, AD 690-780, and *ca.* cal. AD 1150.

6.7.4 Testate amoebae

The testate amoebae data have been plotted in Figure 6.16 using the same zones as the macrofossil data, allowing a direct comparison between the two techniques. The stratigraphy is dominated by the more xerophilous taxa, *Difflugia pulex* and *Hyalosphenia subflava*, although phases of higher abundances of more hygrophilous taxa, specifically *Amphitrema flavum* and *Arcella discoides* type, can be identified. It is also noted that although Hendon (1998) ultimately suggests the xeric nature of *Difflugia pulex* due to its associations with other xerophilous species, its distribution at Langlands Moss indicates high abundances when the xerophilous species (*H. subflava* notably) are low or absent from the stratigraphy. This may reflect the cosmopolitan distribution of *D. pulex*, also noted by Hendon (1998), and indicate that this species may not necessarily display xeric tendencies at every site.

Zone LAG-a (300-250cm): At 288cm there are high values of the hydrophilous *Arcella discoides* type, as well as *Amphitrema flavum* (>30% for each taxa) suggesting relatively high water tables within the mire. However, these taxa are succeeded by very high abundances (>80%) of the xeric testate *Hyalosphenia subflava* throughout the rest of the zone (the taxa become slightly more diverse only at 272cm) indicating lowered water tables, and hence low levels of effective precipitation. The testate evidence suggests that the *Sphagnum* section *Acutifolia* which dominates the lower section of LAG-a corresponds with relatively higher water tables, before the possible fires and subsequent *Eriophorum vaginatum* dominance indicating drier climates. Testate concentrations are low at the start of the zone, but rise to peak concentrations around 256cm of *ca.* 39000 individuals per cm^3.

Zone LAG-b (250-154cm): Levels of *Hyalosphenia subflava* are declining at the start of the zone, being replaced initially by *Assulina muscorum*, *Difflugia pulex*, and the more hygrophilous *Cyclopyxis arcelloides* type. *Amphitrema flavum* dominates this zone, alternating with fluctuations in abundance of *Difflugia pulex.* The two initial peaks of *Amphitrema flavum*, at 232cm and 214cm respectively, coincide with the occurrence of small quantities of the hydrophiles *Hyalosphenia elegans* and *Hyalosphenia papilio.* The low values of these taxa suggest they are responding to ephemeral events such as precipitation quantity and quality (McMullen, 1999), as Warner and Chmielewski (1992) have shown that the response of testate amoebae to hydrological change can be very rapid. These peaks of *Amphitrema flavum* also coincide with the fluctuating *Sphagnum imbricatum* maxima shown in Figure 6.13, suggesting that the cyclic alternation between *Sphagnum imbricatum* and *Sphagnum* section *Acutifolia* reflects changing hydrological conditions within the bog.

The hydrophilous *Amphitrema wrightianum* emerges between 200-192cm in association with high levels of *Amphitrema flavum*, indicating a higher water table. *Difflugia pulex* and *Assulina muscorum* then replace the other taxa, with small amounts of *Hyalosphenia subflava* between 176-168cm, indicating a return to drier conditions. The water table appears to fluctuate again at the end of the zone, as *Amphitrema* spp. return.

Zone LAG-c (154-98cm): *Arcella discoides* type is present at 144cm, along with *Amphitrema flavum*, and smaller levels of *Hyalosphenia subflava.* Woodland (1996) suggests that the small size of *Arcella discoides* type allows it to exist within a wide range of water table values due to its ability to inhabit shallow water films, and thus the presence of *Hyalosphenia subflava* (18%) and *Difflugia pulex* (12%) at 144cm make water table depth inferences difficult. High levels of *Sphagnum imbricatum* at 144cm (Figure 6.13) do not aid interpretations, due to its bimodal water level distribution.

A rise in abundance of *Difflugia pulex* at 136cm may indicate a lower water table as it corresponds with comparatively high levels of Ericaceae (Figure 6.13), although from this depth upwards in this zone *Difflugia pulex* is gradually replaced by *Arcella discoides* type, suggesting a return to higher water tables, with increased effective precipitation. *Arcella discoides* type is replaced at the end of the zone by *Hyalosphenia subflava*, which goes on to dominate the following zone.

Zone LAG-d (98-70cm): The zone is dominated by two species, *Hyalosphenia subflava* at the start of the zone, with *Difflugia pulex* replacing it at the end of the zone. The high levels of UOM and Ericaceae in the macrofossil diagram also suggest this zone to have comparatively low water tables throughout, so it is therefore possible that the changing species indicate changes within the vegetation (*cf.* McMullen, 1999). It would appear that large levels of UOM are associated with peaks of *Hyalosphenia subflava*, whereas large peaks in Ericaceae correspond with high levels of *Difflugia pulex*. These associations are not entirely consistent throughout the diagram, but differential rates of preservation, as well as the lower sampling resolution of the testacea, may influence the nature of these relationships.

Zone LAG-e (70-18cm): *Difflugia pulex* dominates the zone, with abundances consistently between 40-80%. There is a peak of *Hyalosphenia subflava* at 48cm, which again correlates with a large peak in UOM, suggesting high rates of decay and a low mire water table. There were not enough testates for a count to be taken at 24cm, possibly indicating high levels of decay and subsequent losses of subfossil thecae. This may suggest a hiatus around 24cm, a level which has high amounts of UOM.

Zone LAG-f (18-0cm): There is no evidence for hygrophilous taxa at 16cm, as these appear to often be associated with the extinction of *Sphagnum imbricatum* (Mauquoy and Barber, 1999b), but instead *Difflugia pristis* type dominates, with noteworthy levels of *Assulina muscorum*, suggesting a more intermediate or low water table. These species assemblages may also be the result of a hiatus at the top of the stratigraphy, as discussed in section 5.3.2. The top of the stratigraphy exhibits the most diverse range of species, as is common with testate amoebae reconstructions from peat stratigraphy (Hendon, 1998).

6.7.5 DCA

Detrending the plant macrofossil data by segments has produced a biplot (Figure 6.17) which shows a clear moisture gradient, with *Sphagnum papillosum* and *Sphagnum* section *Cuspidata* occupying one end of axis 1, whereas the more xerophilous *Scirpus cespitosus*, and generally dry indicators *Eriophorum vaginatum* and UOM occupy the other end of axis 1. As *Sphagnum imbricatum* dominates the stratigraphy (Figure 6.13) most of the sample scores tend to cluster around it, and in a similar fashion to Longbridge Moss, the DCA has placed *Sphagnum imb/mag* at one end of axis 1, this time at the wetter end. However, the positioning of *Sphagnum* section *Acutifolia* in a more central location along axis 1 should ensure that not too much distortion occurs when the scores from axis 1 of the ordination are plotted against the interpolated dates from the age/depth model (Figure 6.18).

The DCA results in Figure 6.18 show well defined fluctuations in the wetness reconstruction. Clearly identifiable wet phases can be seen around *ca.* cal. 1500 BC, 1400 BC, 1240 BC, 1070 BC, 970-740 BC, 660-260 BC, 0 AD/BC, AD 300-370, AD 690-1010, and around *ca.* cal. AD 1330. Comparatively drier phases are reconstructed around *ca.* cal. 1980-1670 BC, 1440 BC, 1140 BC, 1040-1000 BC, 110-60 BC, AD 60, AD 520, and *ca.* cal. AD 1150.

6.7.6 Testate amoebae transfer function

There is a greater degree of variation in the testate amoebae reconstructed water table depths from Langlands Moss (Figure 6.19) than from Longbridge Moss (Figure 6.10). These hydrological changes represent the changing dominance of testacea throughout the profile, mainly from the xerophilous *Hyalosphenia subflava* and *Difflugia pulex* to the more hygrophilous *Amphitrema* species. Wet phases, with a relatively low water table occur around *ca.* cal. 1900 BC, 1370-1110 BC, 500-400 BC, AD 110-240, and *ca.* cal. AD 690-1330. Comparatively drier phases can be identified around *ca.* cal. 1820 BC, 1760-1650 BC, 310 BC - AD 50, and *ca.* cal. AD 440-520.

6.7.7 Comparison of the three proxy techniques

The three proxy techniques are plotted together in Figure 6.20, and the inferred climatic deteriorations (represented as greyscale lines) are shown to be coherent between each reconstruction. Clear phases of climatic deteriorations are recorded (Figure 6.20) around *ca.* cal. 1900 BC, 1560 BC, 1240 BC, 930 BC, 500 BC, AD 300, AD 600, and *ca.* cal. AD 900. The reconstructions cannot be defined clearly after *ca.* cal. AD 1300 due to the limitations of the age/depth model (see Section 5.3.2). Dry climatic shifts are difficult to tie down to specific dates, as the onset of a dry phase may result in peat erosion or a cessation of growth. However, phases when the climate was relatively dry and where each reconstruction is in agreement can be discerned between the wet phases, and are identified at around *ca.* cal. 1700 BC, 1040 BC, 110 BC, and *ca.* cal. AD 520.

Figure 6.12 Langlands Moss: Stratigraphy

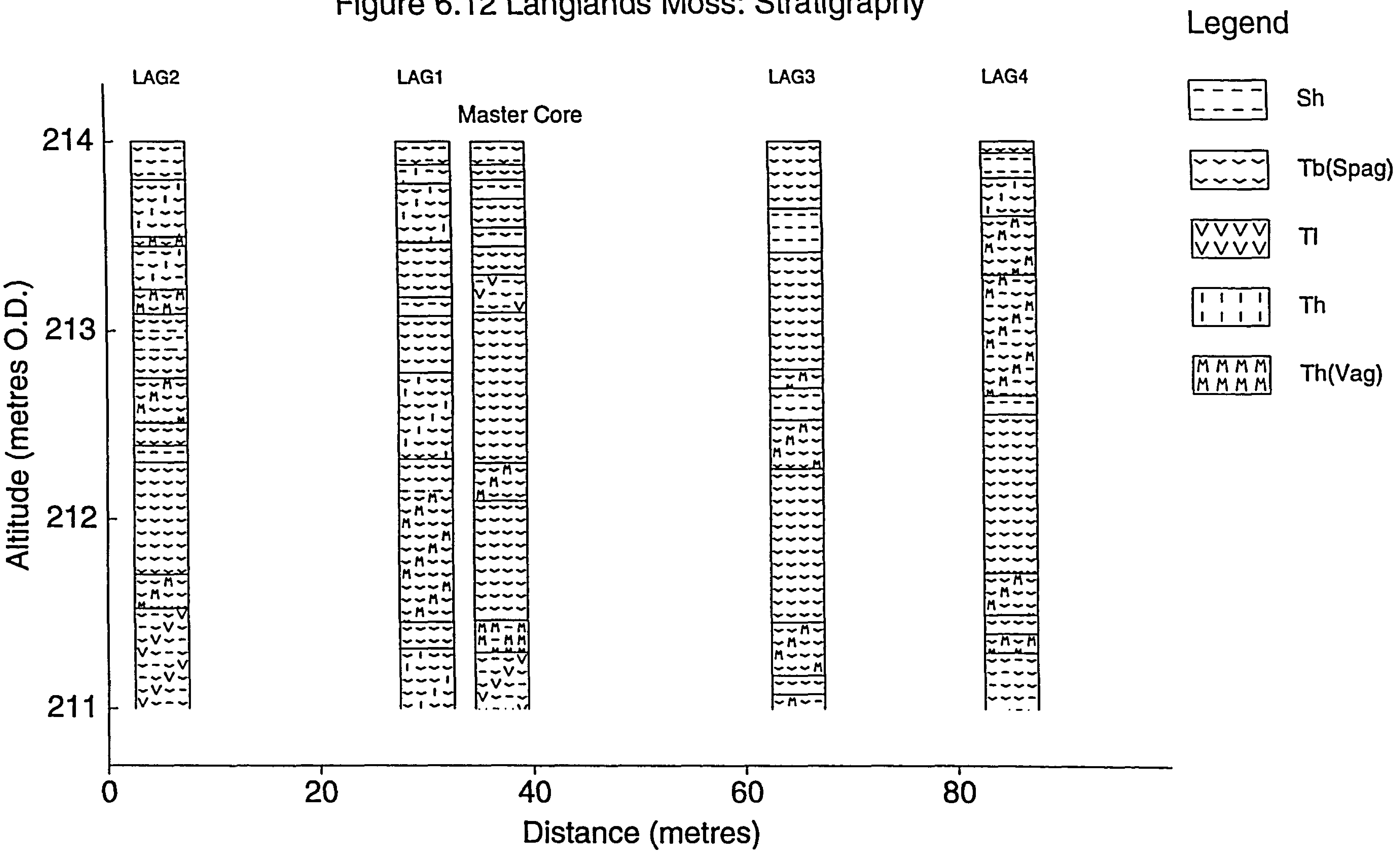


Figure 6.13 Langlands Moss – Macrofossil Diagram

Analysis: Pete Langdon

Figure 6.14 Langlands Moss - Humification data. The raw data are presented at the top, and have been detrended by linear regression, producing the lower data set.

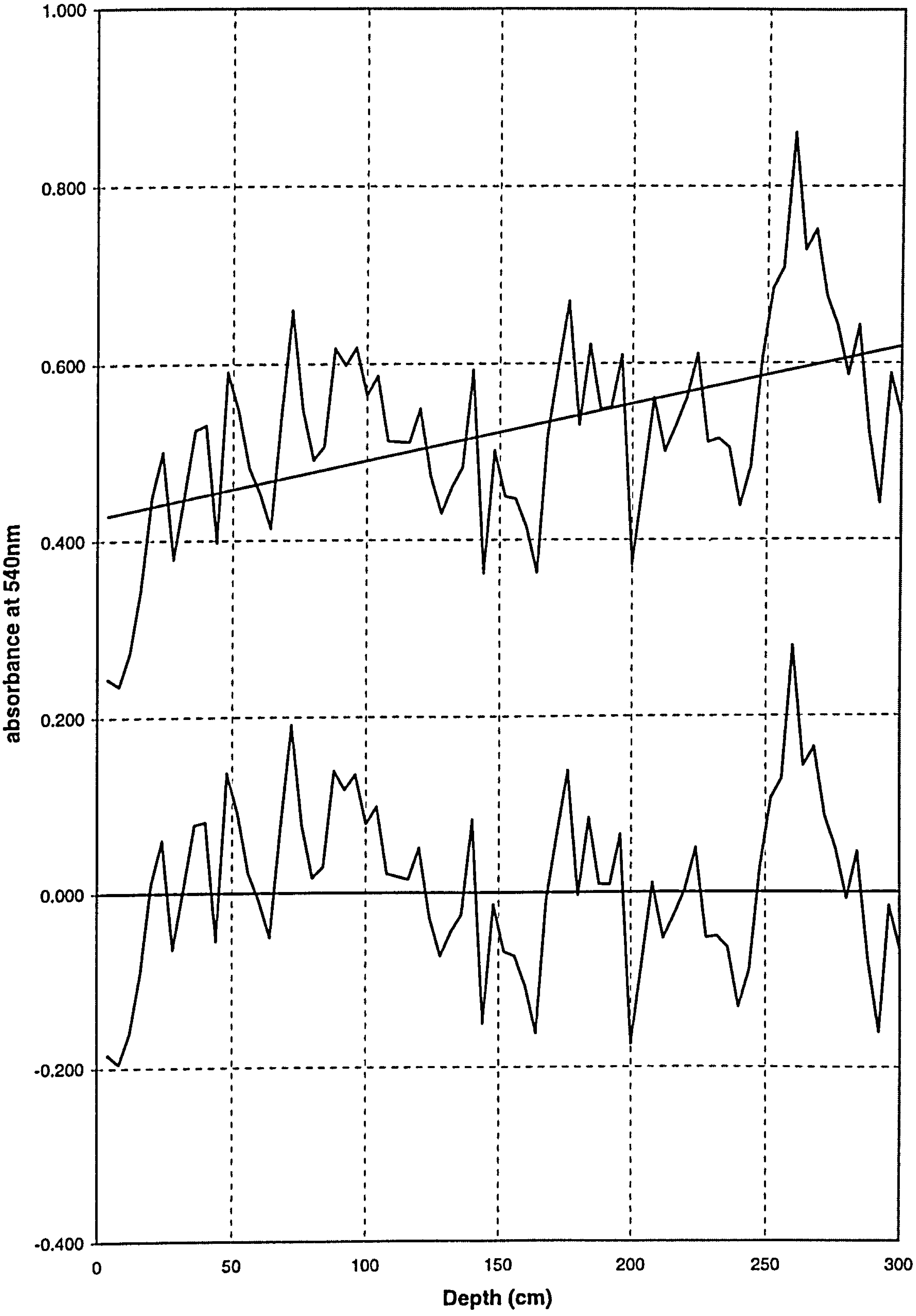


Figure 6.15 Langlands Moss - Detrended humification vs time.

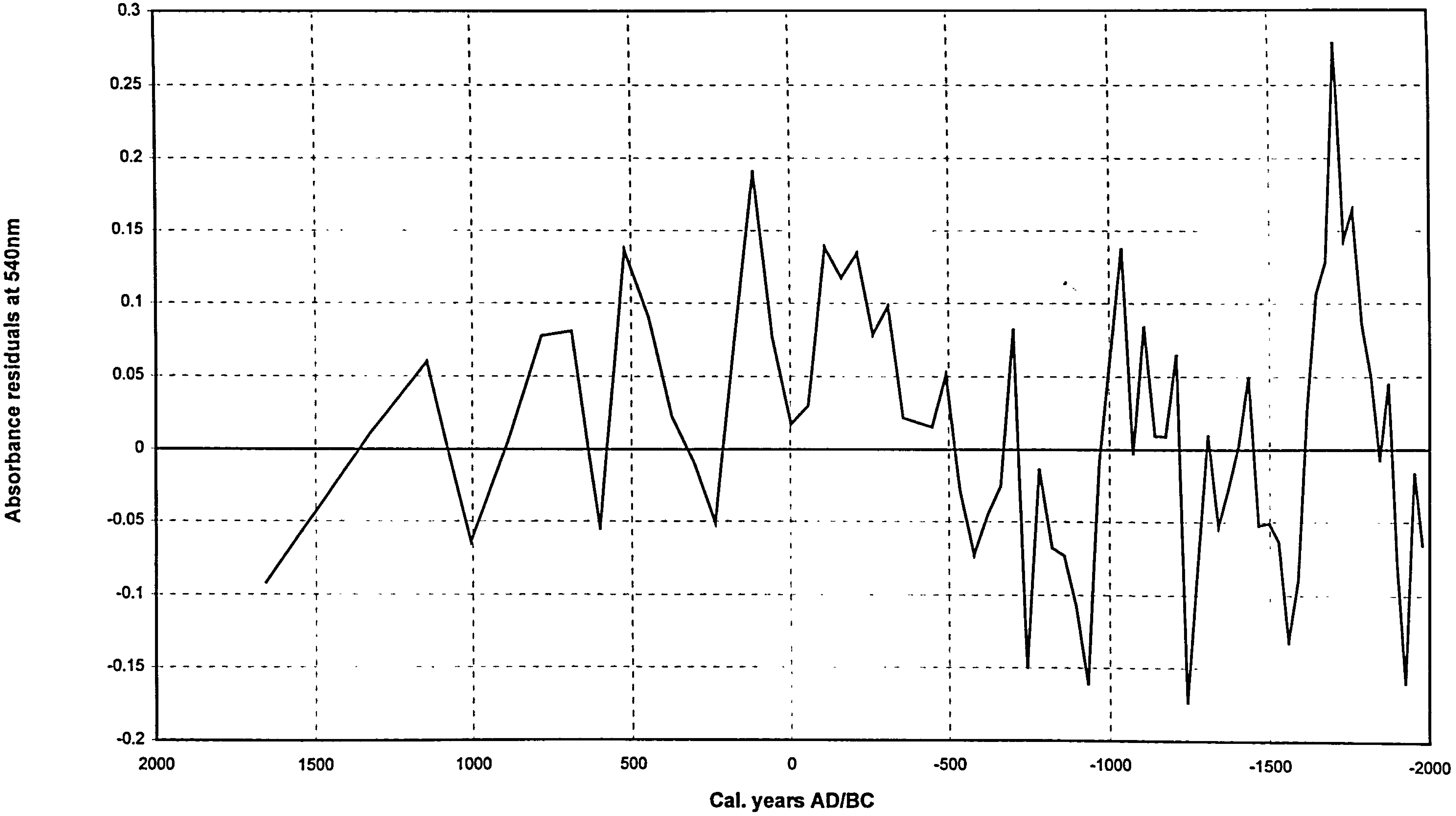


Figure 6.16 Langlands Moss – Testate amoebae

Analysis: Pete Langdon

Figure 6.17 Langlands Moss - Plant macrofossils: detrending by segments.

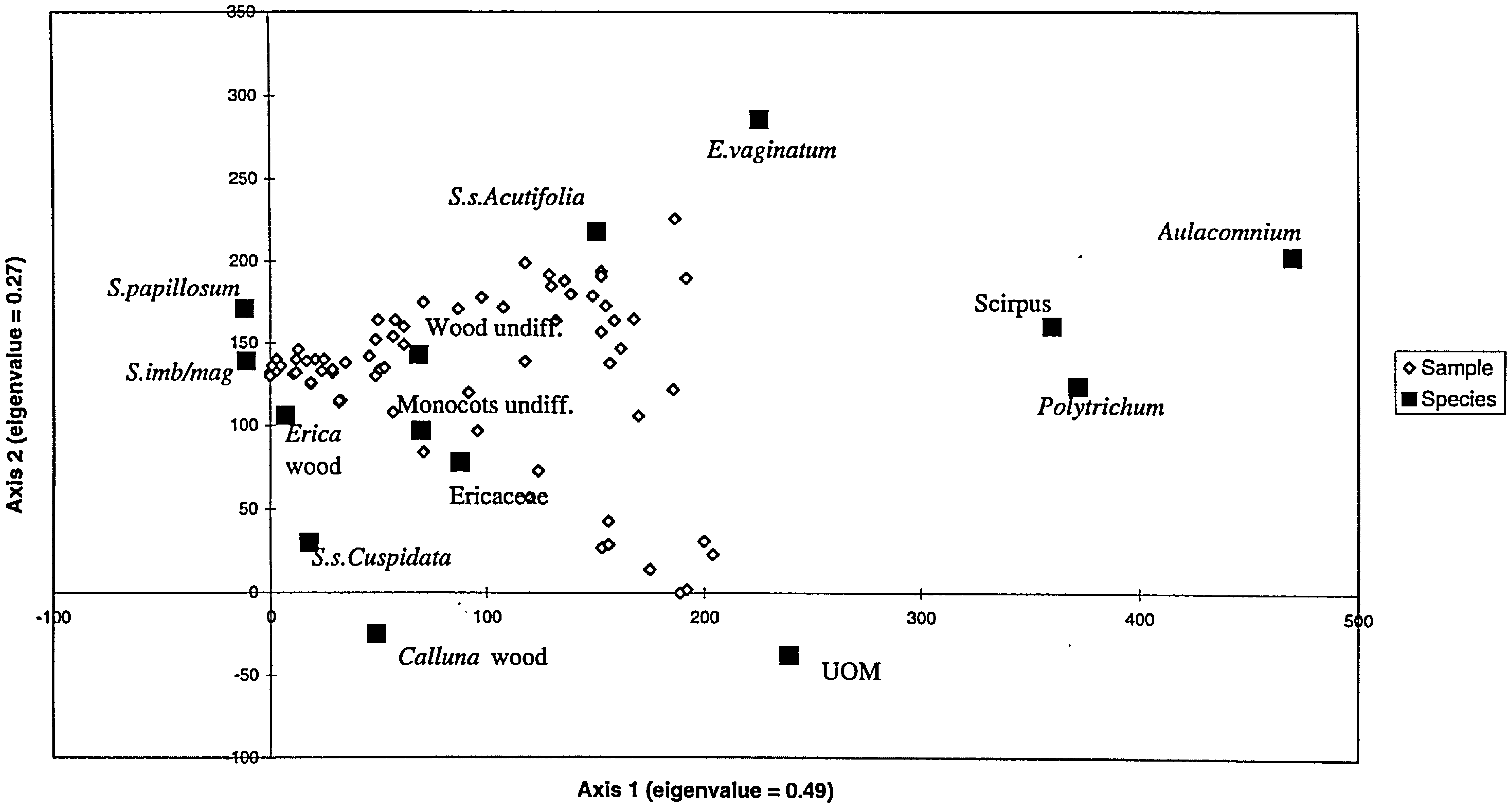


Figure 6.18 Langlands Moss - Plant macrofossils DCA: axis 1 scores vs time.

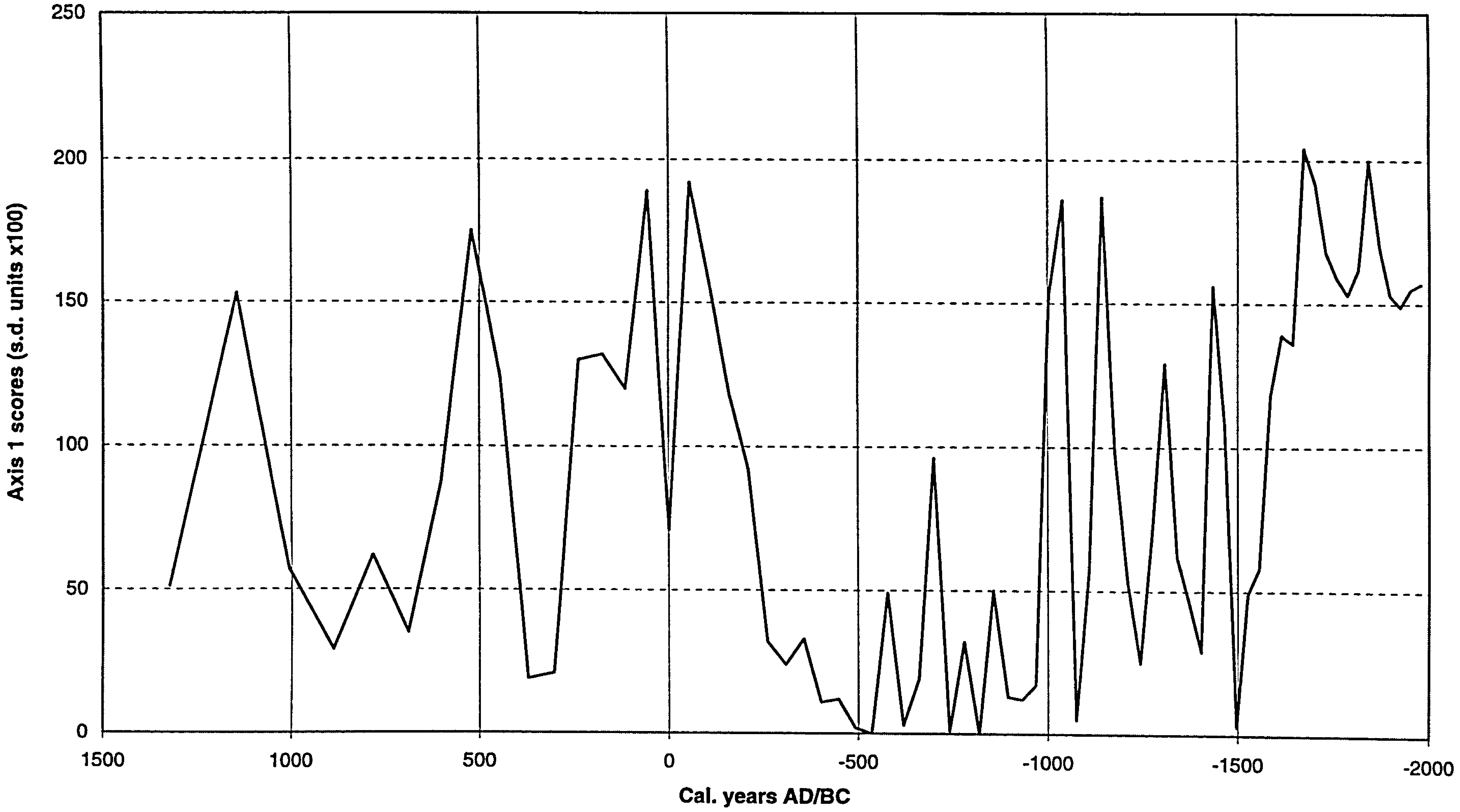


Figure 6.19 Langlands Moss - Testate amoebae reconstructed water table depths +/- sample prediction errors vs time.

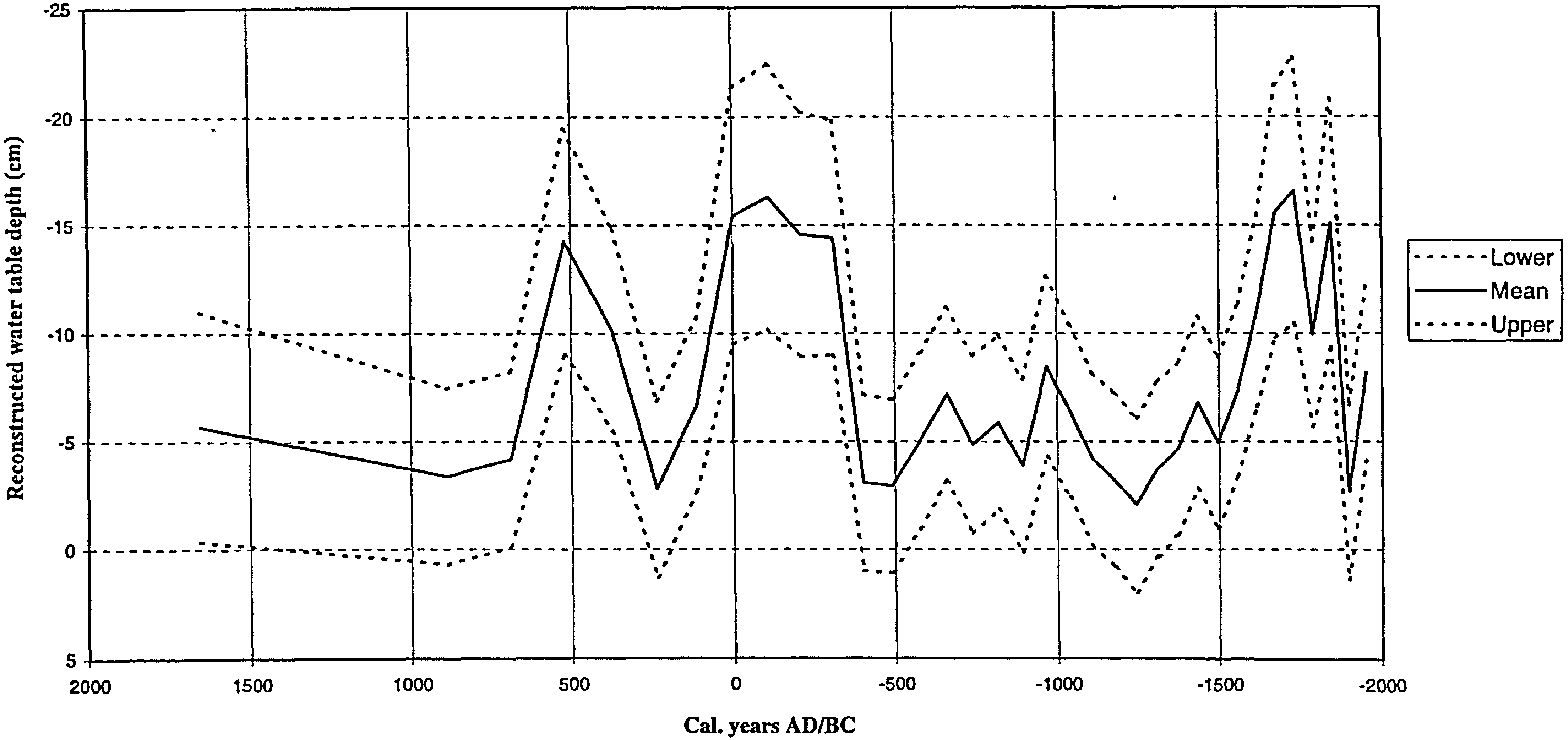
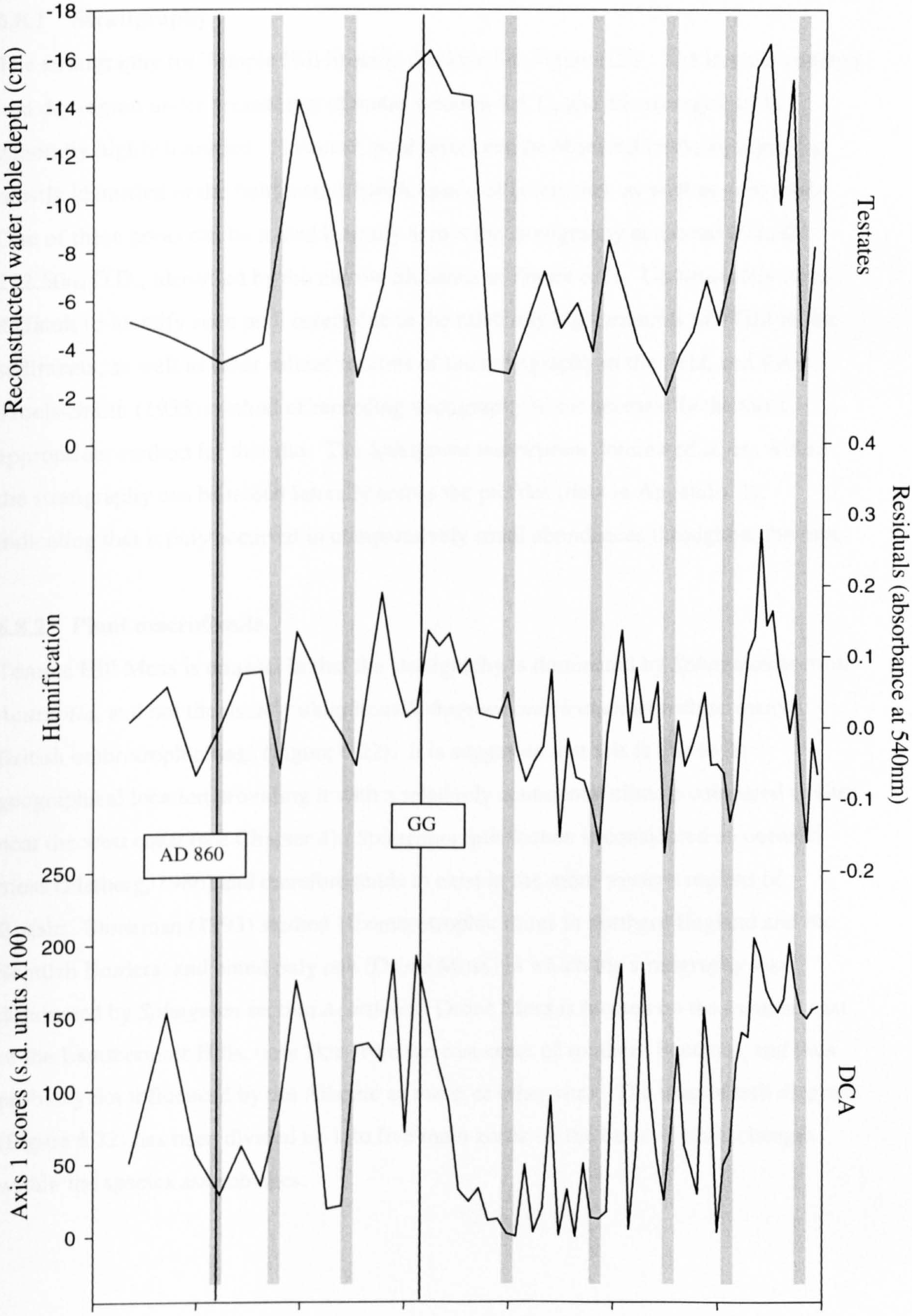


Figure 6.20 Langlands Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae plotted against time. Solid vertical lines indicate tephras (GG=Glen Garry)



6.8 Temple Hill Moss

6.8.1 Stratigraphy

The stratigraphy for Temple Hill Moss is displayed in Figure 6.21. The moss is situated in a dry region under present day climates (section 4.4.1), and the stratigraphy is generally highly humified. However, pool layers can be observed in the stratigraphy, mostly identified in the field from *Rhynchospora alba* remains, as well as pool muds. One of these pools can be traced laterally across the stratigraphy at around 292.20-292.50m O.D., identified by the narrow Sh bands in Figure 6.21. Unfortunately it was difficult to identify such pool layers due to the relatively high amounts of UOM in the sediments, as well as other salient features of the stratigraphy in the field, and the Troels-Smith (1955) method of recording stratigraphy is not necessarily the most appropriate method for this site. The *Sphagnum imbricatum* dominated layers within the stratigraphy can be traced laterally across the profiles (data in Appendix 1), indicating that it only occurred in comparatively small abundances throughout the mire.

6.8.2 Plant macrofossils

Temple Hill Moss is unusual in that the stratigraphy is dominated by *Sphagnum* section *Acutifolia*, and not the usually ubiquitous *Sphagnum imbricatum* as with so many British ombrotrophic bogs (Figure 6.22). It is suggested that this is due to its geographical location providing it with a relatively continental climate compared to sites near the west coast (see Chapter 4). *Sphagnum imbricatum* is considered an oceanic moss (Flatberg, 1986) and therefore tends to exist in the more western regions of Britain. Stoneman (1993) studied 10 ombrotrophic mires in northern England and the Scottish Borders, and noted only one (Drone Moss) in which the stratigraphy was dominated by *Sphagnum* section *Acutifolia*. Drone Moss is located on the extreme east of the Lammermuir Hills, only 3km from the east coast of southern Scotland, and thus probably not influenced by the Atlantic as much as other sites. The macrofossil diagram (Figure 6.22) has been divided up into five main zones on the basis of main changes within the species assemblages.

Zone TEM-a (484-360cm): A small amount of *Phragmites* is present at 484cm, in association with high amounts of UOM at the bottom of this zone. The UOM declines in abundance to 450cm as *Sphagnum* section *Acutifolia* is able to grow on the mire surface, accompanied by *Polytrichum juniperum/alpestre* type and *Hypnum cupressiforme*, at the time of the Lairg tephra deposition, cal. 5060-4780 BC. The presence of these brown mosses suggests that the *Sphagnum* may have been growing on a hummock, with a relatively low water table.

The Sphagna became overgrown by an *Eriophorum vaginatum* tussock at 428cm, until the *Sphagnum* section *Acutifolia* was able to re-establish itself between 420-404cm. The second phase of *Sphagnum* growth took place with no brown mosses and very little UOM or Ericaceae, suggesting that the *Sphagnum* may have been growing on a low hummock/lawn and the water table may have been relatively higher, inferring an increase in effective precipitation, dated to cal. 4900-4600 BC (SRR-6340). High levels of UOM exist when the *Sphagnum* abundance decreases, and was overgrown again by an *Eriophorum vaginatum* tussock at 396-388cm. The *Eriophorum* was eventually overgrown again by *Sphagnum* section *Acutifolia*, with high levels of UOM between 388-372cm, indicating high levels of decomposition, and hence a low water table.

Zone TEM-b (360-274cm): Relatively high levels of *Sphagnum* section *Acutifolia* exist throughout the whole of the zone, the only major dip being at 308-300cm when an *Eriophorum vaginatum* tussock grew over the *Sphagnum*. The presence of *Dicranum scoparium* at the start of the zone suggests that water levels were relatively low, as this moss tends to be found on hummock tops (Corley, 1992). However, after its decline UOM levels are negligible, and there are no major peaks in Ericaceae, suggesting that material was able to pass through the acrotelm relatively quickly, which implies a comparatively high water table was present. The *Sphagnum* section *Acutifolia* may therefore have grown on a low hummock/lawn microform.

Scirpus cespitosus expands over the *Sphagnum* just before the rise in *Eriophorum vaginatum* at 316cm, possibly indicating a lowered water table, hence the *Eriophorum* was able to colonise the microform.

Zone TEM-c (274-54cm): This zone contains all the *Sphagnum imbricatum* found at the site, and coupled with the field stratigraphy, reveals that the amount of *Sphagnum imbricatum* within the mire is relatively small. The start of this zone consists of a phase of abundant *Sphagnum* section *Acutifolia*, with an associated occurrence of the hydrophilous *Sphagnum* section *Cuspidata*, dated to cal. 2890-2600 BC (SRR-6338). Increased amounts of UOM at 272cm may indicate pool muds. The suggested palaeoenvironment is of a low lawn/pool microform, with the pool expanding and contracting in relation to changes in effective precipitation, as depicted by the levels of *Sphagnum* section *Cuspidata* and *Eriophorum angustifolium* within the zone.

Sphagnum imbricatum is present in small amounts in the early part of the zone, but first flourishes at 224cm, when there is a small presence (<20%) of *Eriophorum vaginatum*, suggesting that possibly the water table had fallen, allowing the expansion of *Sphagnum imbricatum*. High levels of UOM are also associated with *Sphagnum imbricatum* growth, suggesting high levels of decay, and a hummock microform with associated low water table depths. There is a phase of wetter conditions, represented by an abundance of *Sphagnum* section *Cuspidata* between 188-184cm, replacing *Sphagnum imbricatum*, dated at cal. 1430-1220 BC (SRR-6337). The *Sphagnum* section *Acutifolia* growing at this time were probably the more hygrophilous species, *Sphagnum subnitens* or *Sphagnum molle* (*cf.* Stoneman, 1993). Following this wet phase, *Sphagnum imbricatum* dominates the stratigraphy, from 176-152cm, which is between *ca.* cal. 1320-920 BC from the interpolated dates. It is difficult to suggest mire hydrological characteristics from macrofossils alone when *Sphagnum imbricatum* dominates the stratigraphy, necessitating the use of the other proxies (Mauquoy and Barber, 1999a).

Peaks in UOM of *ca.* 30% between 148-140cm, and 108-80cm suggest phases of relatively low water tables, although a decline in UOM corresponding with increased abundance of *Sphagnum* section *Acutifolia* around the Glen Garry tephra suggests the possibility of raised water table depths. As the levels of UOM and *Sphagnum imbricatum* decline towards the end of the zone, they are replaced by *Sphagnum* section *Acutifolia* and a persistently low level of *Sphagnum* section *Cuspidata*, coupled with *Rhynchospora alba* and *Eriophorum angustifolium*, indicating raised water table depths.

This may have been responsible for the extinction of *Sphagnum imbricatum* at Temple Hill (dated at cal. AD 20-230: SRR-6336), especially as it has been relatively unable to dominate the stratigraphy for any considerable period of time.

Zone TEM-d (54-22cm): Exceptionally high levels of UOM dominate, with levels of *Eriophorum vaginatum* also high at around 20%. There is virtually no *Sphagnum* present, although other mosses are found (*Hypnum cupressiforme* and *Racomitrium lanuginosum*), possibly having colonised the peat in a response to fires, as high levels of macroscopic charcoal are found. The interpolated dates for this zone are between *ca.* cal. AD 900-1500, and it could be possible that muir burning was taking place during these times. Irrespective of whether the fires were anthropogenically induced, or due to drier climates, there is a possibility that hiatuses within the stratigraphy exist.

Zone TEM-e (22-0cm): A distinct shift in wetness of the bog is recognised as the Sphagna return with the colonisation of *Sphagnum* section *Cuspidata*. *Drepanocladus fluitans* is also present at 4cm, which has been associated with *Sphagnum cuspidatum* leaves in a peat face at Emmen (Casparie, 1972).

6.8.3 Humification

A very slight trend can be seen in the absorbance values for Temple Hill Moss (Figure 6.23) so the detrending technique used for Longbridge Moss was applied to the data. Using the interpolated dates from the age/depth model (see Section 5.3.3) the results can be plotted against time (Figure 6.24), and phases of low humic content, implying wetter climatic conditions, can be identified at *ca.* cal. 4700 BC, 4000 BC, 3470-3150 BC, 2700 BC, 1510-1450 BC, with smaller fluctuations at *ca.* cal. 590 BC, AD 670, and from *ca.* cal. AD 1500. Phases of greater humic content, and thus potentially drier climates occur at *ca.* cal. 4890 BC, 4400-4300 BC, 2930 BC, 2420 BC, 1890 BC, 1760 BC, 1320 BC, 1060-860 BC, 720-590 BC, 160 BC, AD 130, and *ca.* cal. AD 920-1160.

6.8.4 Testate amoebae

The testate amoebae data can be seen in Figure 6.25. The stratigraphy is dominated by *Amphitrema flavum*, although notable peaks in *Hyalosphenia subflava* indicate much lower water tables.

Zone TEM-a (452-360cm): Testate concentrations were too low to count below 452cm, as well as at 388cm. *Difflugia pulex* and *Assulina muscorum* dominate the assemblages within this zone, along with relatively high abundances of *Nebela militaris* and the xeric *Trigonopyxis arcula*, suggesting that the water table was relatively low throughout this zone.

Zone TEM-b (360-274cm): A change in species assemblage from *Difflugia pulex* to *Amphitrema flavum* at 356cm indicates a shift to wetter conditions. High abundances of *Amphitrema flavum* persist through the zone until they start to be replaced by *Assulina muscorum* at 288cm. Comparatively high values of *Difflugia pulex* at 340cm suggest a possible brief lowering of the water table, although a peak in abundance of *Hyalosphenia papilio* at 304cm implies a comparatively high water table.

Zone TEM-c (274-54cm): High levels of *Amphitrema flavum* continue throughout the zone apart from when they are replaced by large abundances of *Hyalosphenia subflava* between 176-152cm and 112-96cm, indicating lowered water tables for these periods, dated between *ca.* cal. 1320-920 BC and *ca.* cal. 230 BC - AD 60 respectively, using the interpolated dates. The first drier phase of growth matches the peak growth of *Sphagnum imbricatum*, suggesting it was restricted to its compact hummock microform. The second phase coincides with relatively high levels of *Sphagnum imbricatum*, which then fall, presumably as the *Sphagnum imbricatum* tries to adapt to increasingly wetter conditions, which it cannot, and eventually becomes extinct at 64cm.

The wetter conditions in this zone are emphasized by an abundance of *Amphitrema wrightianum*, which generally coincide with peaks in levels of *Amphitrema flavum*. The main phases are between 240-216cm, 144-128cm, and at the end of the zone, associated with the extinction of *Sphagnum imbricatum*, from 72-64cm, and have interpolated dates of *ca.* cal. 2300-1940 BC, 790-520 BC, and *ca.* cal. AD 520-670 respectively.

The Glen Garry tephra occurs between phases of *Amphitrema wrightianum* and *Hyalosphenia subflava* dominance, suggesting intermediate water table depths, although 30% of the testacea are composed of *Difflugia pulex*, suggesting a comparatively dry mire surface.

Zone TEM-d (54-22cm): There were not sufficient testates to undertake a count at 48cm, and the three levels which were counted all produced xerophilous fauna assemblages comprising mainly *Assulina muscorum*, *Euglypha rotunda*, *Hyalosphenia subflava*, and *Nebela militaris*, indicating a low water table depth.

Zone TEM-e (22-0cm): High levels of *Arcella discoides* type exist, indicating a high water table, with a diverse assemblage of species represented in the top few samples.

6.8.5 DCA

The first and second axis of the ordination are plotted against each other (Figure 6.26) and show a hummock-hollow gradient along axis 1, which has an eigenvalue of 0.48, being far greater than on any of the other axes. Over 4 s.d. units are represented on axis 1, suggesting that the species at each end of the axis are mutually exclusive, as in a unimodal species response model species abundances rise and fall over 4 s.d. units along an environmental gradient (ter Braak and Prentice, 1988). The presence of *Sphagnum* section *Acutifolia* towards the centre of axis 1 relates to its bimodal distribution, although the positioning of *Sphagnum imbricatum* at one end of axis 1 relates to its presence under relatively dry conditions at Temple Hill, which is corroborated with evidence from the testate amoebae and humification data.

The axis 1 scores have been plotted against time (Figure 6.27) using the interpolated dates from the age/depth model, and serve as an effective precipitation proxy curve, as the sample scores represent a water level gradient. Notable wet phases occur around *ca.* cal. 4690-4600 BC, 1940 BC, 1450-1380 BC, 450-380 BC, AD 520-750 and *ca.* cal. AD 1690 to present. There is evidence of a general wet phase between *ca.* cal. 4000-2120 BC with subtle fluctuations, although this phase is mainly dominated by *Sphagnum* section *Acutifolia* (Figure 6.22) and hence can only be distinguished as significant or not by using the other proxies. Dry climatic phases can be identified around *ca.* cal. 5450-5170 BC, 4880-4790 BC, 4500-4300 BC, 2000 BC, 1820 BC, 1250-920 BC, AD 60, and *ca.* cal. AD 920-1600.

6.8.6 Testate amoebae transfer function

Comparatively large fluctuations in reconstructed water table depths from the testate amoebae data can be seen in Figure 6.28, relating to the variations between the hygrophilous *Amphitrema* species and peaks in the more xerophilous *Hyalosphenia subflava* and *Assulina muscorum*. Phases of relatively high water table depths, and hence wetter climates can be seen around *ca.* cal. 3900 BC, 3370-3150 BC, 2530-1940 BC, 1450 BC, 720-520 BC, AD 360-750, and *ca.* cal. AD 1690-1780. Comparatively drier climatic phases can be recognised around *ca.* cal. 4500 BC, 2930-2650 BC, 1820 BC, 1320-930 BC, 90 BC - AD 60, and *ca.* cal. AD 1340-1510.

6.8.7 Comparison of the three proxy techniques

The three effective precipitation proxy curves have been plotted in a composite diagram (Figure 6.29) which show good agreement between the reconstructions, and specific phases of wetter and drier climates which can be detected in all three proxies. The greyscale lines on Figure 6.29 relate to climatic deteriorations, and can be recognised in each of the three major proxies around *ca.* cal. 4700 BC, 3900 BC, 3350 BC, 2300 BC, 1900 BC, 1450 BC, 850-500 BC, AD 500-600, and *ca.* cal. AD 1700-1800. These wet phases present a first order impression of cyclicity, which is explored fully in section 6.13. Comparatively drier phases from Temple Hill Moss are recognised around *ca.* cal. 4900 BC, 4400 BC, 2900 BC, 1750 BC, 1250-920 BC, AD 100, and *ca.* cal. AD 920-1160.

Figure 6.21 Temple Hill: Stratigraphy

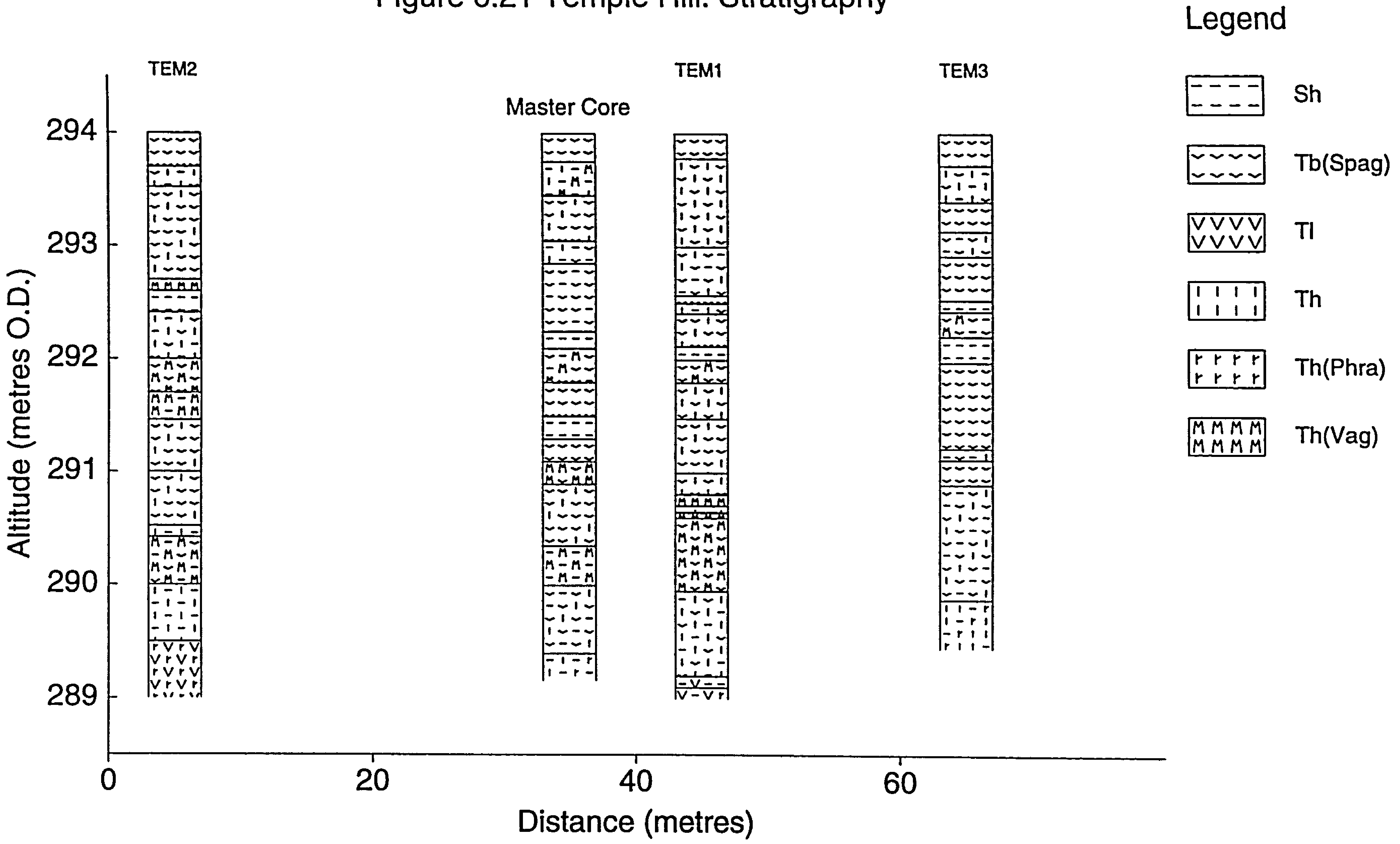


Figure 6.22 Temple Hill Moss – Macrofossil diagram

Analysis: Pete Langdon

Figure 6.23 Temple Hill Moss - Humification data. The raw data are presented at the top, and have been detrended by linear regression, producing the lower data set.

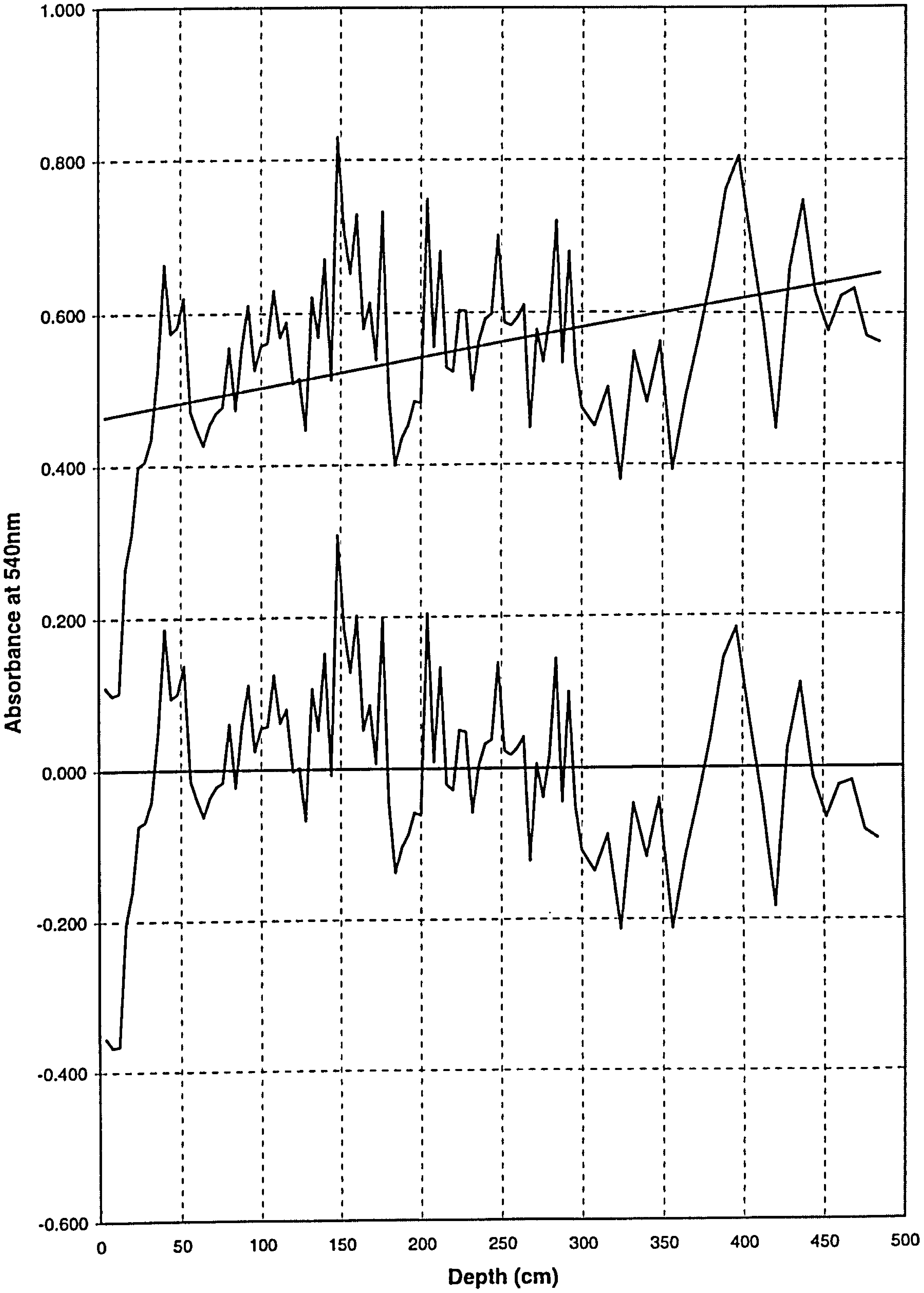


Figure 6.24 Temple Hill Moss - Detrended humification vs time.

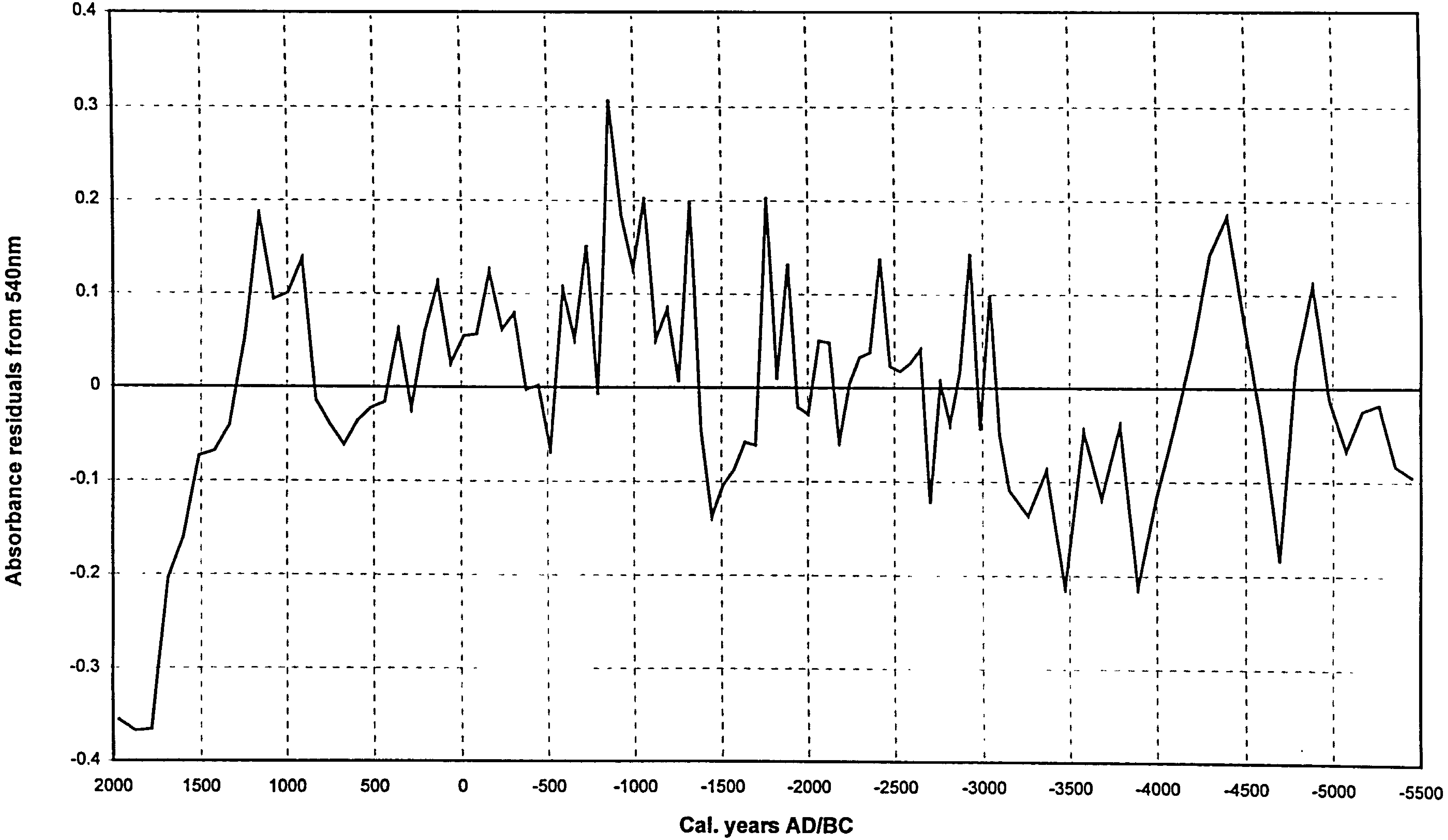
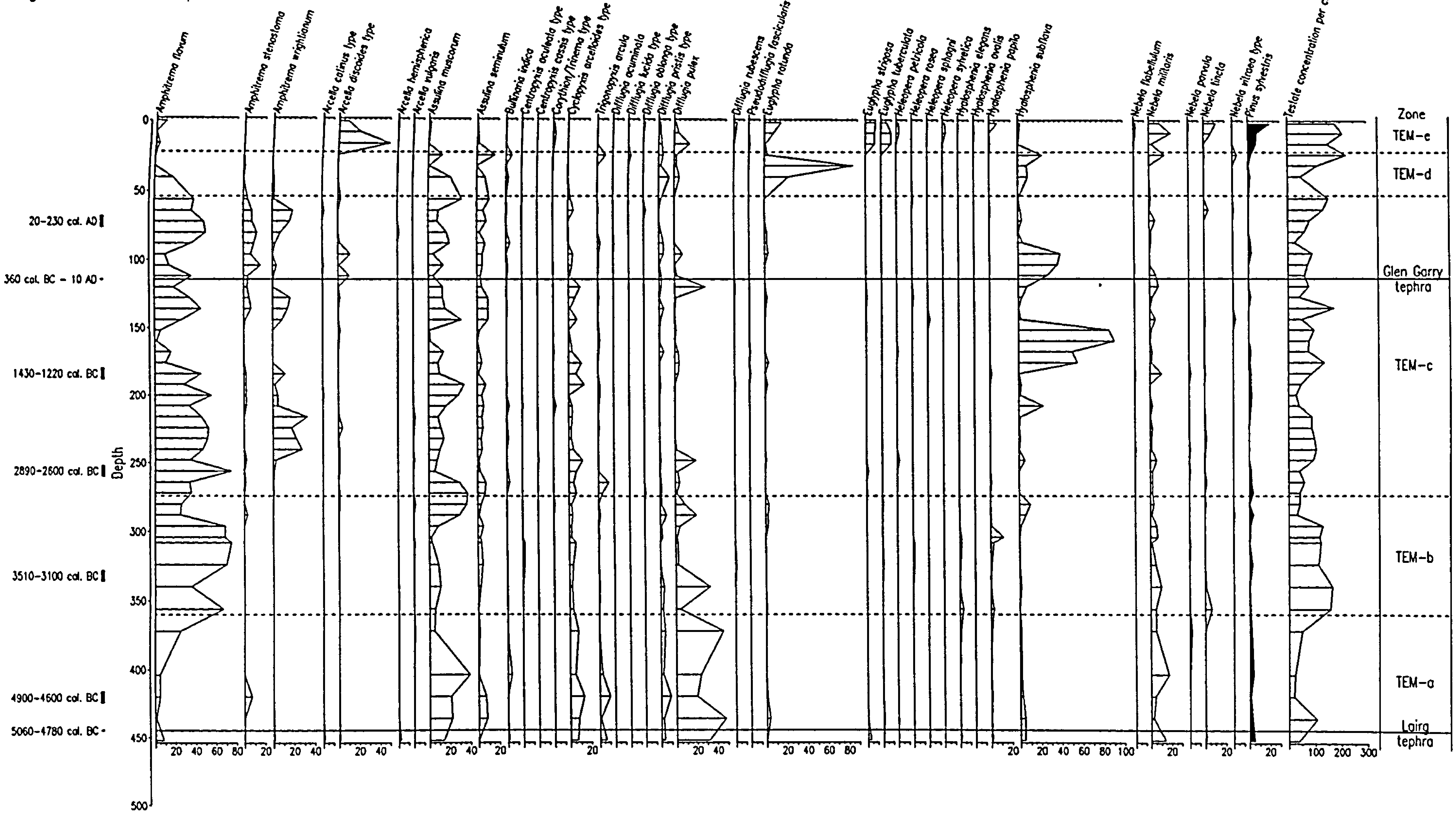


Figure 6.25 Temple Hill Moss – Testate amoebae



Analysis: Pete Langdon

Figure 6.26 Temple Hill Moss: detrending by segments.

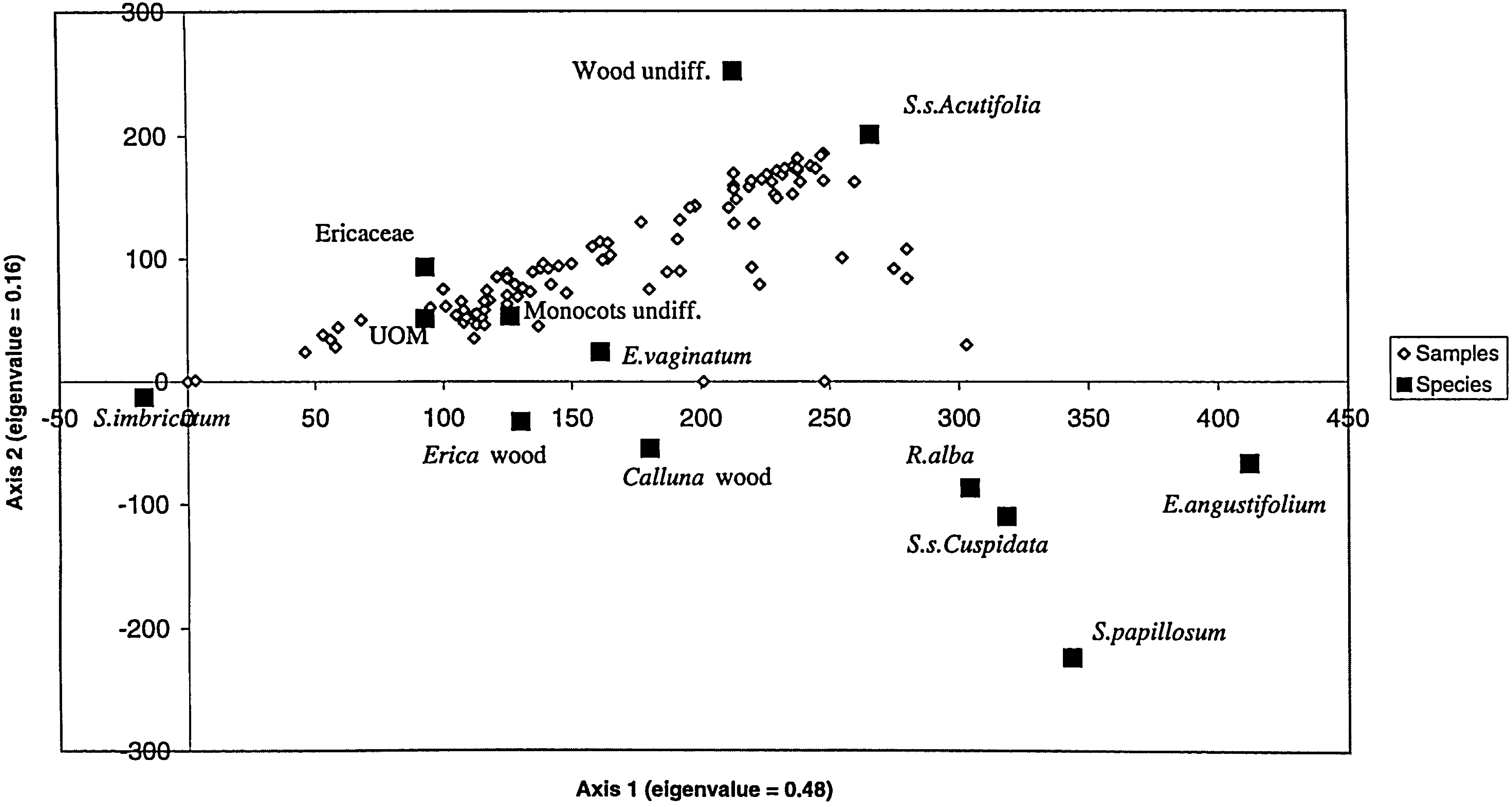


Figure 6.27 Temple Hill Moss DCA: axis 1 scores vs time.

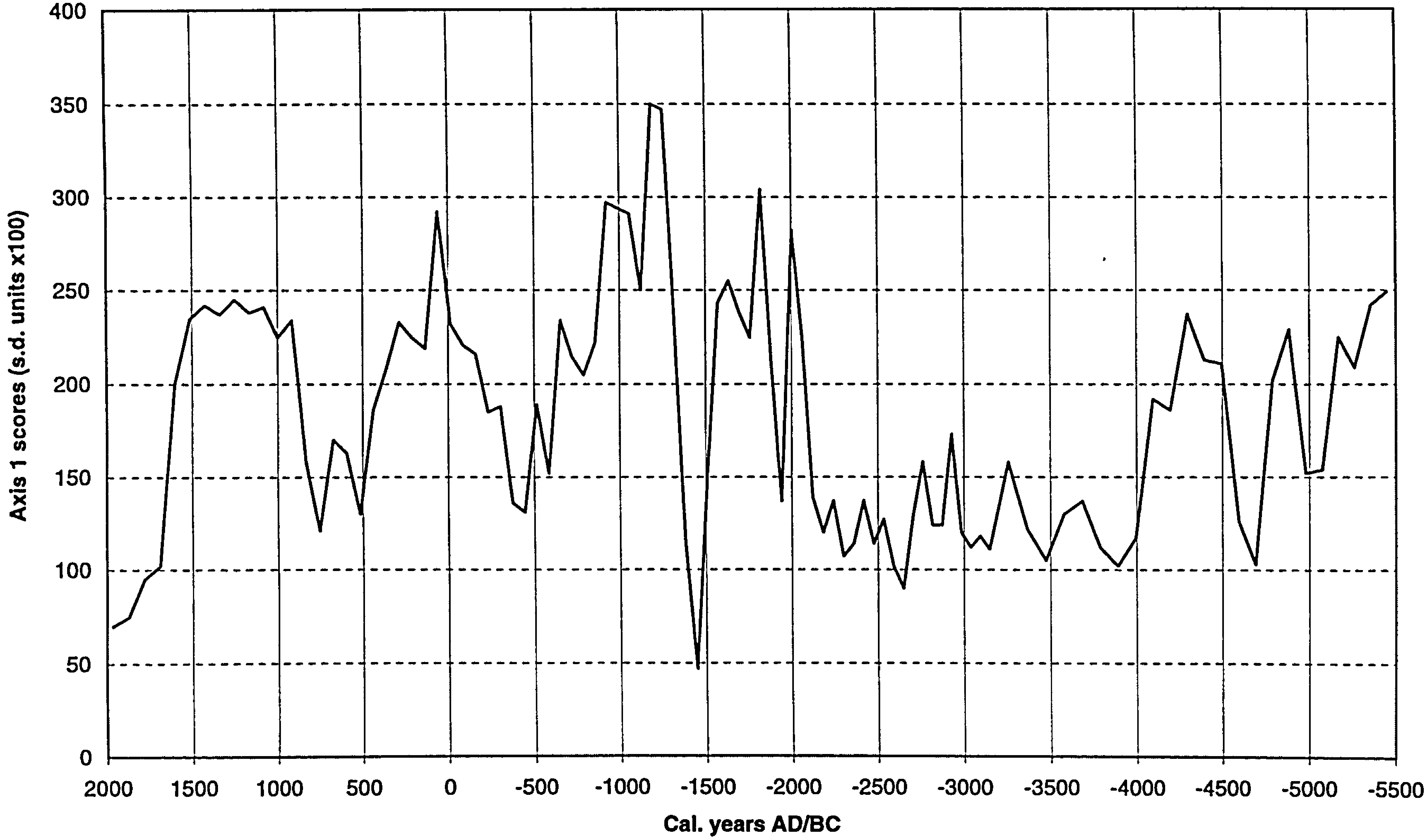


Figure 6.28 Temple Hill Moss - Testate amoebae data reconstructed water table depths +/- sample prediction errors vs time.

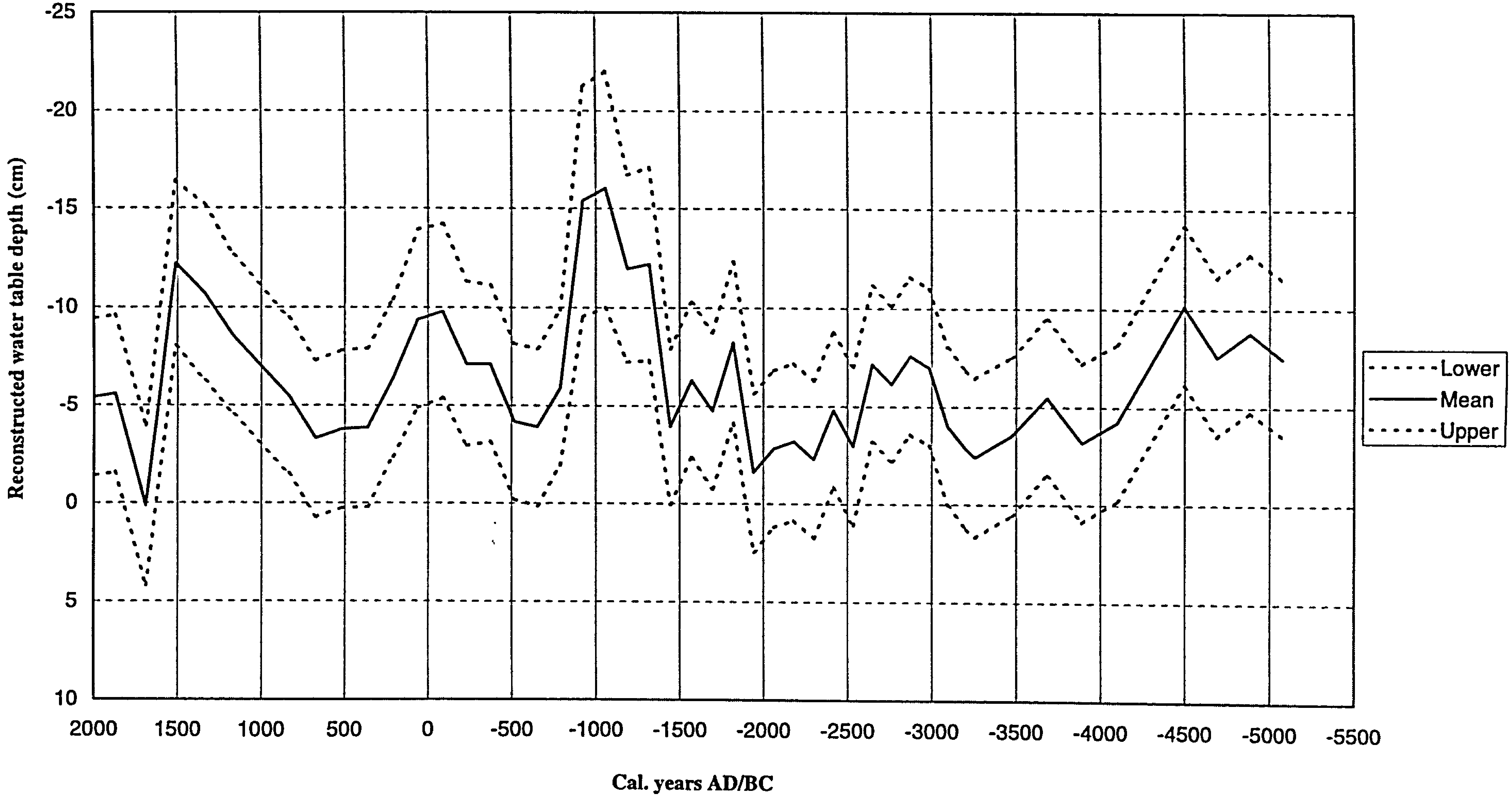
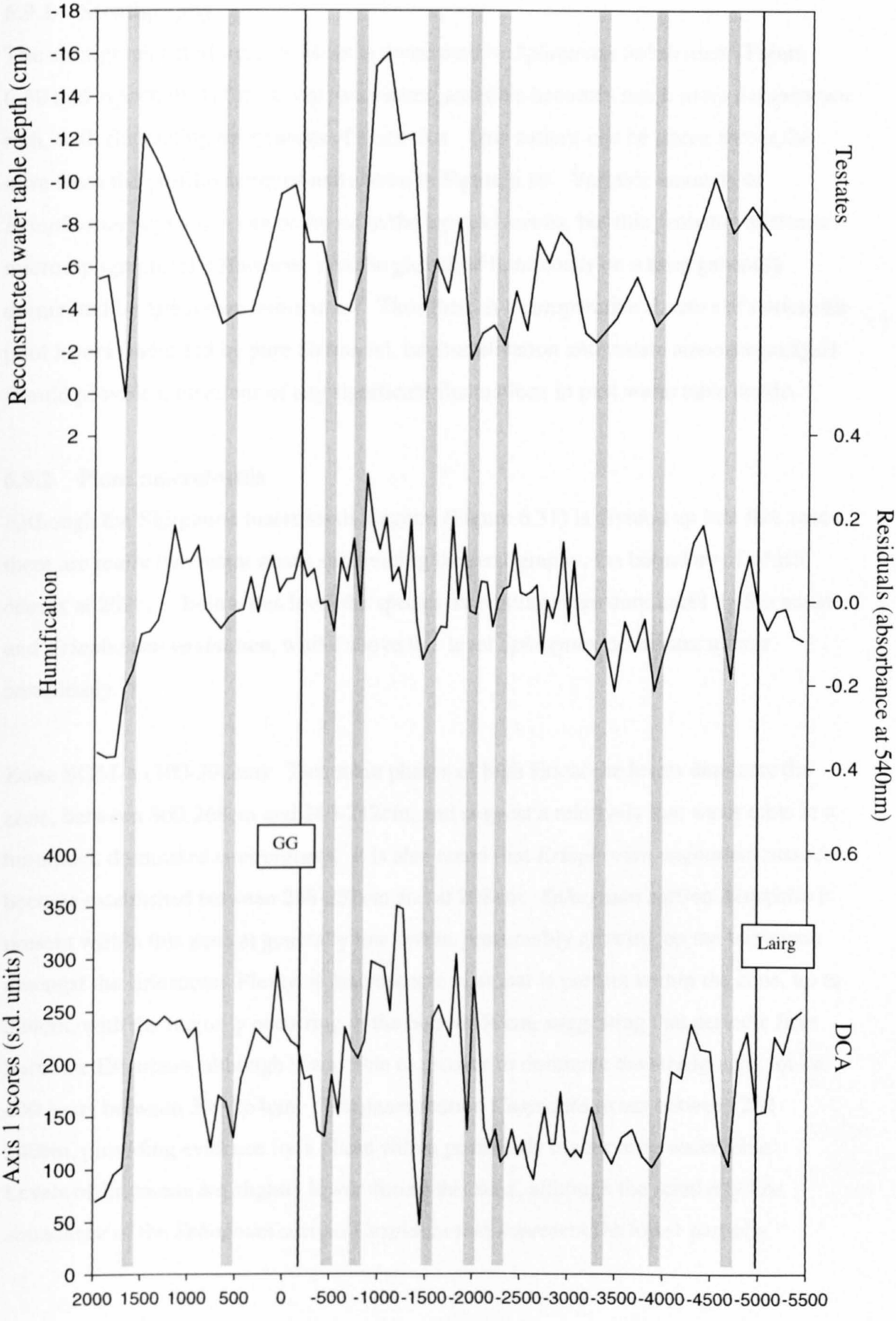


Figure 6.29 Temple Hill Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae data plotted against time. Solid vertical lines indicate tephras.



6.9 Shirgarton Moss

6.9.1 Stratigraphy

The stratigraphy at Shirgarton Moss is dominated by *Sphagnum imbricatum* (Figure 6.30 and Appendix 1) for the top two metres, and then becomes much more *Eriophorum* rich, with fluctuating abundances of Ericaceae. This pattern can be traced across the mire from the profiles sampled and shown in Figure 6.30. Variable amounts of *Eriophorum vaginatum* can be found in the top two metres, but this probably relates to microtopographical differences, and the growth of hummocks on a mire generally dominated by *Sphagnum imbricatum.* Thus there is a comparative absence of noticeable pool layers (indicated by pure Sh bands), but humification and testate amoebae analysis should provide indications of any significant fluctuations in past water table depths.

6.9.2 Plant macrofossils

Although the Shirgarton macrofossil diagram (Figure 6.31) is divided up into five zones, there are really two major zones subdividing the stratigraphy, the boundary of which occurs at 202cm. Below this level the species assemblages are dominated by Ericaceae and *Eriophorum vaginatum*, whilst above this level *Sphagnum* dominates almost completely.

Zone SGM-a (300-202cm): Two main phases of high Ericaceae levels dominate the zone, between 300-268cm and 240-212cm, and suggest a relatively low water table in a hummock dominated environment. It is also noted that *Eriophorum vaginatum* tussocks become established between 268-252cm and at 208cm. *Sphagnum* section *Acutifolia* is present within this zone at generally low levels, presumably growing on the hummock amongst the Ericaceae. Plenty of macroscopic charcoal is present within the zone, up to 224cm, with the majority occurring in the bottom 50cm, suggesting that periodic fires burnt the Ericaceae, although it was able to recover to dominate the stratigraphy for *ca.* 400 years between 300-264cm. *Sphagnum* section *Cuspidata* exists between 252-240cm, providing evidence for a phase with a potentially higher mire water table. Levels of Ericaceae are slightly lower during this time, although the relatively low abundance of the *Sphagnum* section *Cuspidata* may represent the lower part of a hummock, as it has been demonstrated that these species can exist in such a microform utilising the lateral transport of water (Rydin, 1985).

Zone SGM-b (202-78cm): *Sphagnum imbricatum* replaces *Eriophorum* at the start of the zone, possibly due to the raising of the water table as a result of increased precipitation, dated at cal. 910-760 BC (SRR-6344). *Sphagnum imbricatum* then dominates throughout the zone, with significant phases of *Sphagnum* section *Cuspidata* occurring at 128cm and 108cm, dated to *ca.* cal. AD 10 and *ca.* cal. AD 280 respectively from the interpolated dates. Consistently low values of *Sphagnum* section *Cuspidata* occur throughout the zone, possibly responding to an expanding/contracting low lawn/pool microform. Small amounts of *Eriophorum angustifolium* occur at 156cm and 144cm, indicating a wet lawn/pool microform was present.

Zone SGM-c (78-54cm): The plant macrofossils suggest a shift to wetter conditions as *Sphagnum imbricatum* is replaced by *Sphagnum papillosum* and *Sphagnum* section *Cuspidata*. The initial change in species assemblage almost certainly documents the onset of climatic deterioration at 76cm, as high levels of *Sphagnum* section *Cuspidata* are present (46%), which then remain at lower levels throughout the zone, probably still being too wet for *Sphagnum imbricatum* to return. This is dated to cal. AD 1120-1290 (SRR-6342), although the age/depth model from Shirgarton (Section 5.3.4) suggests this date is too young. It is dated to *ca.* cal. AD 770 by interpolation.

Zone SGM-d (54-30cm): *Sphagnum imbricatum* returns to dominate the assemblages for one last phase, possibly as a result of a change in the water table depth, and replaces *Sphagnum papillosum*. Small levels of *Scirpus cespitosus* are present in this zone, implying that the water table may have been lowered relative to the previous zone. *Sphagnum imbricatum* starts to decline towards the end of the zone (36cm) and is replaced again by *Sphagnum papillosum*, becoming extinct at 32cm, dated to cal. AD 1220-1390 (SRR-6341).

Zone SGM-e (30-0cm): The *Sphagnum papillosum* replacement of *Sphagnum imbricatum* is brief, as *Sphagnum magellanicum* becomes the dominant *Sphagnum* species. Interpreting the water table depth is difficult with the one dominant species able to tolerate variable water table depths, although a small but distinct rise in Ericaceae is apparent throughout this zone, indicating that the water table depth may have been lowered.

6.9.3 Humification

A major trend can be seen in the humification curve (Figure 6.32) and thus the detrending procedure was carried out on the raw data. The large departure and subsequent high humification values from 204cm downwards suggests that a species signal may be present in the data, as the top 200cm of the stratigraphy are *Sphagnum* dominated compared to the lower peats. Figure 6.33 plots the detrended data against the interpolated dates from the age/depth model. Clear peaks of low humification can be seen at *ca.* cal. 1750-1710 BC, 1380 BC, 850-200 BC, and less prominent peaks at *ca.* cal. AD 110, AD 390, and *ca.* cal. AD 570-1040. High levels of humification, and possible climatic amelioration phases occur at *ca.* cal. 1820 BC, 1450 BC, 1340-890 BC, 50 BC, AD 1610, and less prominently at *ca.* cal. AD 510, and from *ca.* cal. AD 1120.

6.9.4 Testate amoebae

The testate amoebae data are shown in Figure 6.34, and are dominated by the hygrophilous *Amphitrema* species and the more cosmopolitan, probably xeric, *Difflugia pulex.*

Zone SGM-a (300-202cm): The xerophilous testate *Difflugia pulex* (Hendon, 1998) dominates the zone, with lower, but ever present, values of *Amphitrema flavum. Nebela militaris* is also present throughout the zone, a taxon which tends to exist in relatively deep water tables (Woodland, 1996). Low water tables are thus inferred for the zone. However, there are small peaks of *Amphitrema wrightianum* of 248cm and 208cm, and although these may be responding to ephemeral events, the peak at 248cm corresponds with a phase of *Sphagnum* section *Cuspidata*, indicating the possibility of raised water table depths and a climatic deterioration.

Zone SGM-b (202-78cm): At 200cm, the start of the zone, *Difflugia pulex* is replaced by the hygrophilous *Hyalosphenia papilio*, and the xerophilous *Hyalosphenia subflava* and *Assulina muscorum*. The concentration of testates at this depth is very low, and may be a result of the expansion of *Sphagnum imbricatum* over a previously relatively dry flora. The fauna is then dominated by the hydrophilous *Amphitrema wrightianum* as well as *Amphitrema flavum*, with the exception of 192-184cm and 136-128cm when *Difflugia pulex* returns, indicating a relative lowering of the water table and suggesting a climatic amelioration. The second dry phase occurs when the Glen Garry tephra was deposited.

Zone SGM-c (78-54cm): This phase is characterised by an extremely high abundance of *Amphitrema wrightianum*, indicating a wet mire surface.

Zone SGM-d (54-30cm): *Amphitrema wrightianum* slowly becomes replaced by *Amphitrema flavum*, possibly indicating a slight drop in the water table. However, there is a large increase in abundance of *Amphitrema wrightianum* at 32cm suggesting a possible climatic deterioration. This occurs at the same level as the extinction of *Sphagnum imbricatum*.

Zone SGM-e (30-0cm): The most diverse fauna is found in the top 30cm of the profile, and is dominated by *Arcella discoides* type, as well as the more xerophilous *Difflugia pulex* and *Heleopera petricola*. Woodland (1996) suggested that although *Arcella discoides* type is regarded as hydrophilous, its ability to live within shallow water films allows it to live within a range of water table depths. The presence of the xerophilous species may therefore suggest a comparatively lower water table.

6.9.5 DCA

The DCA of Shirgarton Moss (Figure 6.35) plots axis 1 and axis 2 scores with a moisture gradient being clearly visible. The positioning of *Sphagnum imb/mag* and *Sphagnum* section *Acutifolia* towards the centre of axis 1 indicates that these species grow over both wet and dry microforms, as is clear from their ecologies (sections 6.1.19-6.1.21). However, as *Sphagnum imbricatum* and *Sphagnum magellanicum* dominate the majority of the top two metres of the stratigraphy (Figure 6.31), the ordination will fail to pick up much variation within these zones, highlighting the need for the other proxy techniques.

The axis 1 scores are plotted against time (from the interpolated dates of the age/depth model) and shown in Figure 6.36. Periodic wet fluctuations can be seen in the bottom metre of stratigraphy, around *ca.* cal. 1570 BC and more noticeably around *ca.* cal. 1340-1260 BC. The largest fluctuation in the DCA axis 1 scores (Figure 6.36) occurs at *ca.* cal. 850 BC and signifies the onset of dominance of *Sphagnum imbricatum*, and is interpreted as a climatic deterioration. van Geel *et al.* (1996) discuss a climatic deterioration which occurred throughout northwest Europe at around the same time.

After this event, the curve remains relatively static, apart from minor fluctuations at *ca.* cal. AD 60, AD 340 and *ca.* cal. AD 1520-1610, interpreted as wet shifts, due to the presence of *Sphagnum* section *Cuspidata* and *Sphagnum papillosum* at these times respectively (Figure 6.31). The largest climatic shift to wetter conditions over the last 2000 years at Shirgarton is found between *ca.* cal. AD 770-1120, represented by the dominance of *Sphagnum papillosum* (Figure 6.31), subject to the dating control as discussed in section 5.3.4.

Dryer phases can be recognised in Figure 6.36 around *ca.* cal. 1780-1710 BC, 1490-1410 BC, 1180-1060 BC, and *ca.* cal. 930 BC. After this time, the curve does not fluctuate to what could be considered drier conditions.

6.9.6 Testate amoebae transfer function

The reconstructed water table depths from the testate amoebae data (Figure 6.37) show major variations in the period from *ca.* cal. 2000-100 BC, and indicate a relatively wet mire surface from *ca.* cal. 760 BC to present with minor fluctuations, due to the relative dominance of the hydrophilous *Amphitrema wrightianum* (Figure 6.34). A relative decrease in mire water table depths also occurs around *ca.* cal. 1570 BC, with the main wet shifts being recognised around *ca.* cal. 1340 BC, and after *ca.* cal. 760 BC with pronounced drops around *ca.* cal. 490 BC and *ca.* cal. AD 830-1120. Relatively dry phases can be identified around *ca.* cal. 1790-1750 BC, 1490 BC, 850 BC, with slightly lower water table depths occurring around *ca.* cal. AD 10 and *ca.* cal. AD 1430.

6.9.7 Comparison of the three proxy techniques

A composite diagram has been produced from the three proxy reconstructions (Figure 6.38) which show five main periods of agreement for climatic deteriorations (greyscale lines). The wet phases are identified around *ca.* cal. 1570 BC, 1340 BC, 800 BC, AD 1050, and *ca.* cal. AD 1550. The deterioration around 850 BC identified in the humification and plant macrofossil records is only noted in all three proxies by *ca.* cal. 800 BC, as the testate record shows an abundance in the relatively xeric *Difflugia pulex* before this time. It is also possible that the two slight peaks in the DCA record around *ca.* cal. AD 60 and *ca.* cal. AD 340 are recognised in the other two reconstructions, as there are slight dips in the humification record at these times, and the reconstructed depths from the testate data show very high water tables. Dry phases can be interpreted in each proxy around *ca.* cal. 1450 BC, 1100 BC, 900 BC, 50 BC, and *ca.* cal. AD 1270-1430.

Figure 6.30 Shirgarton Moss:Stratigraphy

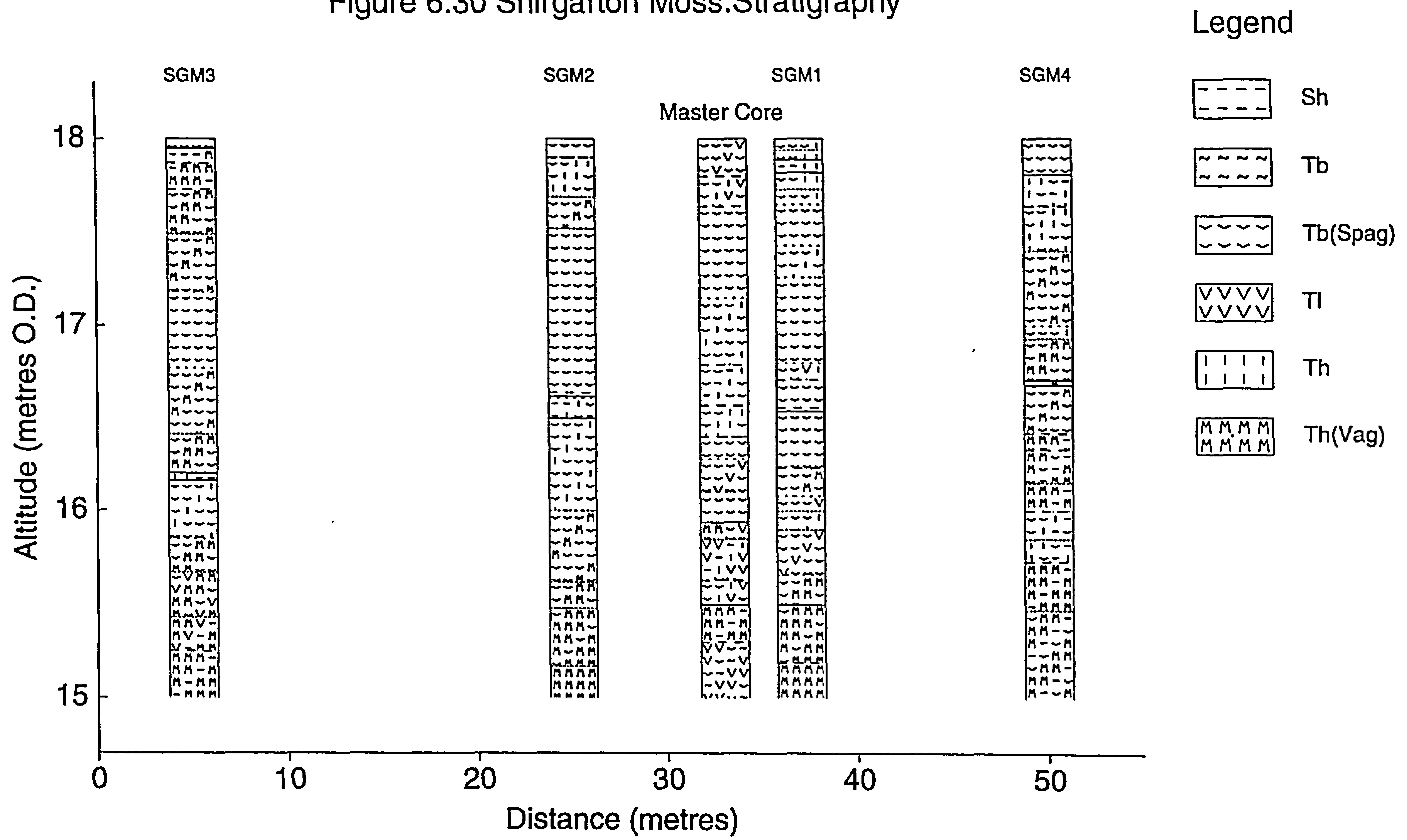


Figure 6.31 Shirgarton Moss – Macrofossil diagram

Analysis: Pete Langdon

Figure 6.32 Shirgarton Moss - Humification data. The raw data are presented at the top, and have been detrended by linear regression, producing the lower data set.

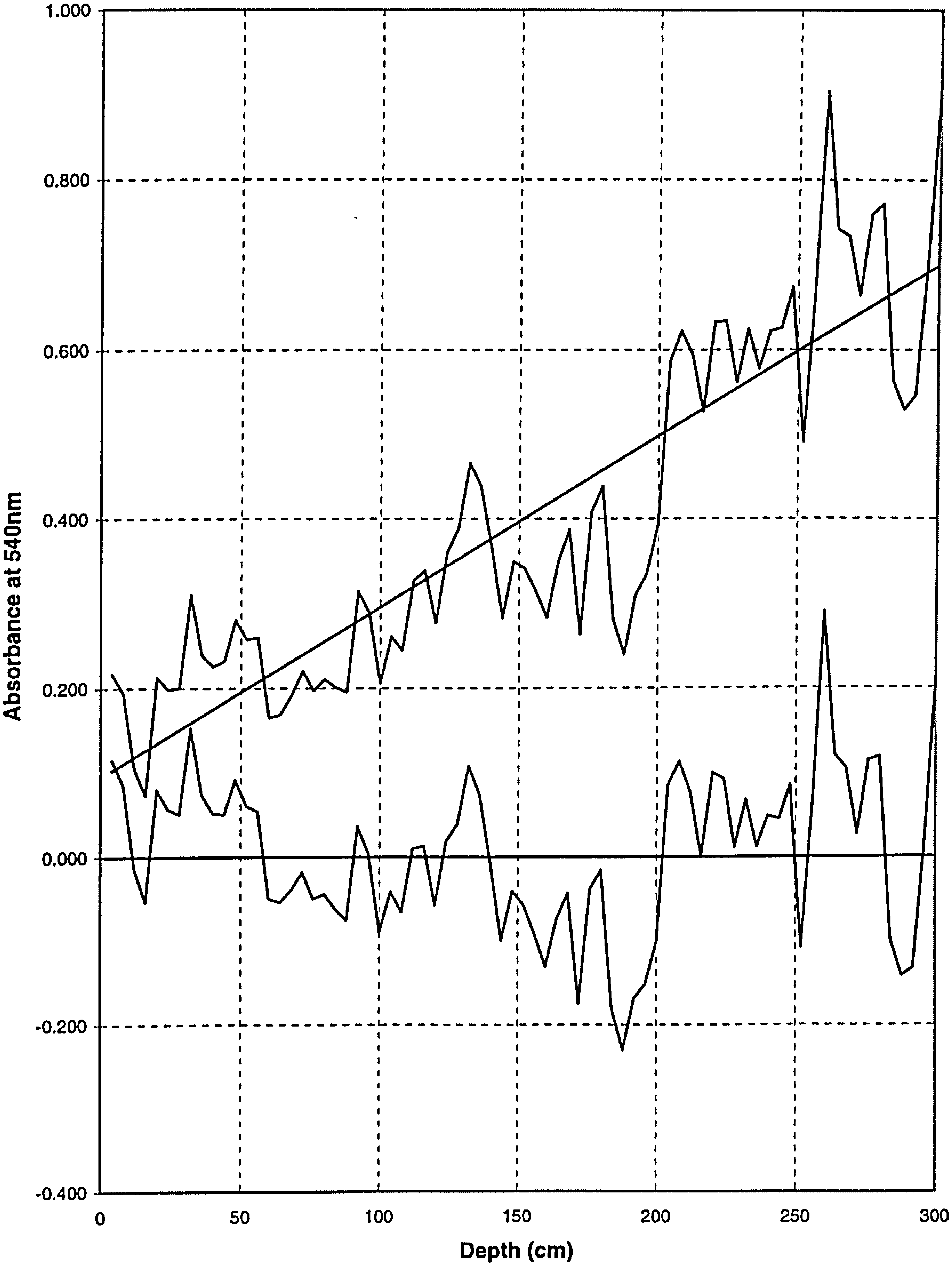


Figure 6.33 Shirgarton Moss - Detrended humification vs time.

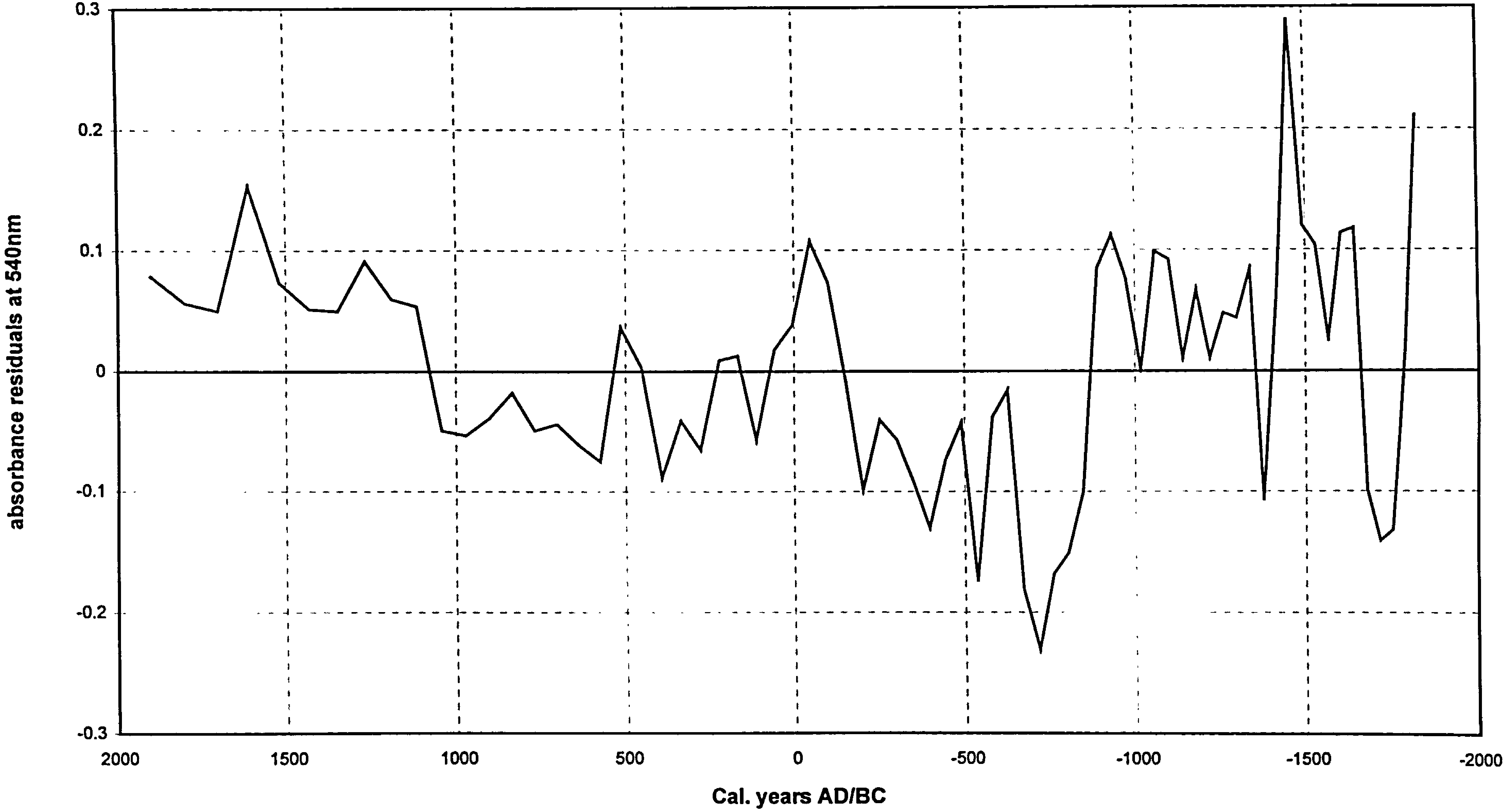


Figure 6.34 Shirgarton Moss – Testate amoebae

Analysis: Pete Langdon

Figure 6.35 Shirgarton Moss - Plant macrofossils: detrending by segments.

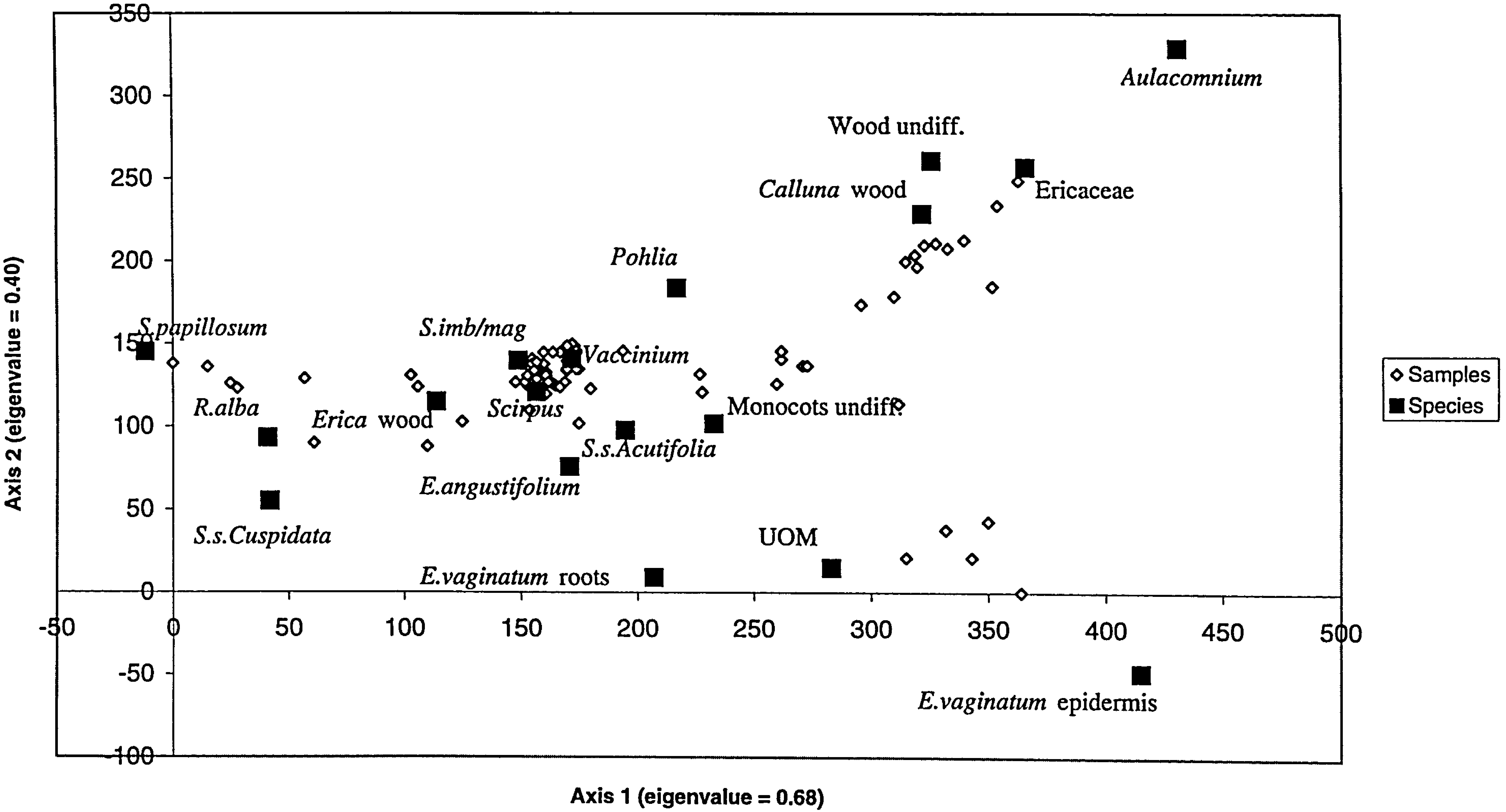


Figure 6.36 Shirgarton Moss - Plant macrofossils DCA: axis 1 scores vs time.

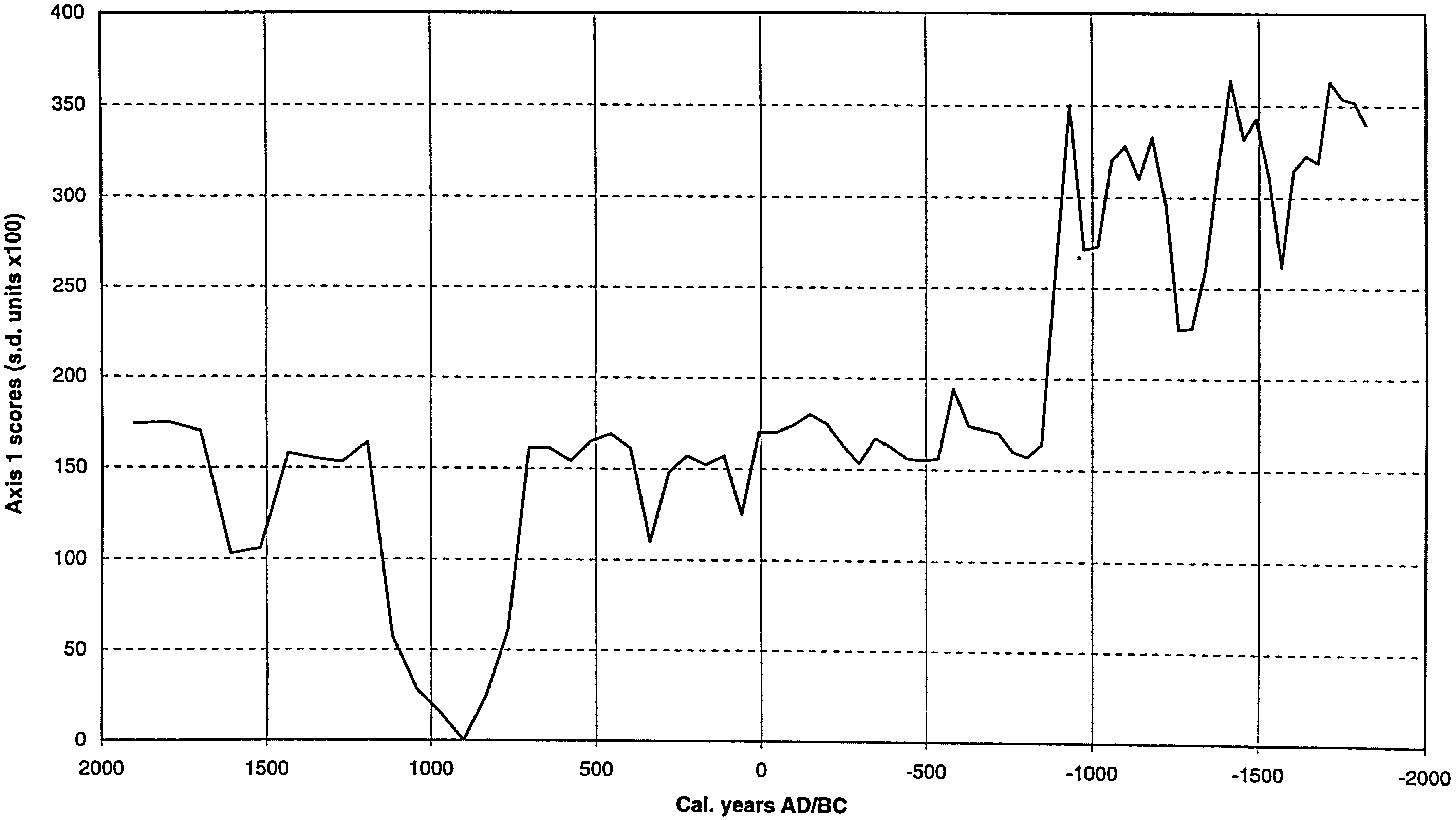


Figure 6.37 Shirgarton Moss - Testate amoebae reconstructed water table depths +/- sample prediction errors vs time.

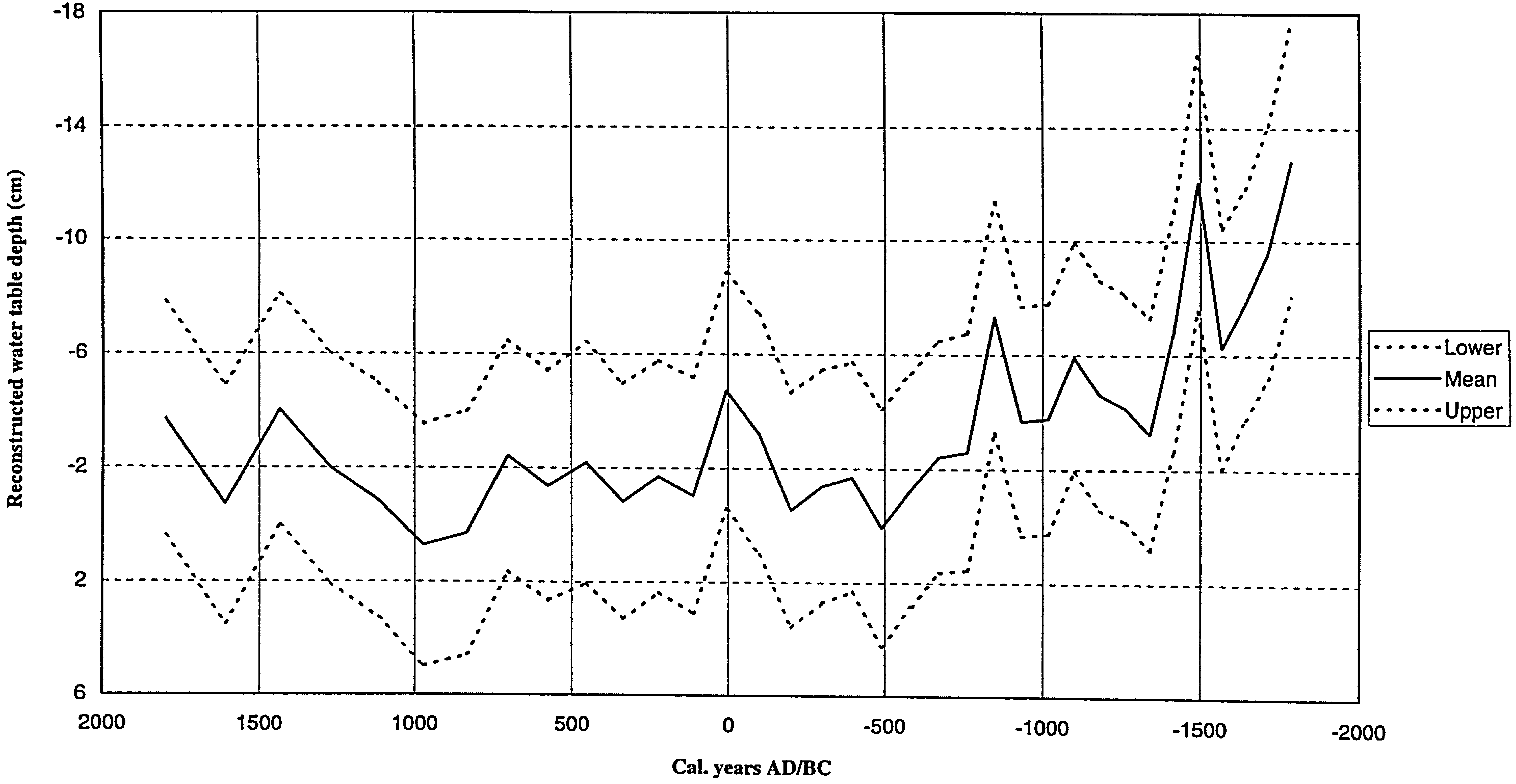
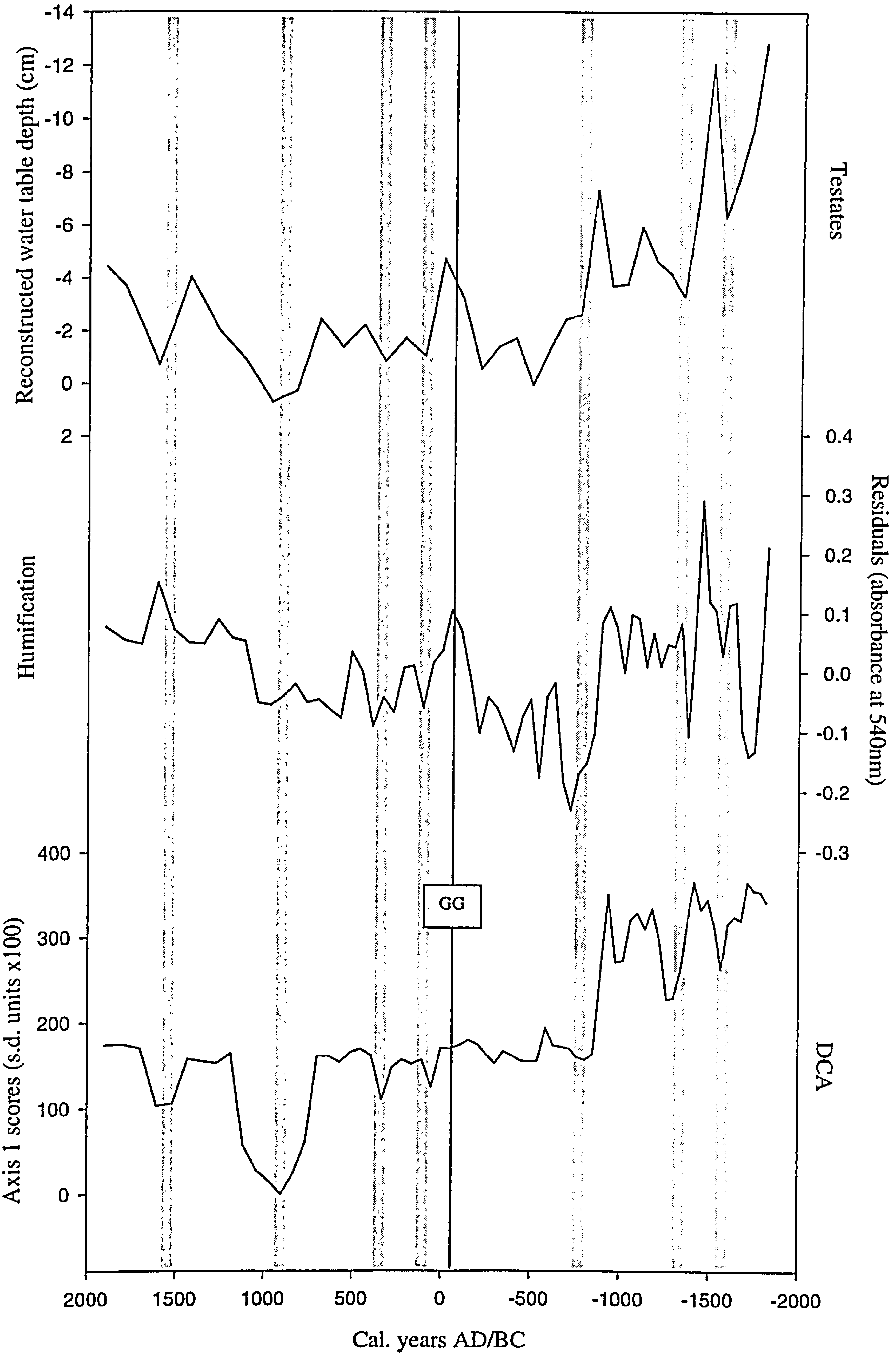


Figure 6.38 Shirgarton Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae data plotted against time. Solid vertical line indicates a tephra.



6.10 Mallachie Moss

6.10.1 Stratigraphy

The stratigraphy at Mallachie Moss can be traced laterally across each profile sampled (Figure 6.39 and Appendix 1). The *Sphagnum* rich surface gives way to a highly humified, monocot rich (often *Eriophorum vaginatum*) layer which can be identified across the whole mire surface (personal observation), indicating a possible past disturbance to the mire. However, the stratigraphy below displays a remarkable coherence between cores, with pool layers (identified by the pure Sh layers) being found at similar depths in each core, forming the basis for sound palaeoclimatic reconstructions (Barber *et al.*, 1998).

6.10.2 Plant macrofossils

Unlike most bogs studied in the British Isles, the peat stratigraphy at Mallachie Moss contains no *Sphagnum imbricatum*, a specifically oceanic moss (Flatberg, 1986; Daniels and Eddy, 1990), suggesting that the site is subject to more continental climate conditions. Mallachie Moss is therefore a site of great importance when comparing it to more oceanic sites. The macrofossil diagram is shown in Figure 6.40, and is divided into six zones.

Zone MAL-a (300-178cm): *Sphagnum* section *Acutifolia* dominates the stratigraphy, with periodic increases in Ericaceae, *Eriophorum vaginatum*, and *Scirpus cespitosus*, on what initially appears to be a cyclical pattern.

An abundance of *Eriophorum vaginatum* at 292cm is accompanied by relatively high levels of Ericaceae, suggesting a hummock microform was present, indicative of a lowered water table. The *Sphagnum* was then able to recover, potentially indicating a rise in water table depths, only to be partially replaced by Ericaceae at 260cm, the same time as the Hekla-4 deposition, suggesting the climate was experiencing a dry phase. High levels of macroscopic charcoal are also present at 260cm, indicating that fires may have occurred. *Sphagnum* is again able to recover, possibly due to fluctuations in the water table, until it is overgrown again by another *Eriophorum* tussock at 224cm.

Scirpus cespitosus is also found at this depth, and after *Sphagnum* section *Acutifolia* was again able to recolonise the mire surface from 220-204cm, *Scirpus* was found in abundance with high levels of Ericaceae between 200-180cm, indicating a return to lowered water table depths.

Zone MAL-b (178-150cm): A relatively high coverage of *Sphagnum* section *Acutifolia* was present in this zone, accompanied by monocots, the remains of which only *Eriophorum vaginatum* and *Scirpus cespitosus* are recognisable. Leaves of *Sphagnum* section *Cuspidata* have also been recovered from this zone, although the presence of *Eriophorum* and *Scirpus* make inferences for past water table depths difficult. However, a peak in *Sphagnum* section *Cuspidata* leaves at 152cm indicates a high water table, and is suggestive of a climatic deterioration. This event has been dated to cal. 1060-830 BC (SRR-6349), and could be related to the climatic deterioration around 800 BC as reviewed by van Geel *et al.* (1996; 1998).

Zone MAL-c (150-118cm): High levels of *Sphagnum* section *Acutifolia* dominate the stratigraphy from 148-132cm, until they became replaced by *Scirpus cespitosus* and *Eriophorum vaginatum* at 128cm, followed by high levels of UOM from 124-120cm. The demise of *Sphagnum* was accompanied by increasing amounts of macroscopic charcoal, indicative of localised fires. It is suggested that the high levels of UOM were related to the fires, and not necessarily indicative of changes in climate. However, such large volumes of charcoal are likely to have been produced from severe fires (Kenworthy, 1963), suggesting they occurred when mire water levels were low as a result of prolonged low effective precipitation (Mauquoy, 1997).

Zone MAL-d (118-42cm): The peak in abundance of *Sphagnum* section *Cuspidata* at 116cm may not be associated with a climatic deterioration, but be the result of an increase in water retaining capacity as charcoal reduces the density of large pores in the upper peats (Mallik *et al.*, 1984). However, very little charcoal was found at 116cm, only above and below stratigraphically, suggesting that fires may have been reduced by an increase in effective precipitation. High levels of Ericaceae, UOM, *Cenococcum* and charcoal are present at 112cm, indicative of relatively lower water tables. This assemblage is followed by a dominance of *Sphagnum* section *Acutifolia* throughout the zone, interspersed with periodic increases in more xerophilous flora assemblages, of *Eriophorum vaginatum*, *Scirpus cespitosus*, UOM, and Ericaceae, notably at 88cm and 76cm. An increase in abundance of *Sphagnum* section *Cuspidata* between 104-96cm suggests a raised mire water table, possibly indicative of an increased effective precipitation regime. *Sphagnum* section *Acutifolia* declined at the end of the zone, being overgrown by an *Eriophorum vaginatum* tussock associated with Ericaceae, signifying a hummock microform and relative lowering of the mire water table.

Zone MAL-e (42-26cm): *Sphagnum* was virtually totally replaced by a UOM 'layer', with smaller levels of Ericaceae and monocotyledons, probably *Eriophorum vaginatum* due to the abundance of *Eriophorum vaginatum* spindles. This layer can be traced across the mire surface, suggesting that an event affected the whole mire. The mire appears to have dried out, possibly aided by a natural climatic amelioration. However, large amounts of charcoal are found within the zone, suggesting that possibly a fire decimated the flora, resulting in a hiatus within the stratigraphy, and coupled with low mire water tables, a dry flora dominated with high levels of decomposition. The age/depth model (section 5.3.5) suggests a hiatus may be present.

Zone MAL-f (26-0cm): The UOM 'layer' was replaced almost exclusively by the hygrophilous *Sphagnum papillosum* and *Sphagnum* section *Cuspidata*, which was eventually succeeded by *Sphagnum* section *Acutifolia* at 8cm. Field evidence shows that large areas of the mire are presently colonised by *Sphagnum magellanicum*, even though none was recorded by this profile.

6.10.3 Humification

Very little deterioration with depth can be seen on the humification profile of Mallachie Moss (Figure 6.41), although a slight trend can be applied to the curve, so the detrending technique was applied to the data. Utilising the interpolated dates from the age/depth model (section 5.3.5) the humification data has been plotted against a time series (Figure 6.42). Periods of lower humification can be identified at *ca.* cal. 2850 BC, 2560-2510 BC, 2250-2200 BC, 2040 BC, 1360 BC, 730-600 BC, 30 BC, AD 510, and *ca.* cal. AD 770-1040. Peats which have undergone phases of increased humification can be recognised at *ca.* cal. 2710-2660 BC, 1470-1410 BC, 1170 BC, 990 BC, 860 BC, 460-390 BC, AD 270-350, AD 600, and *ca.* cal. AD 1230-1630, although there may be a hiatus within the last pair of dates as described above.

6.10.4 Testate amoebae

The testate amoebae data are shown in Figure 6.43. The assemblages are dominated by *Amphitrema flavum*, with other taxa only reaching high abundances periodically.

Zone MAL-a (300-178cm): *Amphitrema flavum* tests dominate the zone, as well as the majority of the stratigraphy. Low but inconsistent values of *Assulina muscorum* are also present, probably responding to ephemeral events, although peaks in the hygrophilous *Hyalosphenia papilio* at 272cm and 248cm may be indicative of a raised water table. Both depths correspond to phases of almost pure *Sphagnum* section *Acutifolia* domination and low humification, indicating a climatic deterioration. The other notable feature from this zone is a rise in abundance of the xerophile *Difflugia pulex* from 200-192cm, coupled with increases in UOM, Ericaceae, and *Scirpus cespitosus*, indicative of a hummock environment, and corresponding lowered water table.

Zone MAL-b (178-150cm): The zone is characterised by an initial peak in *Assulina muscorum*, being replaced by the hydrophile *Amphitrema wrightianum* suggesting a climatic deterioration around 152cm, which coincides with high levels of *Sphagnum* section *Cuspidata*, dated to cal. 1060-830 BC (SRR-6349). High levels of *Amphitrema flavum* are maintained, accompanied by low but consistent values of *Nebela* species, presumably responding to local fluctuations in the water table and vegetation.

Zone MAL-c (150-118cm): A high abundance of *Amphitrema flavum* tests dominate the fauna, with levels of *Amphitrema wrightianum* decreasing throughout the zone, and slight peaks in *Difflugia pulex* at the start of the zone (144cm) and *Assulina muscorum* later (128cm). Testate concentrations are very low at 128cm, beneath 4000 tests per cm^3, and the concentration was so low at 120cm that a count was not undertaken. It is suggested that this is a response to the large fires, indicated from the high abundance of macroscopic charcoal from these samples. Reconstructing the water table depths from these assemblages is therefore difficult.

Zone MAL-d (118-42cm): Levels of *Amphitrema flavum* remain high, with two major exceptions: when they were replaced at 112cm by *Cyclopyxis arcelloides* type, and between 96-80cm, where they were first replaced by the xerophilous *Hyalosphenia subflava*, and then by the hydrophilous *Arcella discoides* type. The peak of *Cyclopyxis arcelloides* type at 112cm coincides with a peak in Ericaceae, high levels of charcoal and *Cenococcum*, and low testate concentrations. It is possible that it is one species which is able to tolerate fires better than others, or alternatively is the first to respond to conditions after a fire. There was no sample taken at 116cm to correlate with the peak in *Sphagnum* section *Cuspidata* at this level. The presence of *Hyalosphenia subflava* from 96-88cm indicates a lowering of the water table, followed by a subsequent rise in water depth by an abundance of *Arcella discoides* type at 80cm. The presence of the xerophilous *Euglypha rotunda* and *Hyalosphenia subflava* at 48cm, replacing already low levels of *Amphitrema wrightianum* suggest a lower water table due to reduced effective precipitation.

Zone MAL-e (42-26cm): Only one sample was obtained from this zone as concentrations were too low for a testate count at 32cm. A huge peak in *Hyalosphenia subflava* at 40cm indicates a low water table depth.

Zone MAL-f (26-0cm): As with previous diagrams, the 'surface' zone, usually containing non-fossilised Sphagna exhibits the most diverse fauna for the stratigraphy. This zone is dominated by *Arcella discoides* type, indicating a high water table under conditions of an increased effective precipitation regime.

6.10.5 DCA

The ordination plot of axis 1 against axis 2 from the DCA is shown in Figure 6.44, on which a clear moisture gradient can be seen. The dry indicators, *Scirpus cespitosus*, UOM, and Ericaceae are all placed at one end of the axis, with the more hygrophilous *Sphagnum papillosum* and *Sphagnum* section *Cuspidata* at the opposite end. The dominant macrofossil throughout the stratigraphy, *Sphagnum* section *Acutifolia* (Figure 6.40), is placed near the middle of axis 1 signifying its bimodal distribution. The axis 1 scores have been plotted against time (Figure 6.45) from the interpolated dates from the age/depth model. The major wet climatic phases can be identified around *ca.* cal. 2610-2510 BC, 2300-1980 BC, 1820-1700 BC, 1470-1410 BC, 920 BC, 800 BC, AD 40-270, AD 510, AD 680-1040, and *ca.* cal. AD 1850 to present. Corresponding dry phases can be recognised around *ca.* cal. 2760 BC, 2350 BC, 1870 BC, 1530 BC, 1300 BC, 860 BC, 460-250 BC, 110 BC, AD 350, AD 600, and *ca.* cal. AD 1320-1740.

6.10.6 Testate amoebae transfer function

The testate amoebae data shows very little variation until the most recent 2000 years (Figure 6.46), due to the dominance of *Amphitrema flavum* throughout this period (Figure 6.43). Slightly higher reconstructed water table depths occur around *ca.* cal. 2510 BC, 2200 BC, 1650 BC, and *ca.* cal. 800 BC, with distinctly more significant fluctuations in the water table depth around *ca.* cal. AD 510, AD 860-950, and *ca.* cal.. AD 1850 to present. There are three suggestions of drier phases, around *ca.* cal. 1530-1470 BC, AD 200-350, and *ca.* cal. AD 1320-1630.

6.10.7 Comparison of the three proxy techniques

The three effective precipitation proxy curves have been plotted together (Figure 6.47) and suggest that the curves can be correlated on up to nine major phases of climatic deterioration (greyscale lines). The main wet phases can be identified around *ca.* cal. 2500 BC, 2200 BC, 1700 BC, 1350 BC, 850 BC, 10 BC, AD 500, AD 950-1040, and *ca.* cal. AD 1850. Dry phases can be recognised around *ca.* cal. 2700 BC, 2350 BC, 1530 BC, 860 BC, 390 BC, AD 350, and *ca.* cal. AD 1420-1630. However, it must be noted that at Mallachie, although the DCA axis 1 scores and humification values correspond to a high degree, there are less fluctuations within the testate data, due to the dominance of *Amphitrema flavum* for the majority of the sequence.

Figure 6.39 Mallachie Moss: Stratigraphy

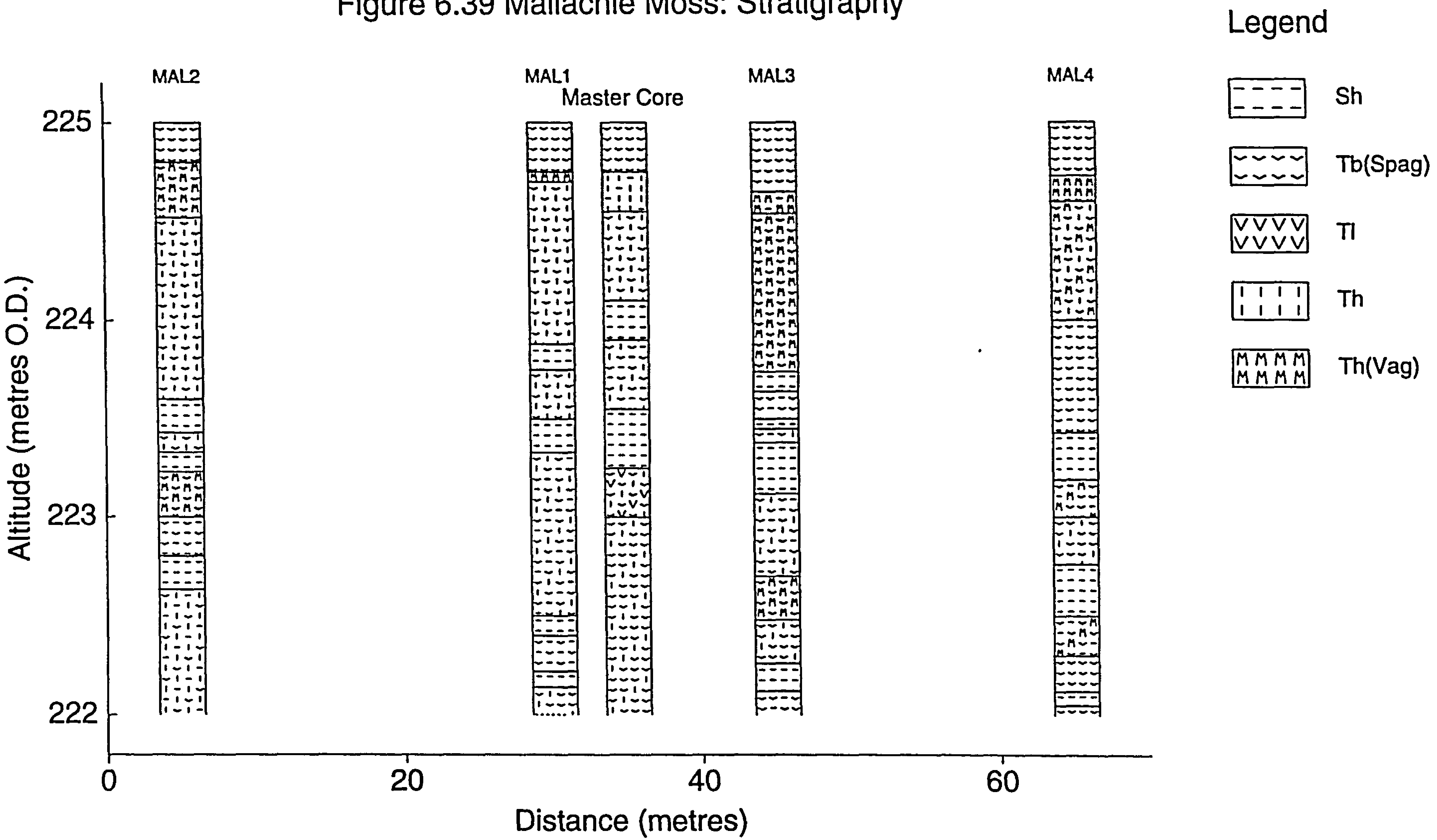


Figure 6.40 Mallachie Moss – Macrofossil diagram

Analysis: Pete Langdon

Figure 6.41 Mallachie Moss - Humification data. The raw data are presented at the top, and have been detrended by linear regression, producing the lower data set.

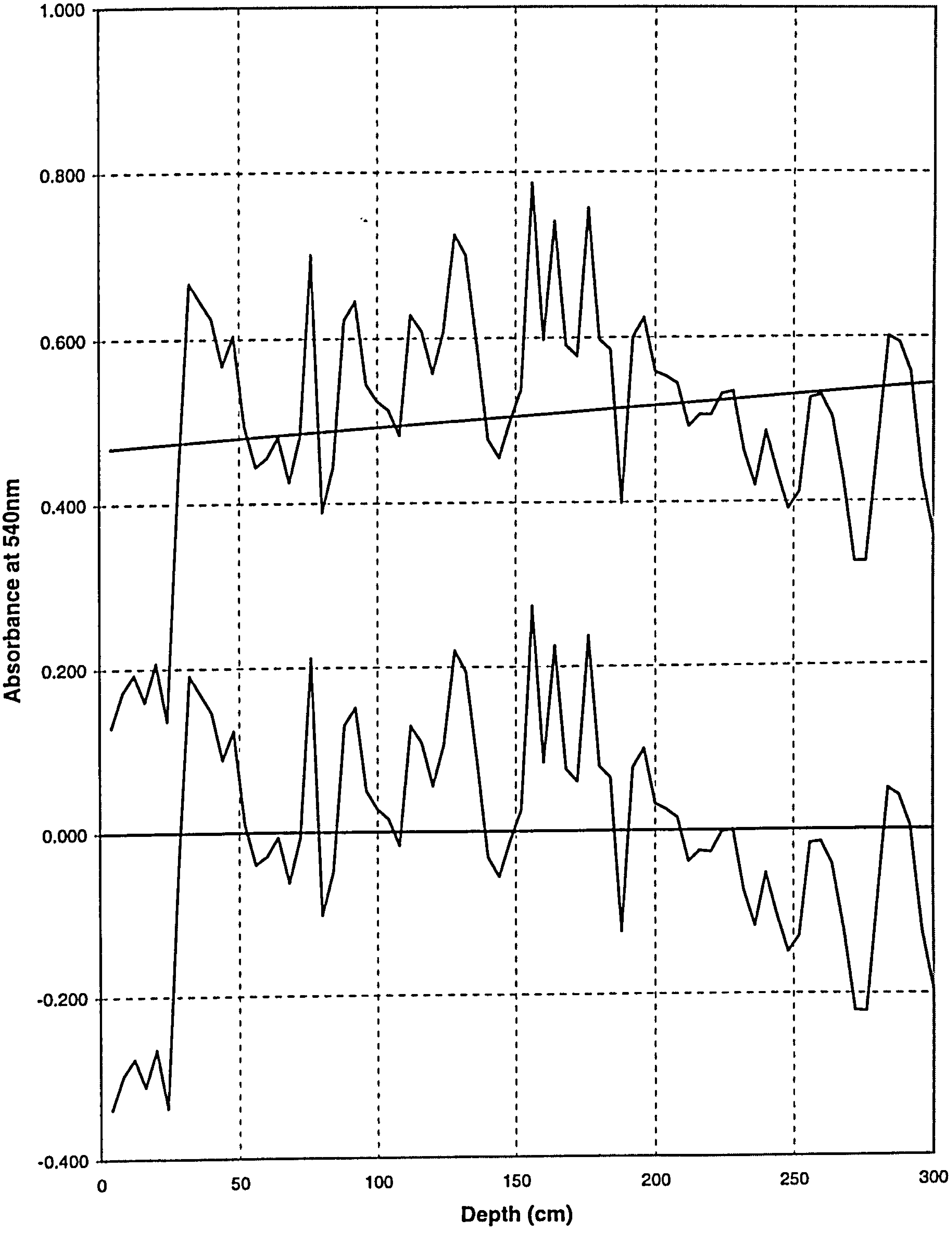


Figure 6.42 Mallachie Moss - Detrended humification vs time.

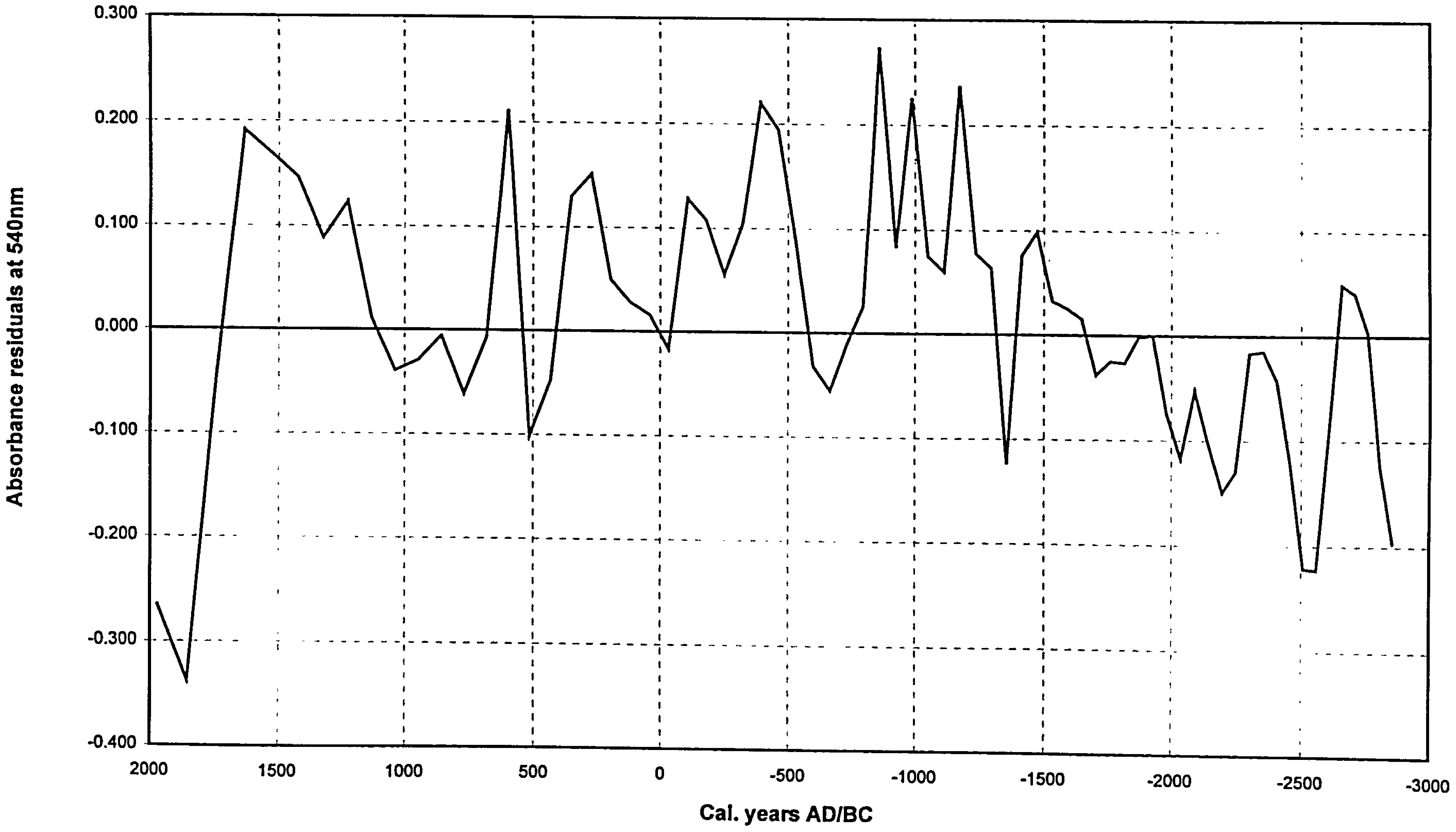


Figure 6.43 Mallachie Moss – Testate amoebae

Analysis: Pete Langdon

Figure 6.44 Mallachie Moss - Plant macrofossils: detrending by segments.

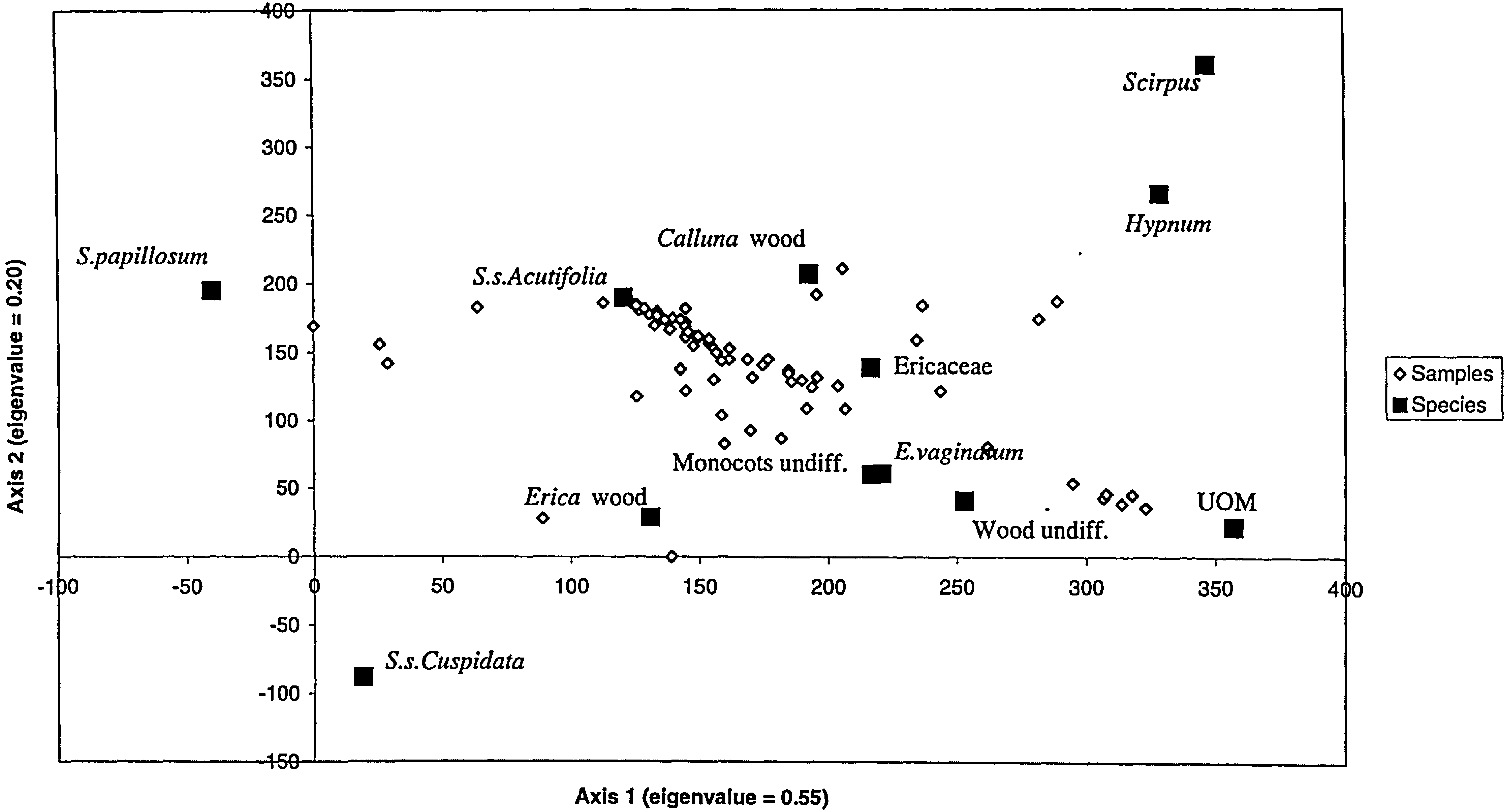


Figure 6.45 Mallachie Moss - Plant macrofossils DCA: axis 1 scores vs time.

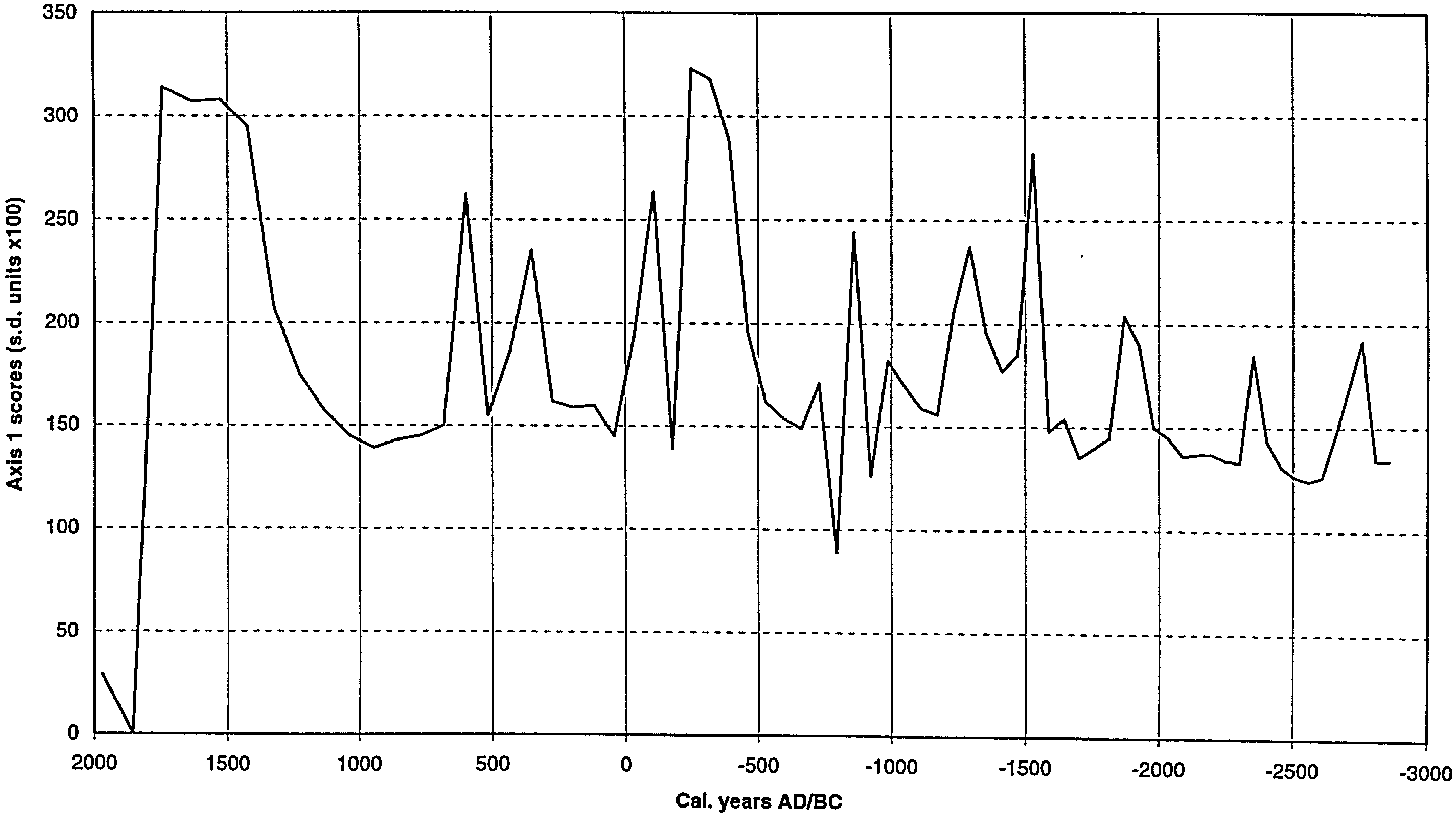


Figure 6.46 Mallachie Moss - Testate amoebae reconstructed water table depths +/- sample prediction errors vs time.

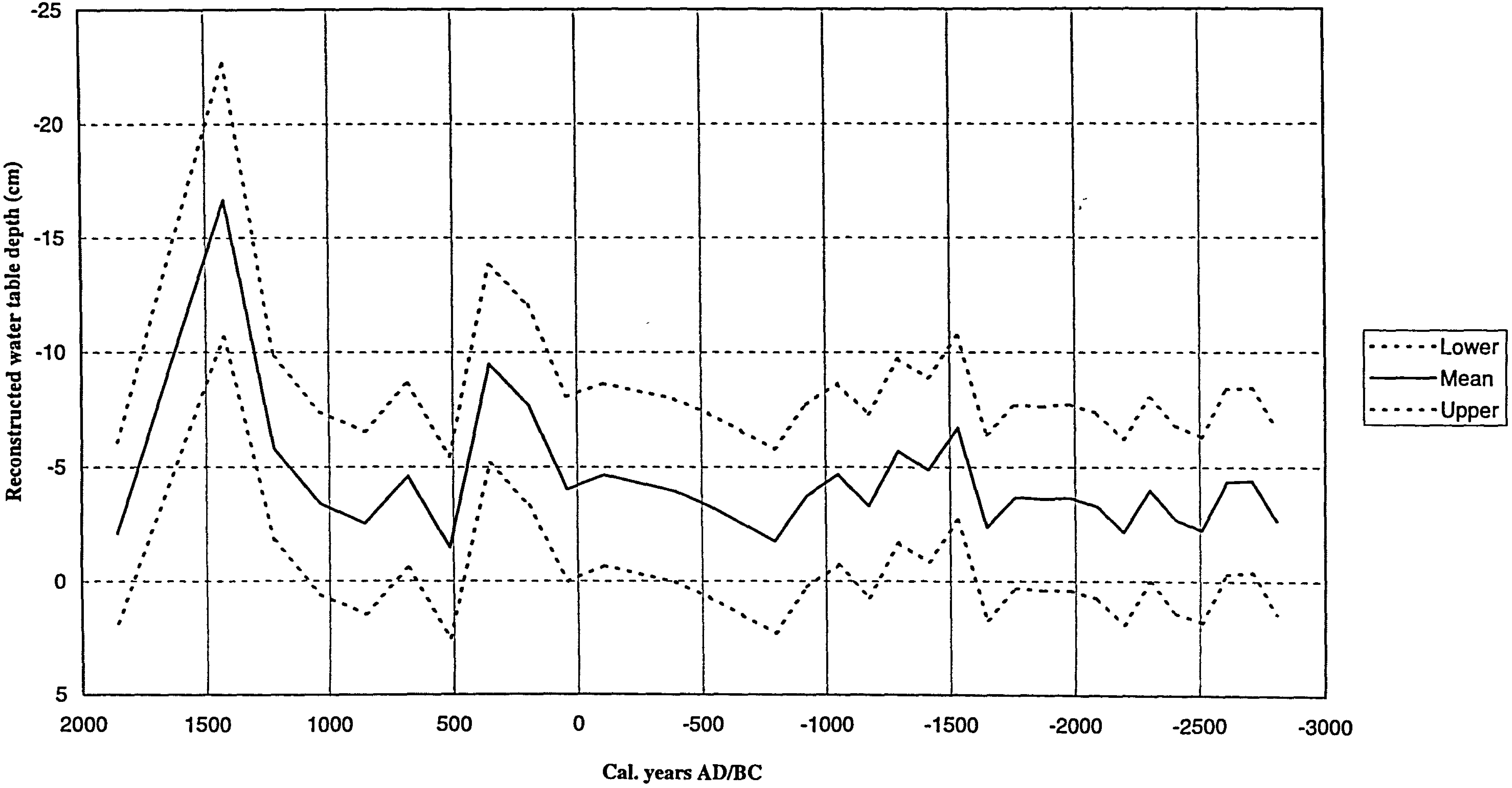
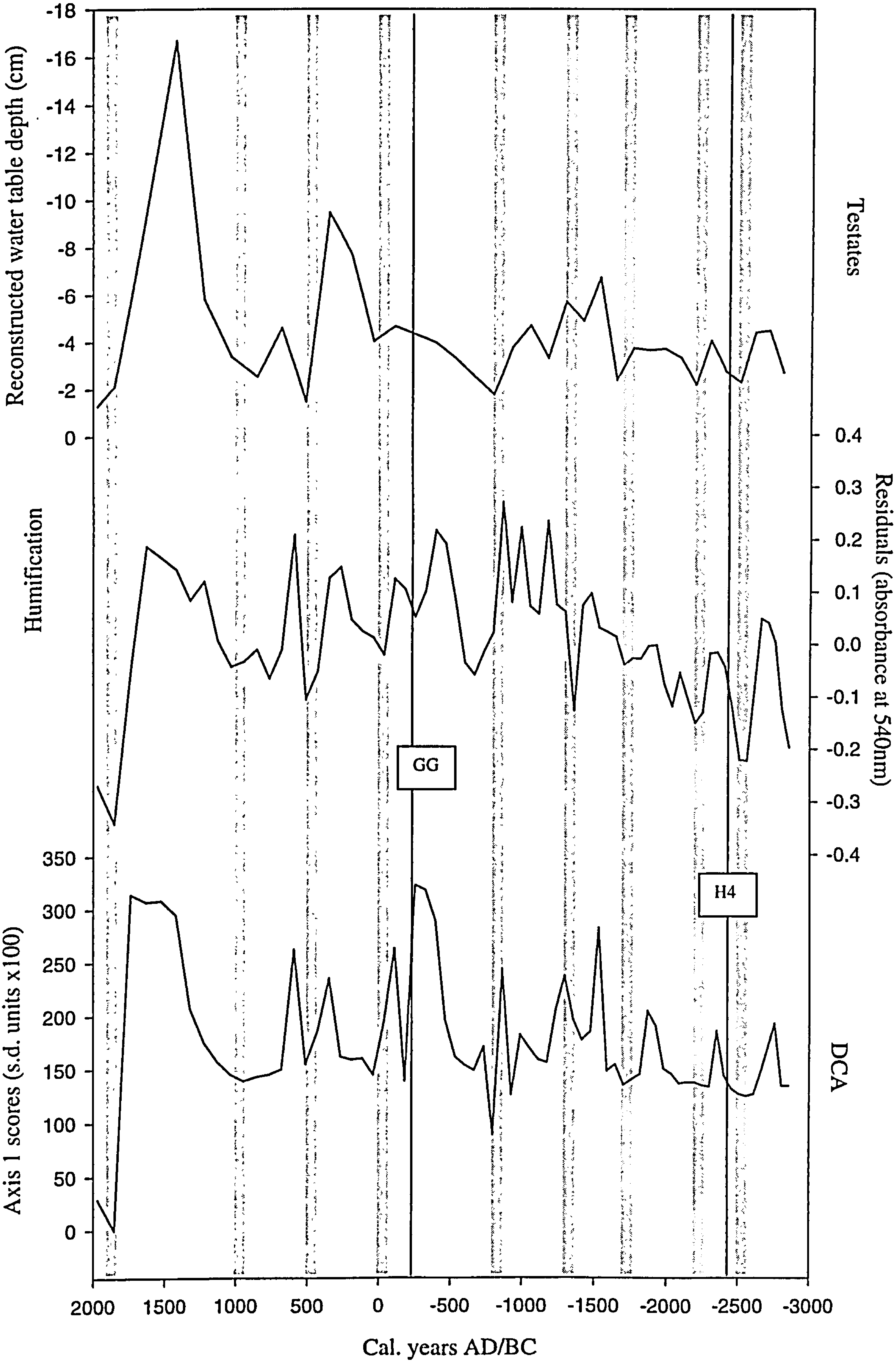


Figure 6.47 Mallachie Moss - DCA scores (plant macrofossils), detrended humification, and testate amoebae data plotted against time. Solid vertical lines indicate tephras.



6.11 Craigmaud Moss

No detailed lithostratigraphical descriptions were undertaken at Craigmaud, as field sampling revealed that the peat was very well humified, with an amorphous appearance, indicating that Troels-Smith (1955) descriptions would be futile at this site. Situated on the northeast coast of Scotland, this site displays the most continental characteristics of any of the sites studied in this project (Birse, 1971). Although the peat is highly decayed, sub-fossil *Sphagnum* remains were identified, although no *Sphagnum imbricatum* was discovered, emphasizing the continentality of the site. Testate amoebae were not counted at Craigmaud.

6.11.1 Plant macrofossils

The plant macrofossil data are displayed in Figure 6.48, and have been divided up into three zones.

Zone CMD-a (152-82cm): UOM levels are generally up to 80% throughout this zone, with the exception of monocotyledon remains (mostly *Eriophorum vaginatum*) found at 152cm, 128cm, 100-96cm, and 84cm. These remains are likely to have survived because they are amongst the species most resistant to decay (Heal *et al.*, 1978), and potentially indicate hummock environments. However, macroscopic charcoal is extremely abundant in every sample within this zone, indicating the mire's susceptibility to fires. The age/depth model (Section 5.3.6) suggests at least one hiatus within this zone, through the repeated burning of the mire, or by the cessation of peat growth, possibly enhanced by low levels of effective precipitation. Kuhry (1994) noted that accumulation rates decrease significantly with increasing fire frequencies.

Zone CMD-b (82-34cm): Low levels of *Sphagnum* section *Cuspidata* are present at 80cm , accompanied by a few leaves of *Drepanocladus fluitans*, indicating wet conditions. A large amount of *Sphagnum* section *Acutifolia* leaves are located between 76-72cm. As these leaves have not been totally decayed they may represent a phase when the aerobic acrotelm was relatively thin, with a comparatively high water table. However, throughout the rest of the zone levels of Sphagna decline, being replaced by monocotyledons, notably *Eriophorum vaginatum* remains, which could indicate that water table depths were lowered, probably in response to a lower regime of effective precipitation.

Zone CMD-c (34-0cm): This zone is characterised by two peaks of *Sphagnum* section *Cuspidata*, accompanied by small peaks in *Sphagnum papillosum*, indicating two separate phases of climatic deterioration. The first occurs at 28cm, and the second at 12cm.

6.11.2 Humification

A pronounced trend in the humification curve can be seen in Figure 6.49, and so the same detrending procedure as at other sites was carried out on the raw data. By applying the age/depth model (Section 5.3.6) to the data a time series of interpolated dates can be developed and plotted against the detrended humification data (Figure 6.50). However, data beyond *ca.* cal. 1000 BC has been discounted because of the possibility of a hiatus. Pronounced peaks of low humification can be seen at *ca.* cal. 740 BC, AD 1410-1530, and *ca.* cal. AD 1720-1780, with less marked fluctuations at *ca.* cal. AD 610 and *ca.* cal. AD 850. More humified peats, suggesting times of less effective precipitation are apparent at *ca.* cal. 440-180 BC, AD 770, AD 1140-1340, and *ca.* cal. AD 1530-1600.

6.11.3 DCA

The DCA plot of axis 1 and axis 2 scores (Figure 6.51) shows a moisture gradient within the macrofossil data, placing the hygrophilous *Sphagnum papillosum* and *Sphagnum* section *Cuspidata* at one end of axis 1, and *Eriophorum vaginatum* and the dry indicator UOM at the other end. Most of the samples are clustered around the UOM component as this tends to dominate the stratigraphy (Figure 6.48). The axis 1 scores are plotted against the interpolated dates from the age/depth model (Figure 6.52). Three major wet phases can be detected in Figure 6.52, occurring around *ca.* cal. AD 610, AD 1470, and *ca.* cal. AD 1720. The lack of major variations between 1500 BC - AD 450 is indicative of the stratigraphy being dominated by UOM with only slight variations in *Eriophorum vaginatum*, and no *Sphagnum* being identifiable. Dry phases can be identified around *ca.* cal. 920 BC, AD 430, AD 1000, AD 1280-1340, AD 1530-1660, and *ca.* cal. AD 1780-1830.

6.11.4 Comparison of the two proxy techniques

Both effective precipitation proxy curves have been plotted together (Figure 6.53) and suggest that four major climatic deteriorations can be detected at Craigmaud. The wet phases can be identified at *ca.* cal. 750 BC, AD 600, AD 1470, and *ca.* cal. AD 1720. Comparatively dry phases can be recognised around *ca.* cal. AD 450, AD 770, AD 1210-1340, and *ca.* cal. AD 1530-1660. It is difficult to identify evidence of any dry phases during the period 1500 BC - AD 450, although there are significant peaks in the humification record around *ca.* cal. 440-180 BC.

Figure 6.48 Craigmaud Moss – Macrofossil diagram

Analysis: Pete Langdon

Figure 6.49 Craigmaud - Humification data. The raw data are presented at the top, and have been detrended by linear regression, producing the lower data set.

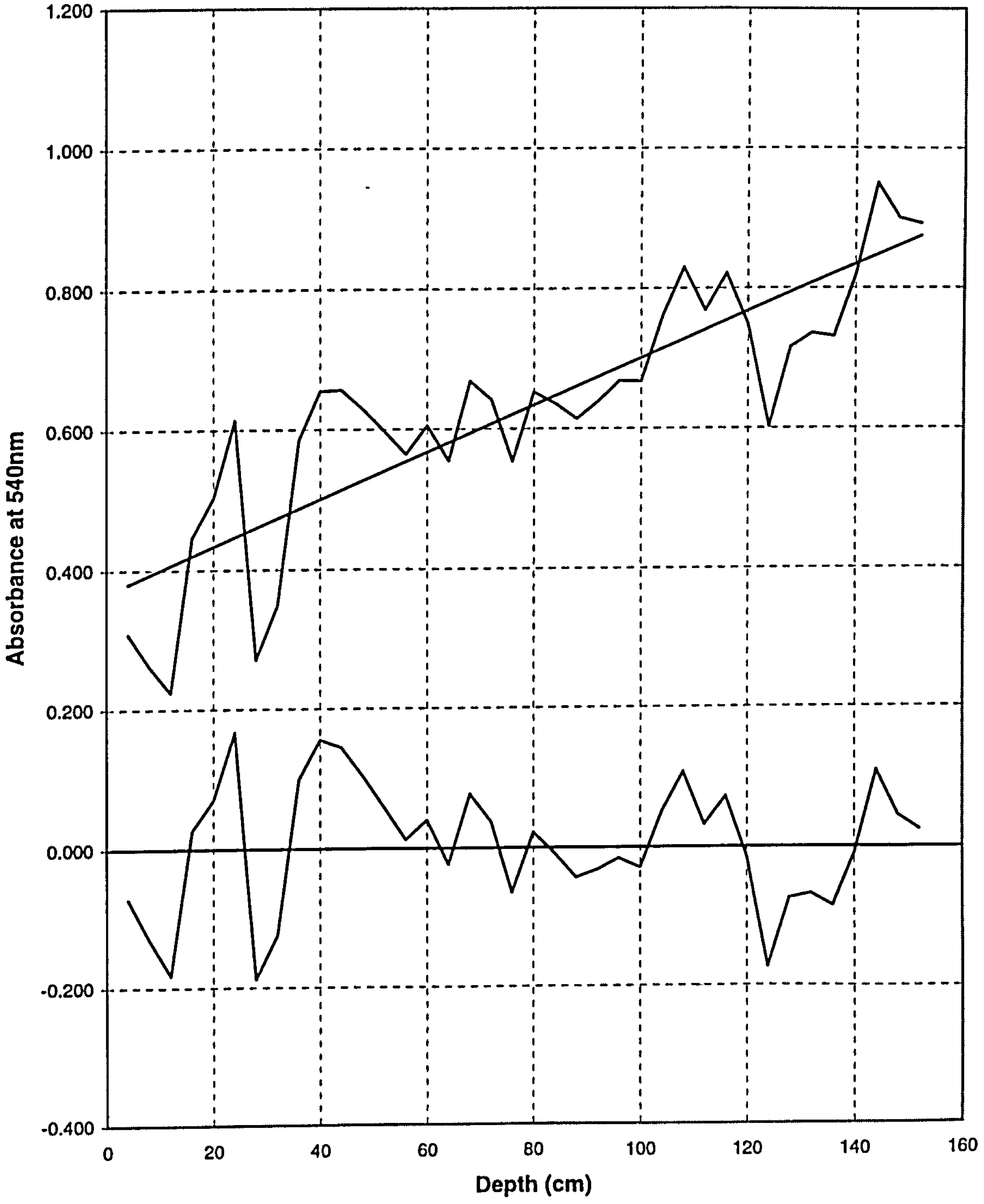


Figure 6.50 Craigmaud Moss - Detrended humification vs time.

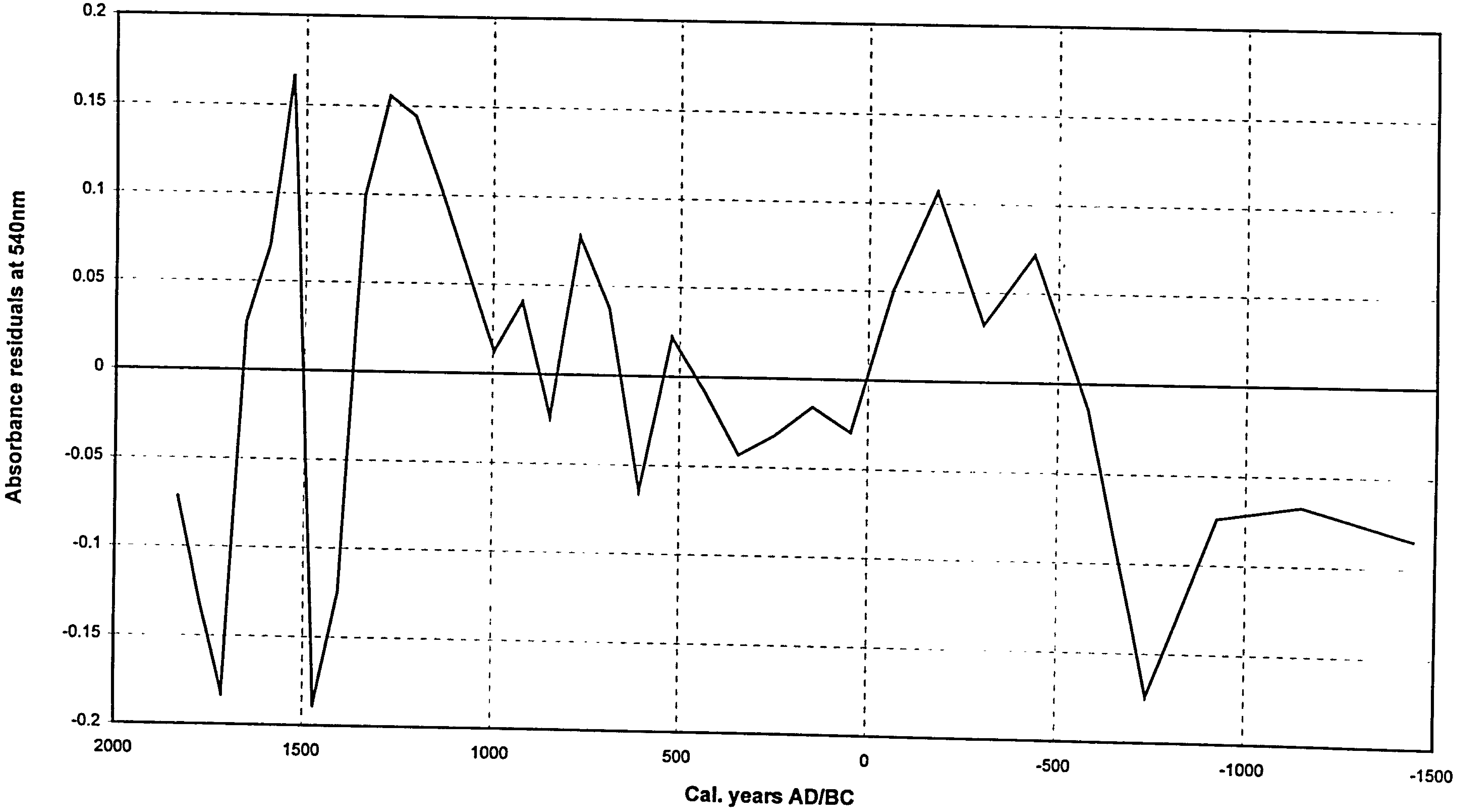


Figure 6.51 Craigmaud Moss - Plant macrofossils: detrending by segments.

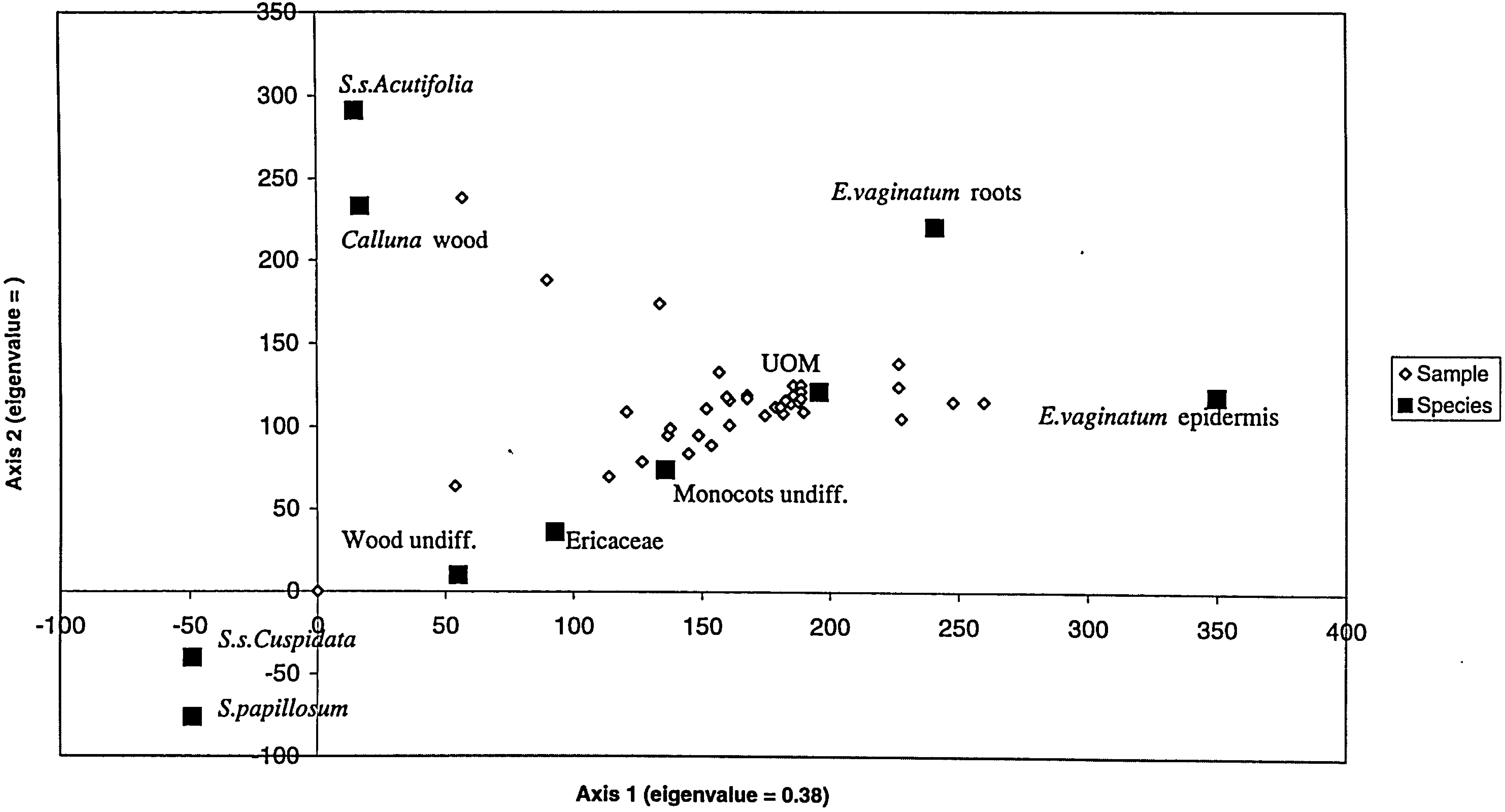


Figure 6.52 Craigmaud Moss - Plant macrofossils DCA: axis 1 scores vs time.

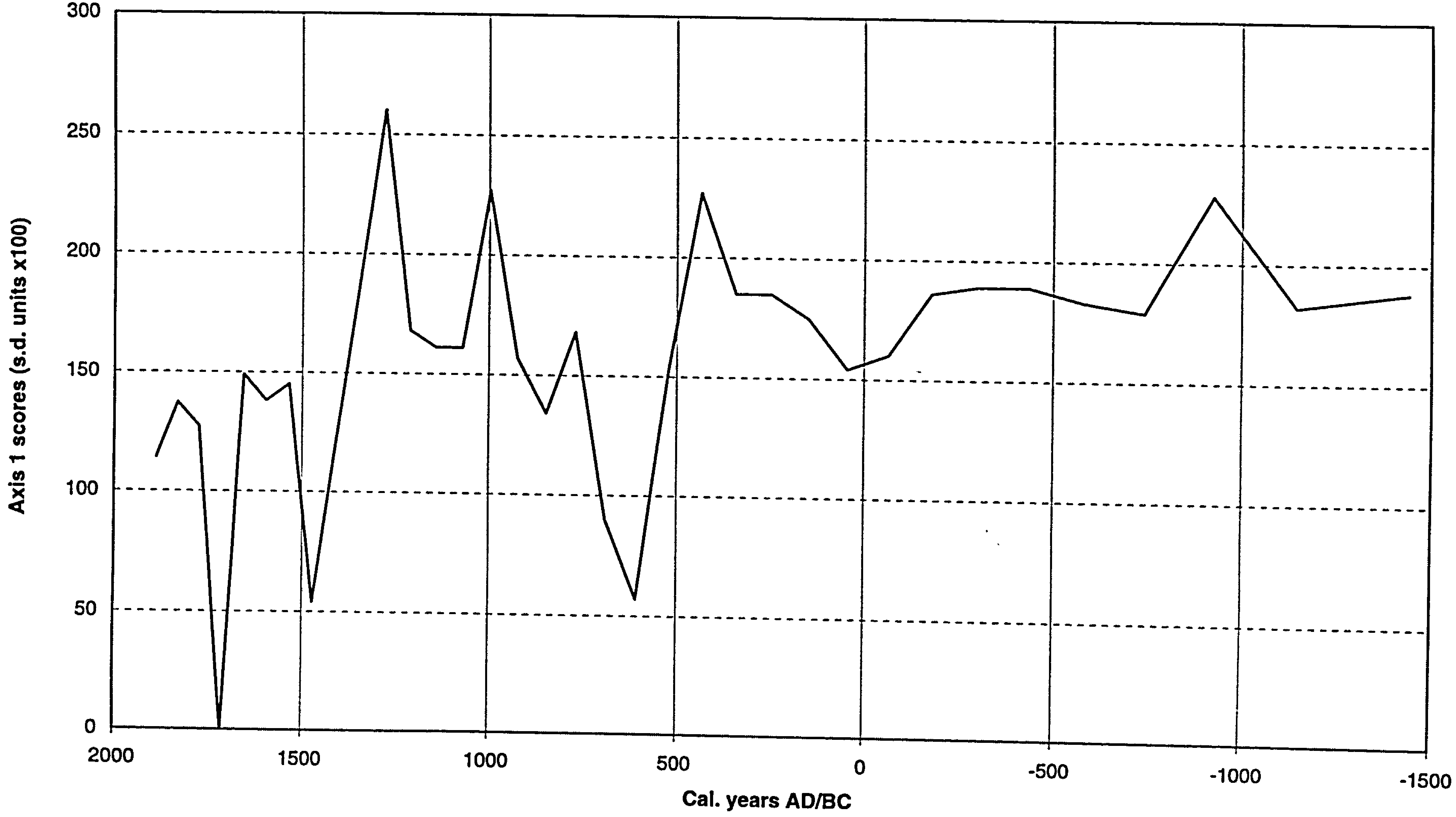
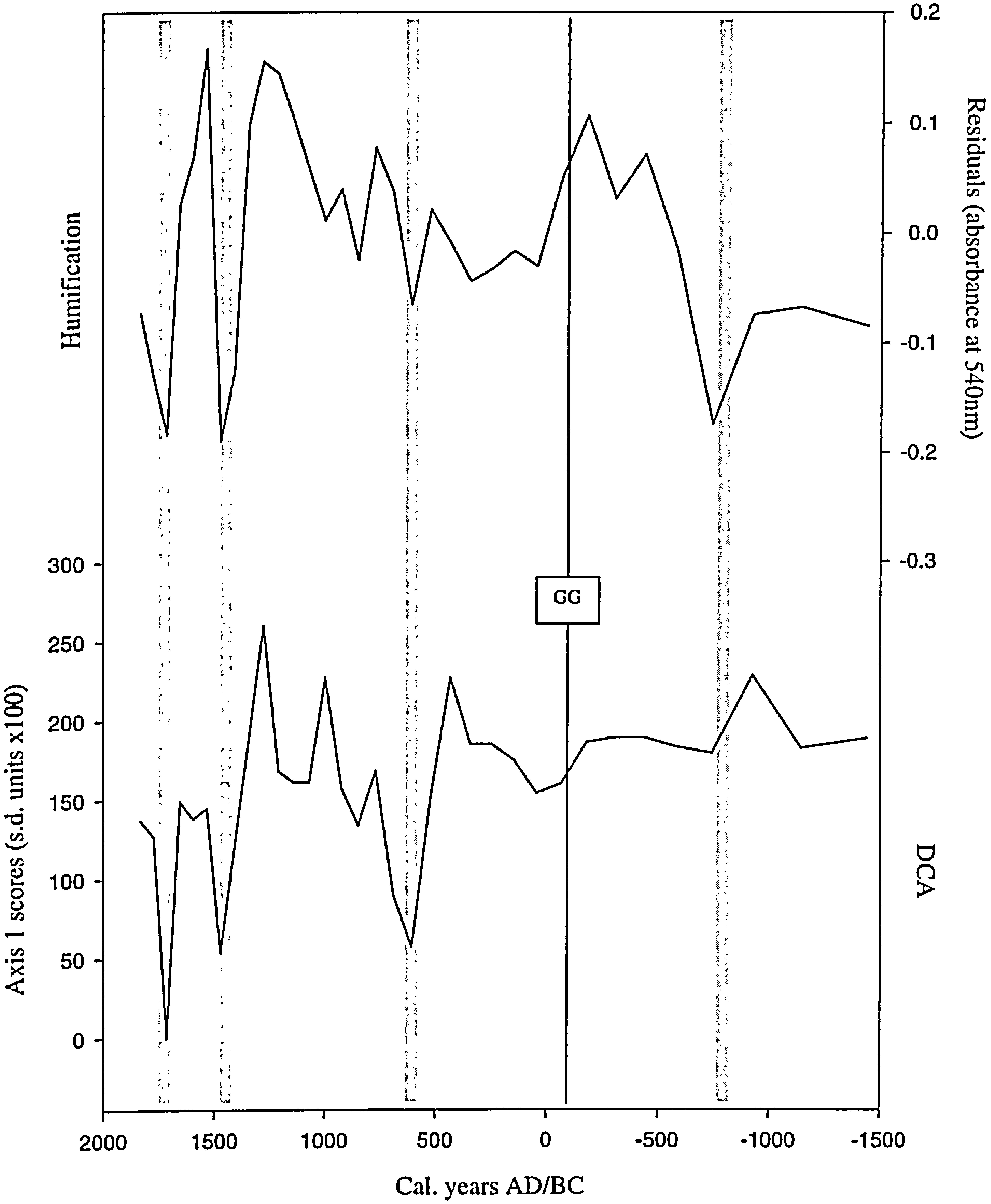


Figure 6.53 Craigmaud Moss - DCA scores (plant macrofossils) and detrended humification plotted against time. Solid vertical line is a tephra.



6.12 Ben Gorm Moss

This blanket mire had only a poor preservation of plant macrofossils (Figure 6.54) , so counts were based on assigning overall peat components on a 1-5 basis from trough estimates, so that the total components came to 5. Then, each component which was present was assigned an abundance on a 1-5 scale, where 1=rare, 2=occasional, 3=frequent, 4=common, 5=abundant. Testate amoebae analysis was undertaken, but concentrations were very low. Below 40cm depth the fauna was dominated by *Hyalosphenia subflava* to the extent that few other species were present, possibly due to selective test decay, so the analysis was curtailed.

The poor preservation of macrofossils, specifically a lack of hygrophilous subfossil remains does not easily allow climatic reconstructions to be undertaken. Hence, without any testate data, the climatic inferences are heavily reliant on the humification data (Blackford and Chambers, 1991; 1995).

6.12.1 Plant macrofossils

Zone BGM-a (138-79cm): The peat is mainly dominated by UOM, monocotylodons, and Ericaceae. Comparatively high levels of *Polytrichum juniperus/alpestre* type are present apart from between 108-104cm where *Eriophorum vaginatum* became dominant. Macroscopic charcoal levels are high in the bottom 20cm suggesting that fires may have been common.

Zone BGM-b (79-25cm): Much more *Eriophorum vaginatum* is present within this zone, although much less Ericaceae are found, possibly being unable to compete with the *Eriophorum.*

Zone BGM-c (25-0cm): Some *Sphagnum* leaves are identifiable, notably *Sphagnum papillosum* between 22-16cm, indicating a relatively high water table, and section *Acutifolia* from 4-8cm. Much less charcoal is present in this zone.

6.12.2 Humification

A distinct trend can be seen in the humification curve (Figure 6.55), and it has thus been detrended according to the previous examples. Using the interpolated dates from the age/depth model a time series has been created and plotted against the detrended humification data (Figure 6.56). Pronounced peaks of low humification, implying a climatic deterioration, are found around *ca.* cal. 1870-1810 BC, 1630 BC (both of which are within the range of the wet shift around 3500-3900 BP identified on the west coast of Scotland by Anderson *et al.*, 1998), 1390 BC, 900 BC, 660 BC, AD 1090-1150, AD 1570, and *ca.* cal. AD 1810. Peats of high humification are identified at *ca.* cal. 1990 BC, 1690 BC, 780 BC, 540 BC, 420 BC - AD 10, AD 600-840, and *ca.* cal. AD 1270.

6.12.3 Comparison of the two analyses

The climatic deteriorations identified around *ca.* cal. 1870-1810 BC, 1630 BC, 1390 BC, 900 BC and *ca.* cal. 660 BC correspond with very low amounts of *Eriophorum vaginatum* on the mire. However, the macrofossil record does not really aid the identification of other shifts in climate, due to the poor preservation of Sphagna in the deeper peats. Thus, humification is relied upon for the major climatic inferences based on the Ben Gorm Moss sequence.

Figure 6.54 Ben Gorm Moss – Macrofossil diagram

Analysis: Pete Langdon

Figure 6.55 Ben Gorm Moss - Humification data. The raw data are presented at the top, and have been detrended by linear regression, producing the lower data set.

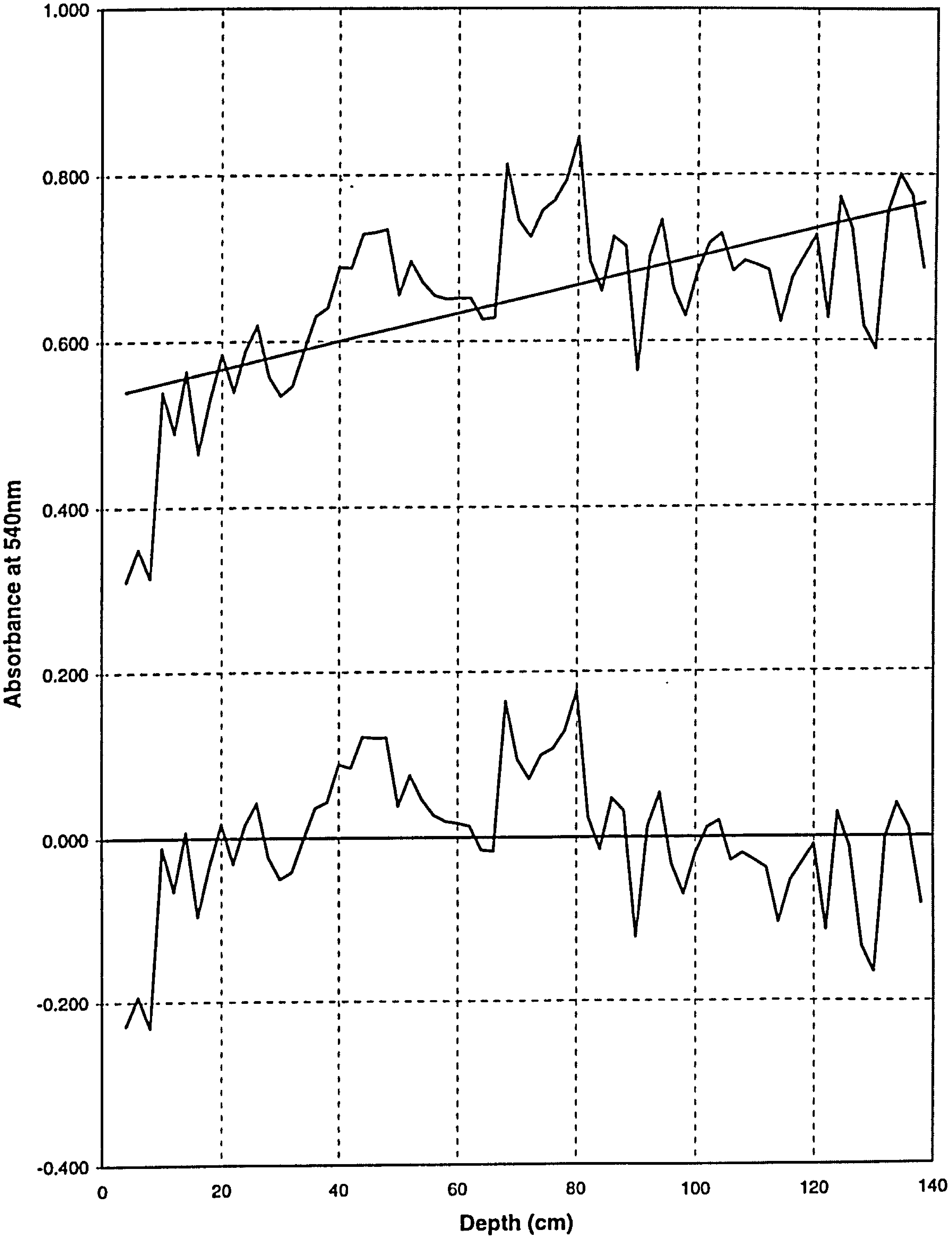
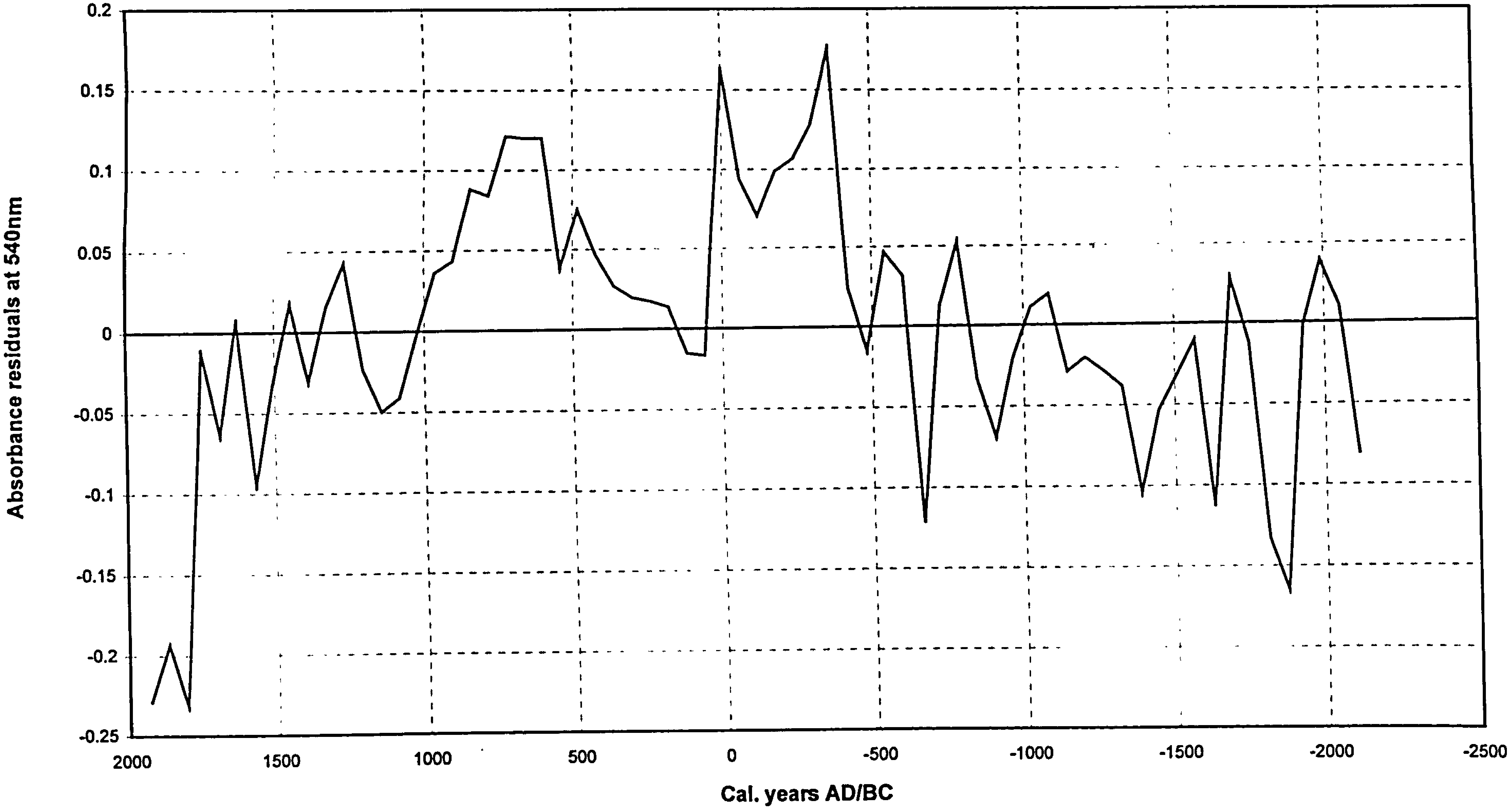


Figure 6.56 Ben Gorm Moss - Detrended humification vs time.



6.13 Time series analysis

Time series analysis of climate related parameters plays a fundamental role in climatic research. The search for a signal of systematic behaviour within a data set, which can be distinguished from the noise (variations which would occur if the data set was purely random) should provide an insight into the underlying mechanisms which govern the physical processes influencing the data. The strength of different variations on different time scales are important characteristics of any time series, and may reveal long term trends within the data, or disclose whether a significant cycle exists at a particular period.

Before submitting the data to time series analysis, the data must be checked for long term trends, which could result in spurious periodicities being produced by the spectral analyses. The humification data have already been detrended due to the slow, but consistent, long-term decay of peat at depths (Clymo, 1984). No visible trend is apparent in the plant macrofossil DCA data with the possible exception of the data from Shirgarton Moss and Craigmaud Moss. These two data sets were thus detrended for the purposes of time series analysis.

Spectral analysis has been used in order to try and detect signals within the time series. Spectral analysis plots the data in the frequency or wavelength domain showing how much variance (and therefore variability) is explained by the different frequencies or wavelength bands. Breaking down this variance enables systematic behaviour to be identified as periodicities, and will aid the understanding of why variations are occurring.

The age/depth models constructed for each core do not provide equal spacing of age intervals, due to slight variations in accumulation rates. Traditional spectral analysis methods, e.g. Fast Fourier Transform, require evenly spaced data so an approach to transforming the data must be considered. Different ways in which to solve this problem exist, the simplest way of which is to interpolate the data by constructing a grid of evenly spaced times, and overlaying it on top of the data, interpolating the values onto the grid. The Fast Fourier Transform can then be used for spectral estimation.

However, it has been noted that this method can behave poorly, especially if there are long gaps in the data, when a spurious bulge of power at low frequencies can be produced (Press *et al.*, 1992).

A different method of spectral analysis for unevenly sampled data was therefore used, in order to mitigate the problems of interpolations. This method was developed by Lomb (1976), based on earlier work by Barning (1963) and Vanicek (1971), which has been expanded by Scargle (1982). The Lomb method evaluates data on the sinusoidal wavelengths at the actual sampled times, forming a normalised periodogram. The data is weighted on a 'per point' basis rather than a 'per time interval' basis (see Press *et al.*, 1992 for details).

No confidence bands have been placed on the periodograms, as the Lomb program used was unable to calculate these, so the results must be viewed only as a guide as to the amount of variance each peak actually shows, and which peaks are statistically significant. Analyses were carried out on the Lomb program (in Press *et al.*, 1992), and have been plotted using Doodler (kindly made available by M. Kelly, Climatic Research Unit, University of East Anglia).

The plant macrofossil DCA reconstructions and humification records from each site were subjected to spectral analysis. The reconstructed water table depths from the testate amoebae data were not analysed, due to the low sampling resolution of the data. As sampling intervals of 4cm were used for the effective precipitation proxy records, this relates to an average age of *ca.* 40-70 years depending on the site, and *ca.* 60 years for Ben Gorm Moss as it was sampled every 2cm. At least two sampling intervals are considered to be necessary for a periodicity to be considered significant (T. Payne, pers. comm.), and thus in general periodicities of under 150 years were not considered. Conversely, at least 7-10 cycles of any periodicity need to be present in a data set to be considered statistically significant (M. Kelly, pers. comm.), and thus where a record extends for *ca.* 4000 years (most of the sites) only periodicities between *ca.* 400-150 years can be considered statistically significant. Many peaks around the 70-80 year cyclicity, which correspond with the well known solar variations, are produced by the periodograms from the sites in this study, although they cannot be suggested to be significant, due to the relatively low sampling resolution present in this study. However, it is suggested that with a higher sampling resolution these periods could prove significant.

Longbridge Moss

Spectral analyses from the plant macrofossil DCA axis 1 scores showed a large amount of low frequency variance, none of which is significant due to the core only being *ca.* 4700 years old. However, a potentially significant peak in variance does exists with a periodicity of 520 years (Figure 6.57), which matches the double peak identified by Aaby (1976). Analysis of the humification data shows similar amounts of low frequency cycles, with a notable period of 1250 years (Figure 6.58). However, this cycle is only replicated by the data set four times, and may not therefore be significant. More impressively, a potentially significant cycle of 470 years can be seen in the humification record (Figure 6.58), which corresponds well with the 520 year cycle identified in the DCA analysis.

Langlands Moss

A periodogram of the DCA data can be seen in Figure 6.59. A low frequency peak, although not statistically significant, can be identified with a period of 910 years, but cycles which do appear to be significant can be recognised as having periods of 580 and 316 years. The humification data has also been subjected to spectral analysis (Figure 6.60) and periodicities can be recognised at 575, 370, and 315 years, being remarkably coherent with the DCA data. There is also a noticeable peak at the 83 year cycle, although not statistically significant. It would therefore appear that Langlands Moss exhibits two significant periodicities, replicated from two separate analyses, at 580 year and 315 year cycles. The possible mechanisms and interpretation of such cycles are discussed in Chapter 7.

Temple Hill Moss

Temple Hill Moss proves to have some of the most significant periodicities of any of the sites. Spectral analyses of both the DCA (Figure 6.61) and humification (Figure 6.62) records reveals a clearly significant periodicity of 1100 years. The record at Temple Hill Moss extends back for *ca.* 7500 years, suggesting that this cycle can be replicated to a significant level. The cycle can also be identified in the wet shifts shown in Figure 6.29. This cycle is consistent with the 1150 year periodicity found at Walton Moss (Hughes *et al.*, 1999/2000), and the cycles identified by Bond *et al.*, (1997) as a mechanism for millennial scale climatic change, possibly linked to the North Atlantic's thermohaline circulation (Bond *et al.*, 1997). The implications for this signal being detected in one of the most eastern, and comparatively least oceanic site is discussed in Chapter 7, along with the nature of the cyclicity.

Shirgarton Moss

Spectral analysis of the DCA data reveals a low peak, probably not significant, with a periodicity of 745 years. The humification data identifies a low peak displaying a cyclicity of 465 years. Thus, the two records are not coherent, and suggest that neither peak is significant, hence neither periodogram is shown. The detection of a major periodicity with the data is probably impeded by the almost total dominance of *Sphagnum imbricatum* throughout the stratigraphy, thus damping down more subtle variations.

Mallachie Moss

Analysis of the DCA data reveals a significant periodicity at 920 years (Figure 6.63), although this cycle is beyond the limits of the core age, which is only *ca.* 4900 years, suggesting only a maximum of five cycles can have occurred in this time. However, there are also prominent peaks at the 120 and 100 year cycles, although these are barely significant when taking into account the relatively low sampling resolution. The humification data reveals much less low frequency variance, but a potentially significant peak at the 365 year cycle (Figure 6.64). However, the two records do not agree on a common periodicity, and thus none can be accepted for this site. A tentative explanation could be the amount of past disturbance to the site (as indicated by relatively high levels of charcoal - see section 6.10.2), hence affecting the accumulation rates of the mire.

Craigmaud Moss

A potentially significant peak at the 280 year cycle was identified in the DCA record (Figure 6.65), and three less prominent, but possibly significant, peaks at the 400, 350, and 280 year cycles can be recognised in the humification data (Figure 6.66). The significance of these results can be overstated, but it does appear that a 280 year cycle can be traced in the records throughout the last 2000 years, as before this time it appears that a hiatus may be present in the stratigraphy (see section 6.11.1).

Ben Gorm Moss

No real structure was observed in the humification data from Ben Gorm Moss.

6.14 Summary

The plant macrofossil, humification, and testate amoebae reconstructions have enabled variations in water table depths, and hence climatic changes to be detected on each site studied. The use of testate amoebae and humification analyses to complement the plant macrofossil data has been vital in detailing stratigraphic changes where one plant community dominates. There appears to be only slight leads and lags between the proxy records, as most seem to agree on a definite phase of wetter/drier climates, suggesting there is little smearing of the records (*sensu* Baillie, 1991) which therefore provide coherent phases of change.

·The tephra isochrones provide additional support for the chronology of each site, and enable precise correlations between proxy records to be made, which will be explored fully in the following chapter. Chapter 7 will also compare the timing and cyclicity of climatic changes at each site, and attempt to synthesise changes over Scotland for the last *ca.* 5000years.

Figure 6.57 Periodogram showing plant macrofossil DCA values from Longbridge Moss over the last ca. 4700 years.

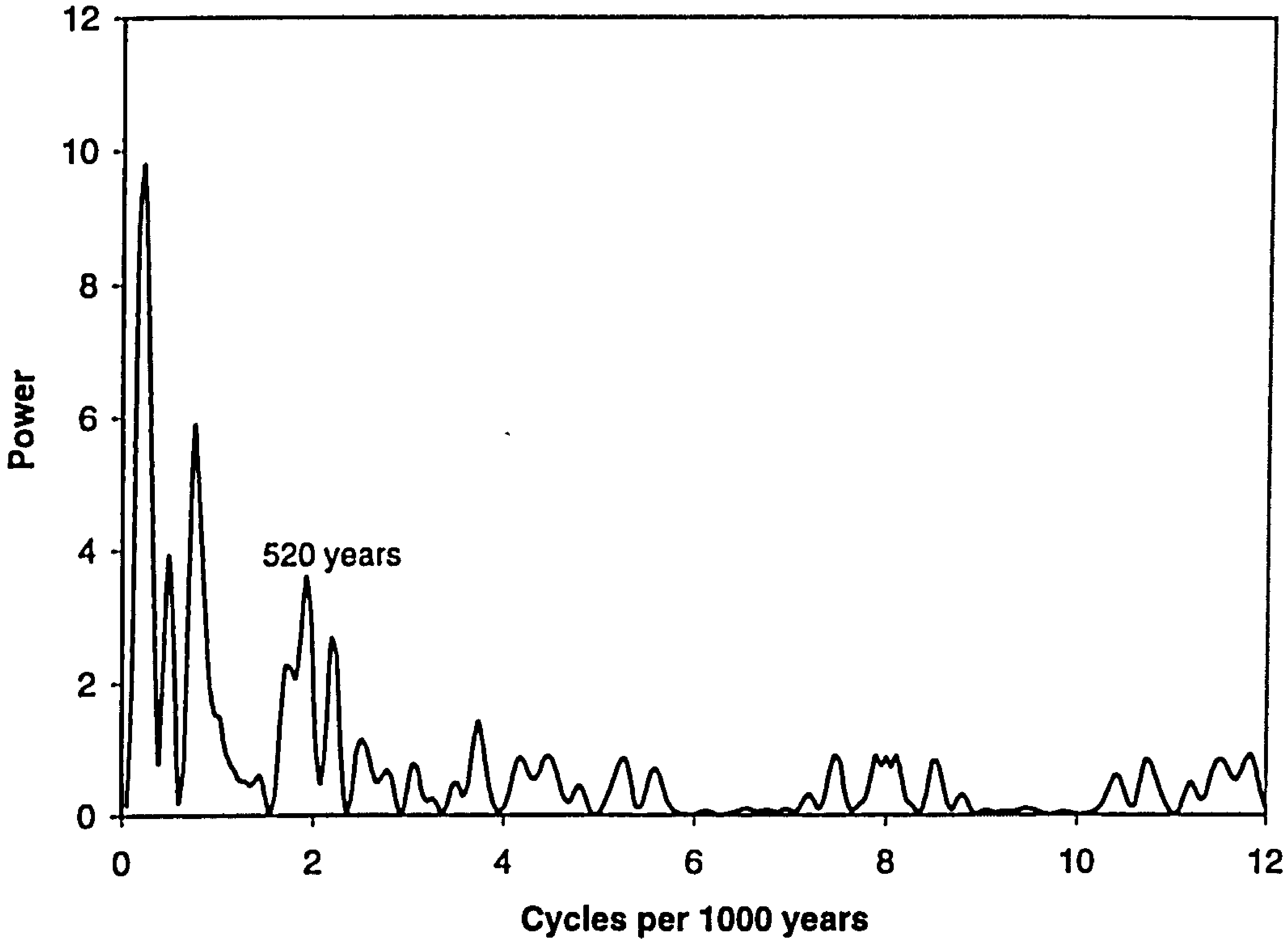


Figure 6.58 Periodogram showing humification values from Longbridge Moss over the last ca. 4700 years.

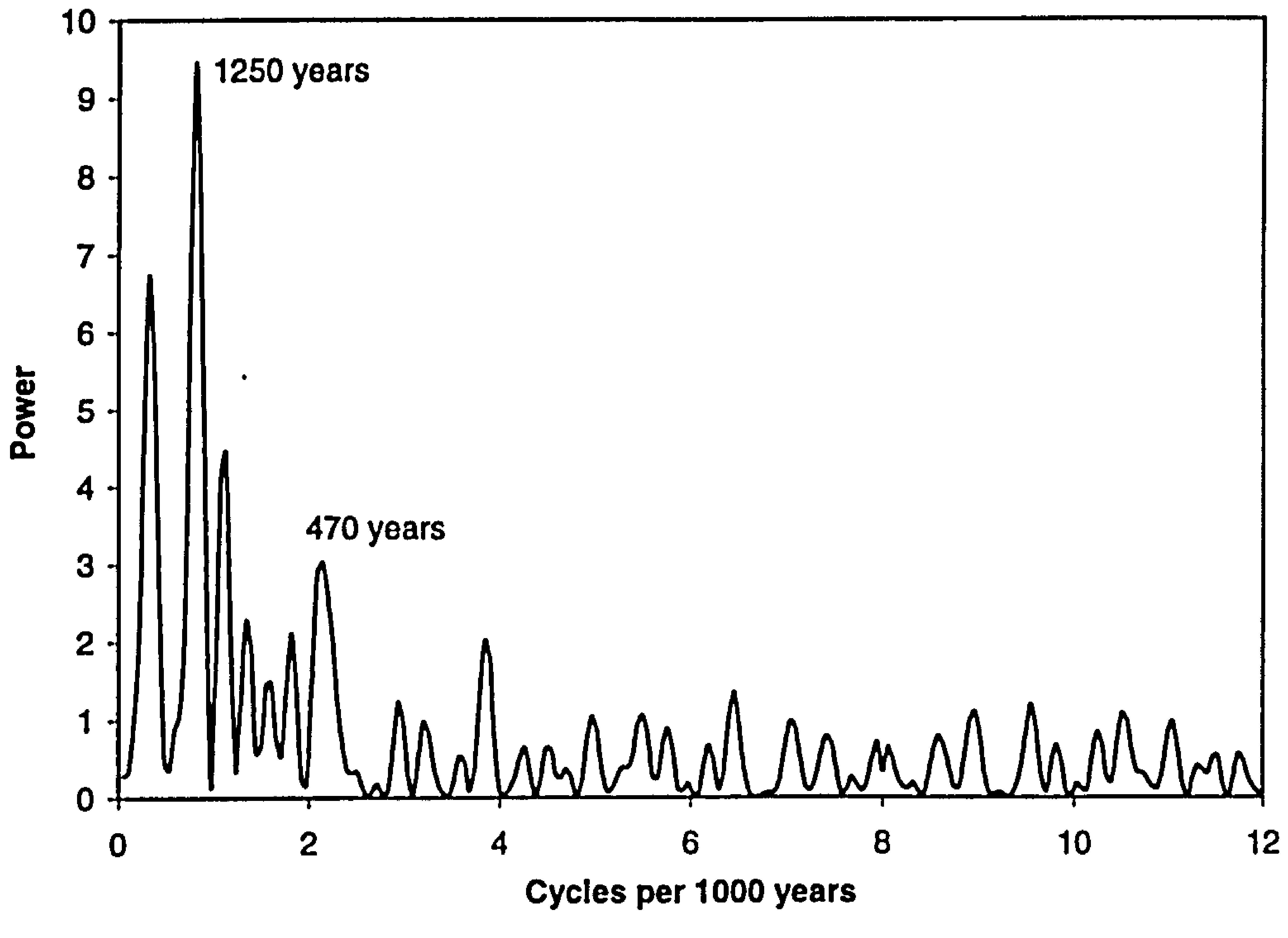


Figure 6.59 Periodogram showing plant macrofossil DCA values from Langlands Moss over the last ca. 4000 years.

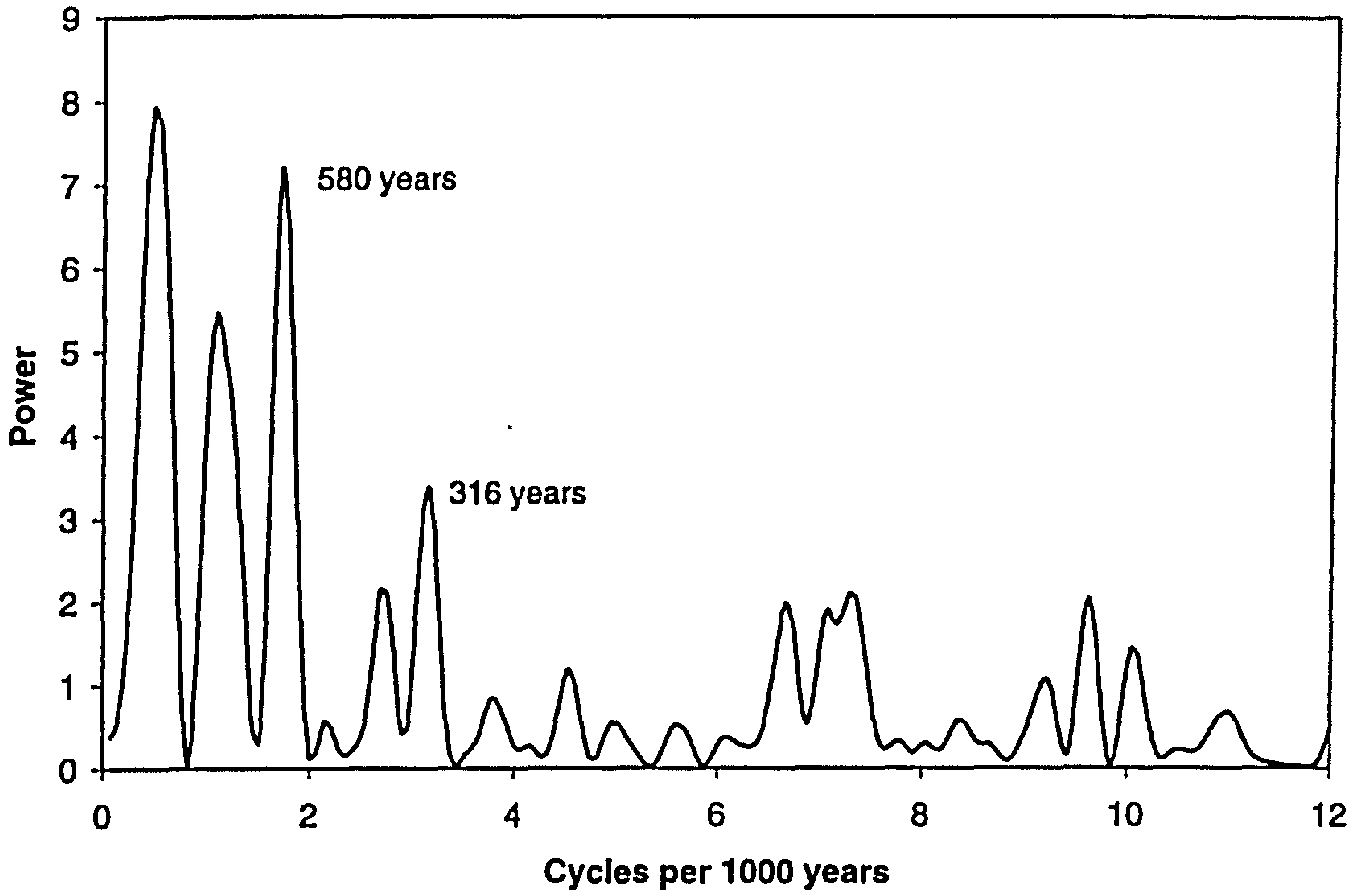


Figure 6.60 Periodogram showing humification values from Langlands Moss over the last ca. 4000 years.

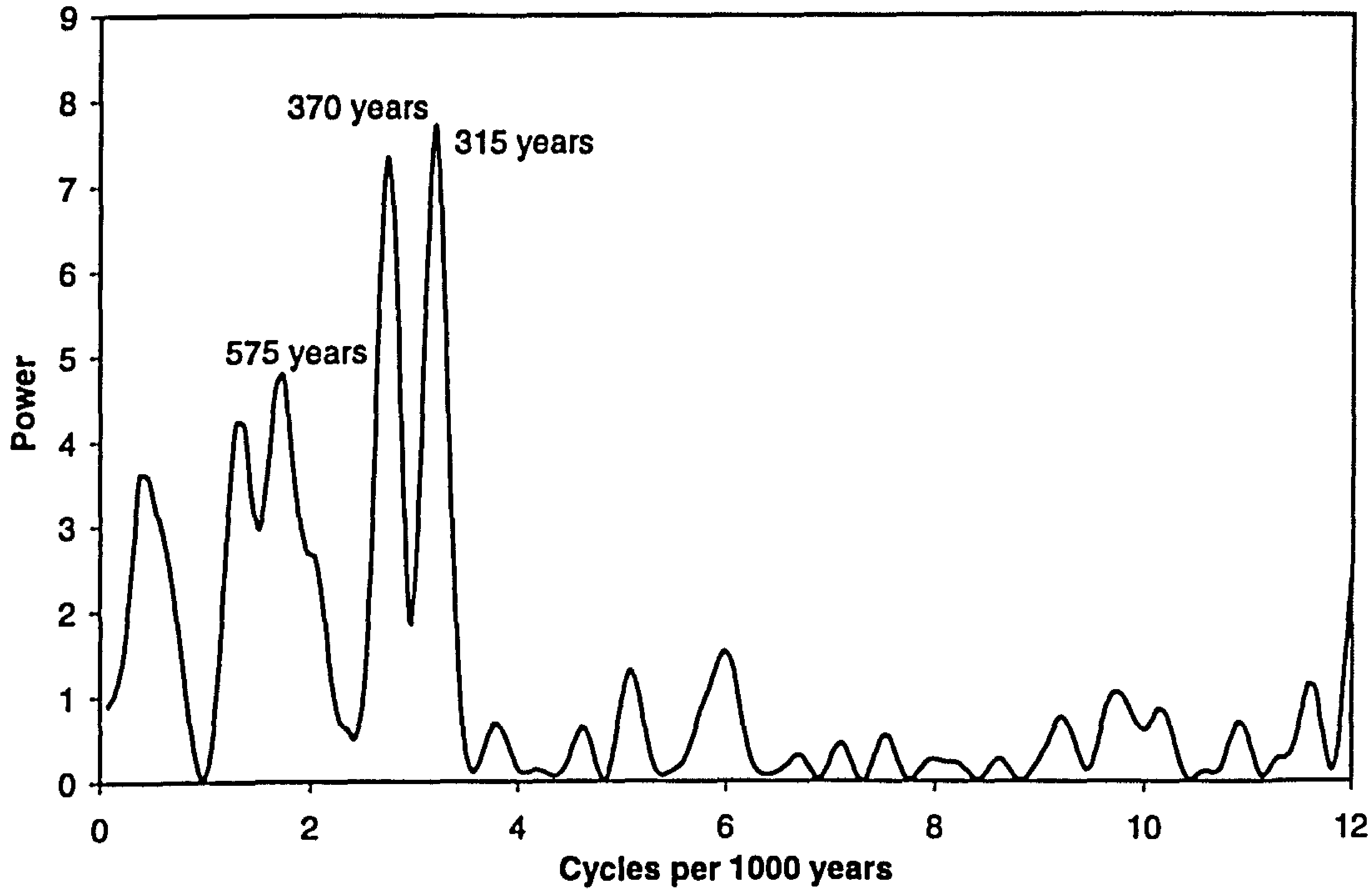


Figure 6.61 Periodogram showing plant macrofossil DCA values from Temple Hill Moss over the last ca. 7500 years.

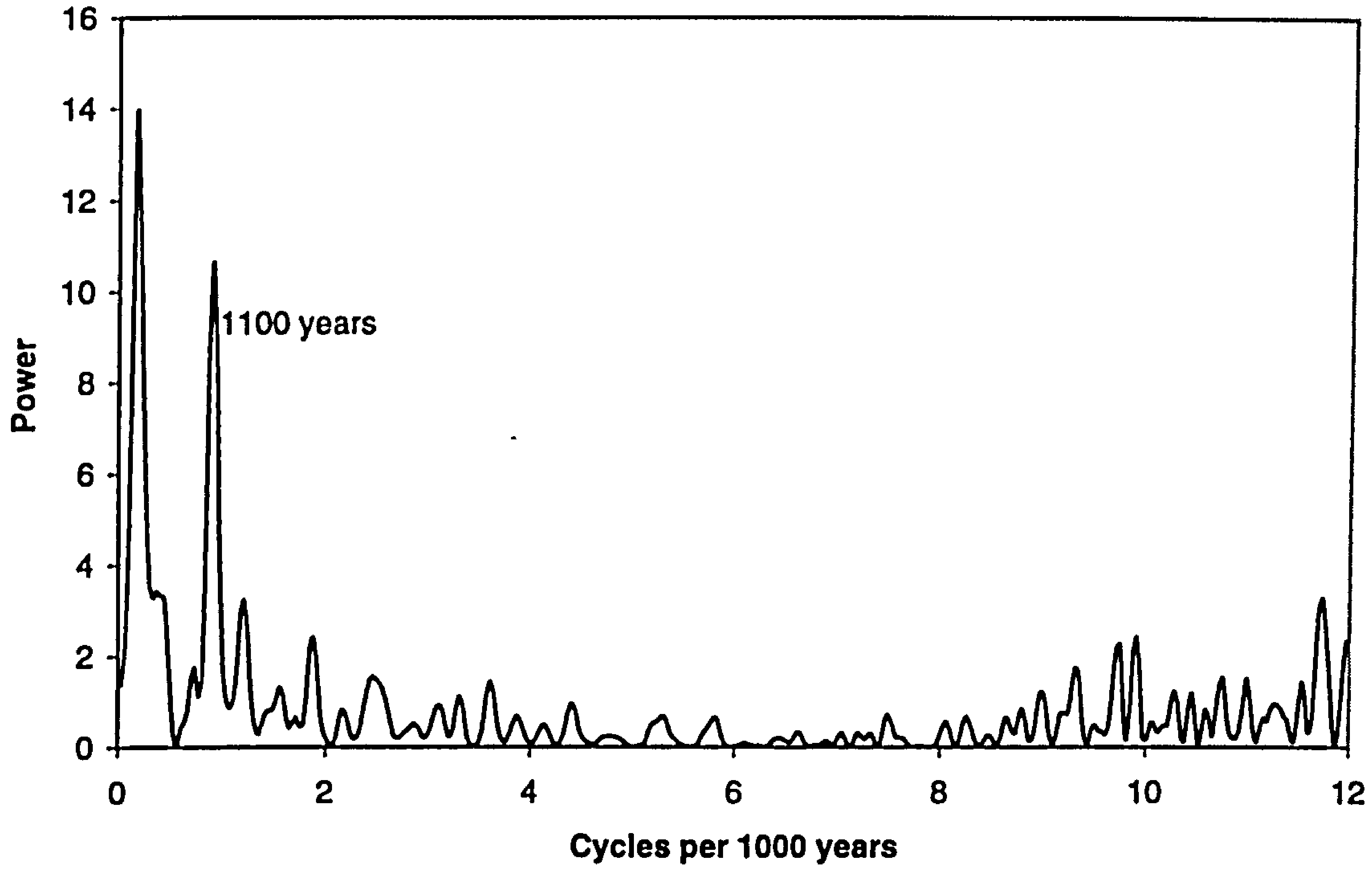


Figure 6.62 Periodogram showing humification values from Temple Hill Moss over the last ca. 7500 years.

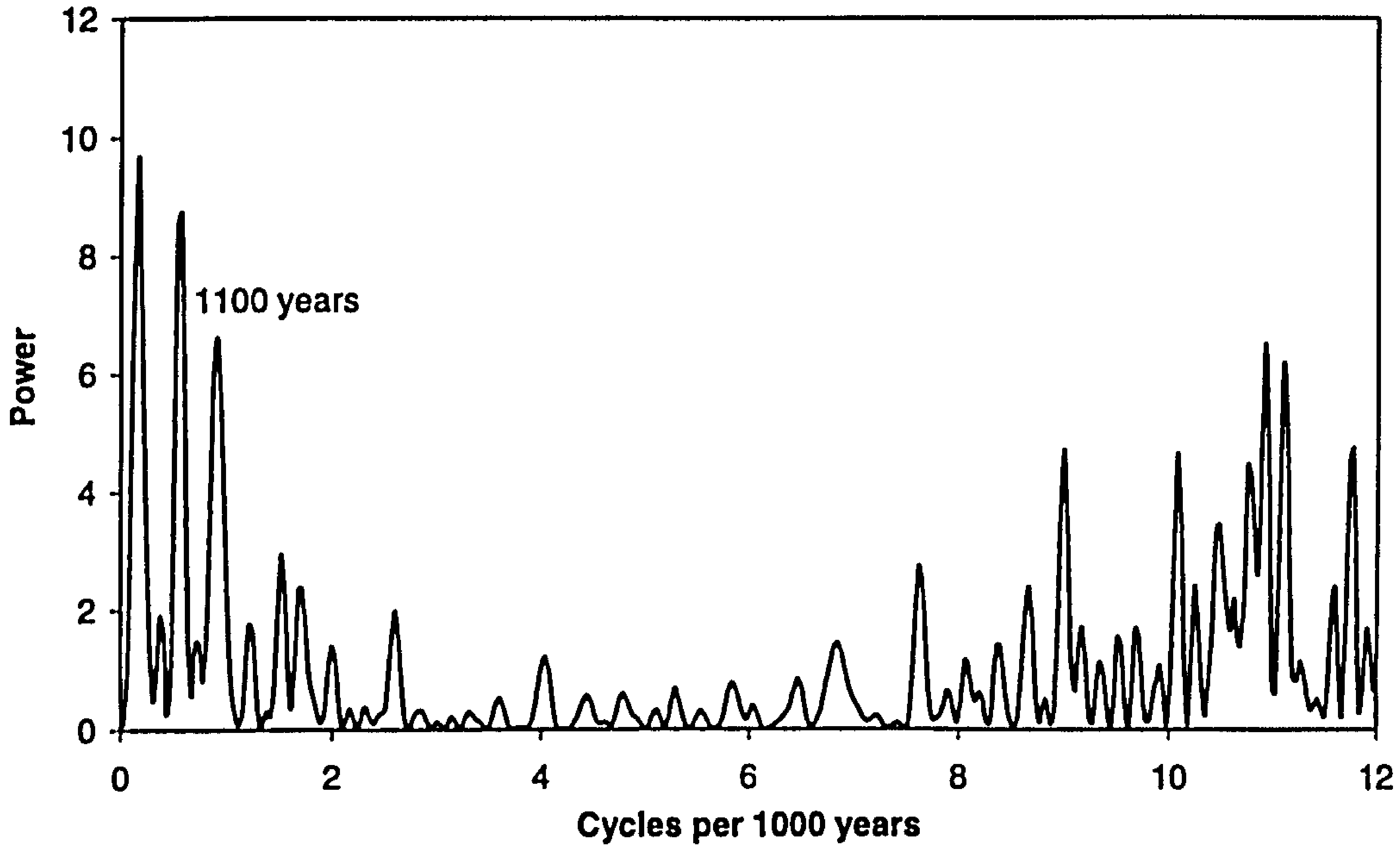


Figure 6.63 Periodogram showing plant macrofossil DCA values from Mallachie Moss over the last ca. 4900 years.

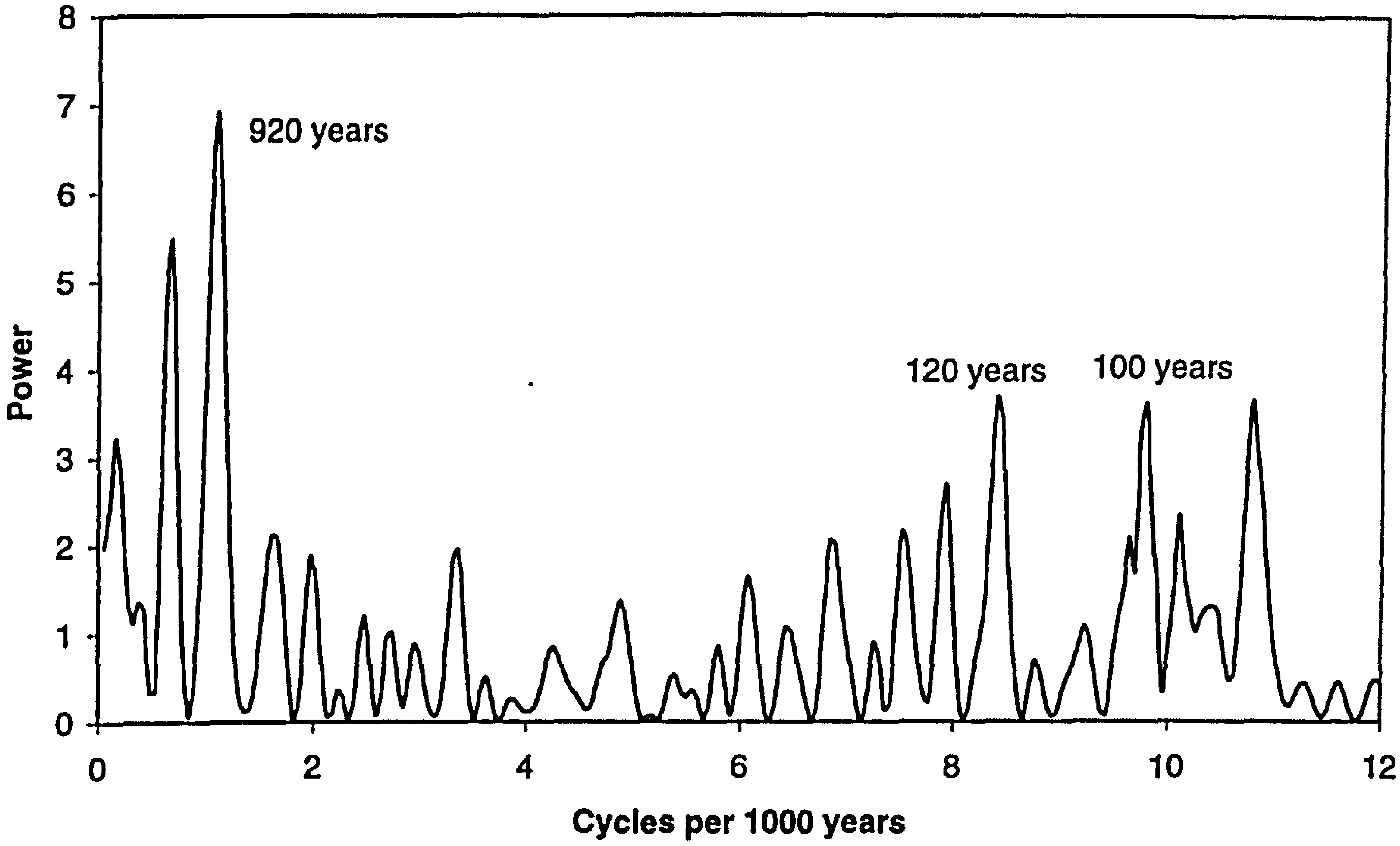


Figure 6.64 Periodogram showing humification values from Mallachie Moss for the last ca. 4900 years.

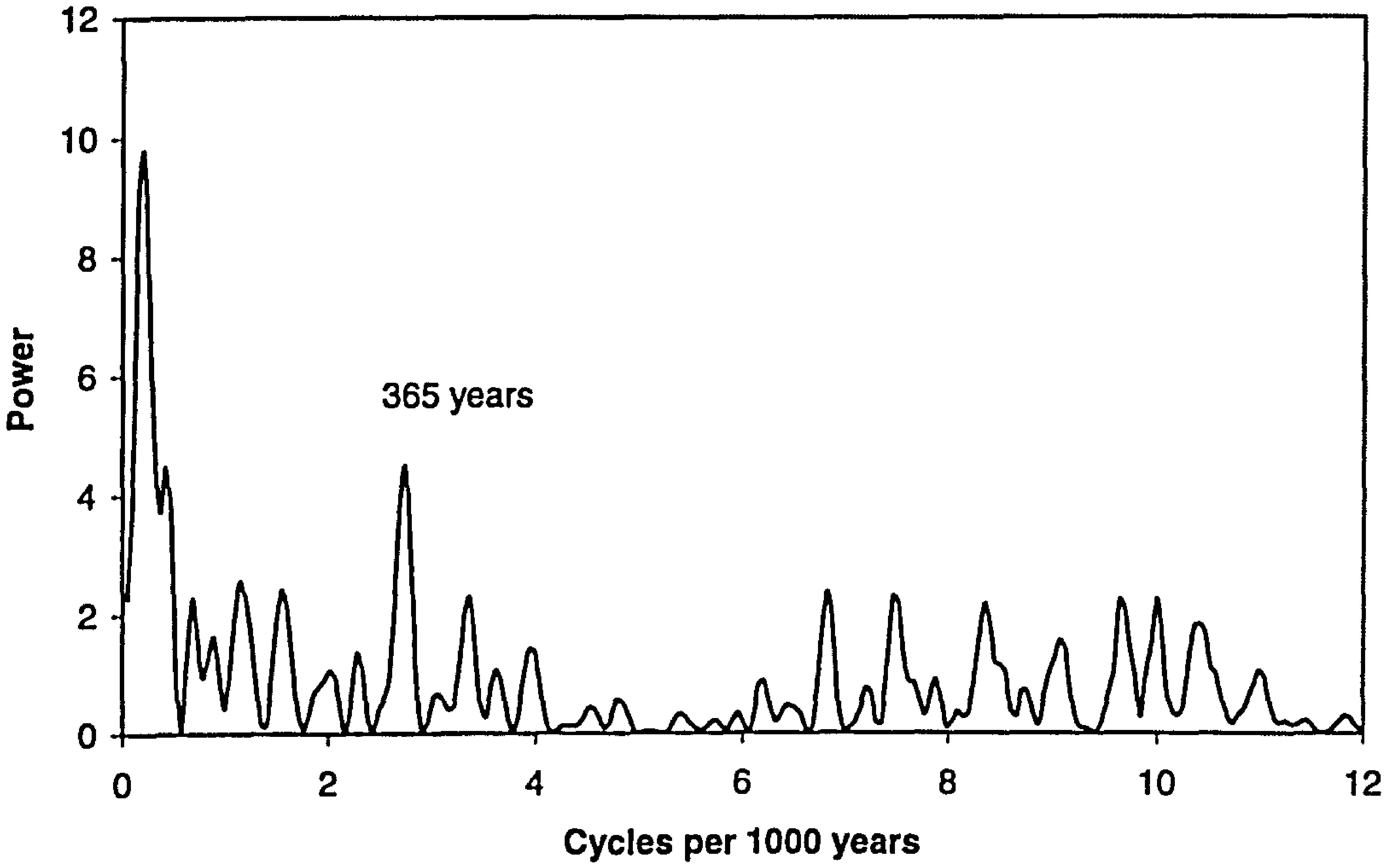


Figure 6.65 Periodogram showing plant macrofossil DCA values from Craigmaud Moss over the last ca. 2000 years.

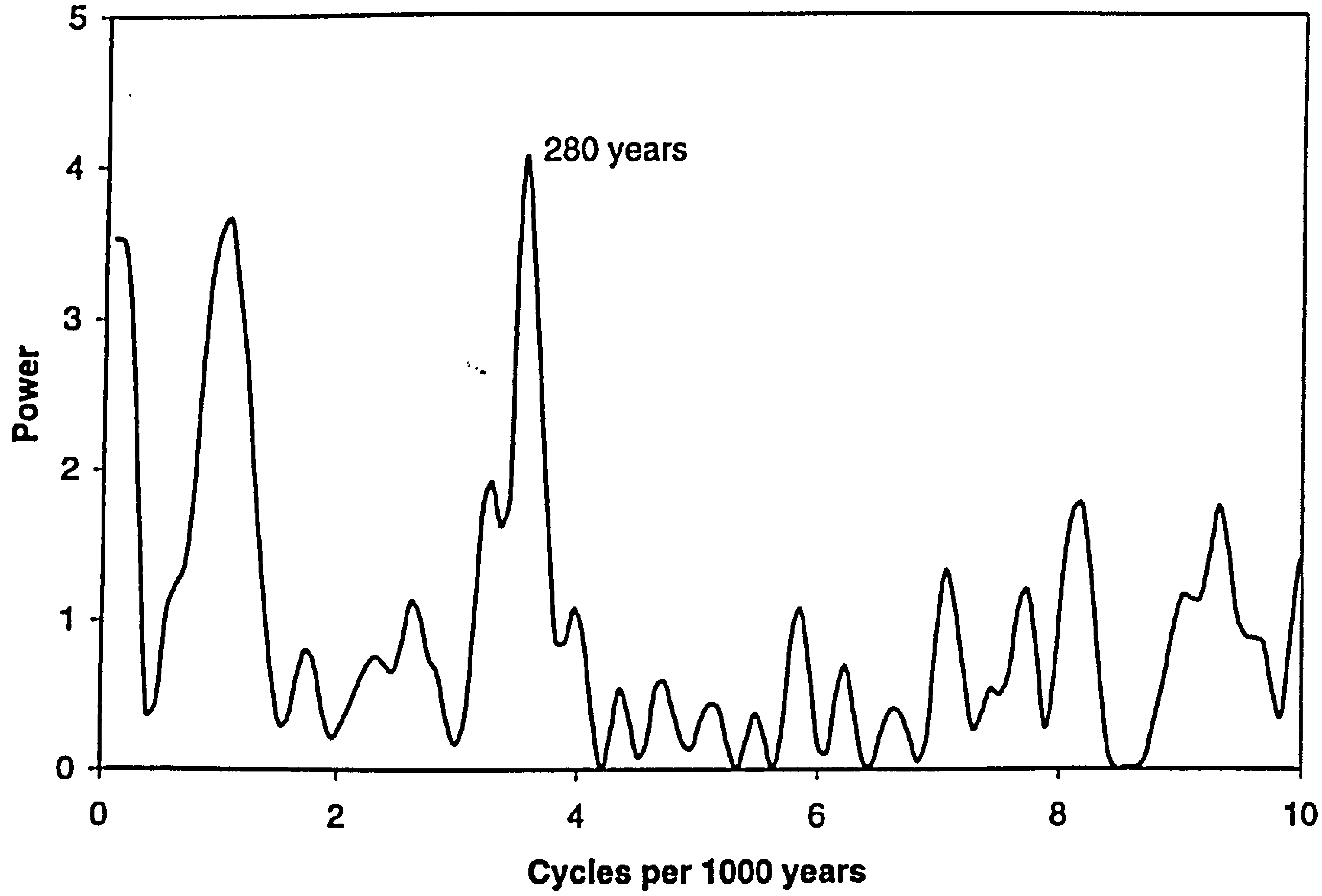
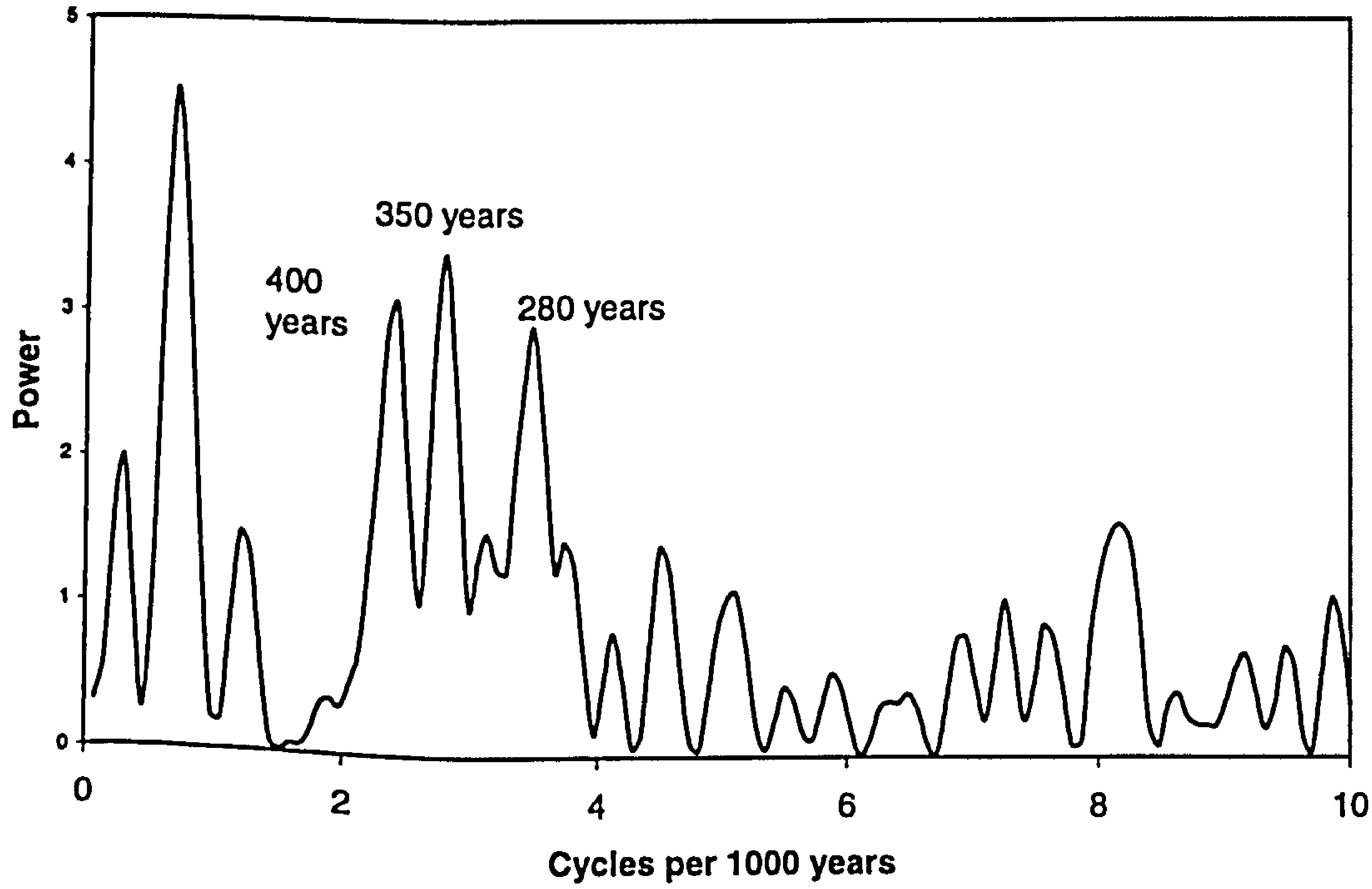


Figure 6.66 Periodogram showing humification values from Craigmaud Moss over the last ca. 2000 years.



Chapter 7 - Discussion

7.1 Climatic changes

Climatic fluctuations have been discussed from each of the seven sites in chapter 6, and compared on an intrasite scale. It is evident that specific wet and dry phases can be determined from each site, shown by identical temporal variations in each of the three proxy reconstructions. The syntheses of climatic changes from each site can now be compared, with the prospect of identifying similar phase changes, based on radiocarbon and tephra dated age/depth models, and synchronous changes using the tephra isochrones. Before this is undertaken, it is necessary to discuss the notion of climatic changes in relation to ecological thresholds, and how they can be recognised in the palaeoecological record.

7.1.1 Climatic phases and ecological thresholds

"The phrase 'climatic fluctuation' is not quantitative, in general has only the vaguest significance, and is expressed when some threshold value in the climatic system is passed" (Conway, 1948).

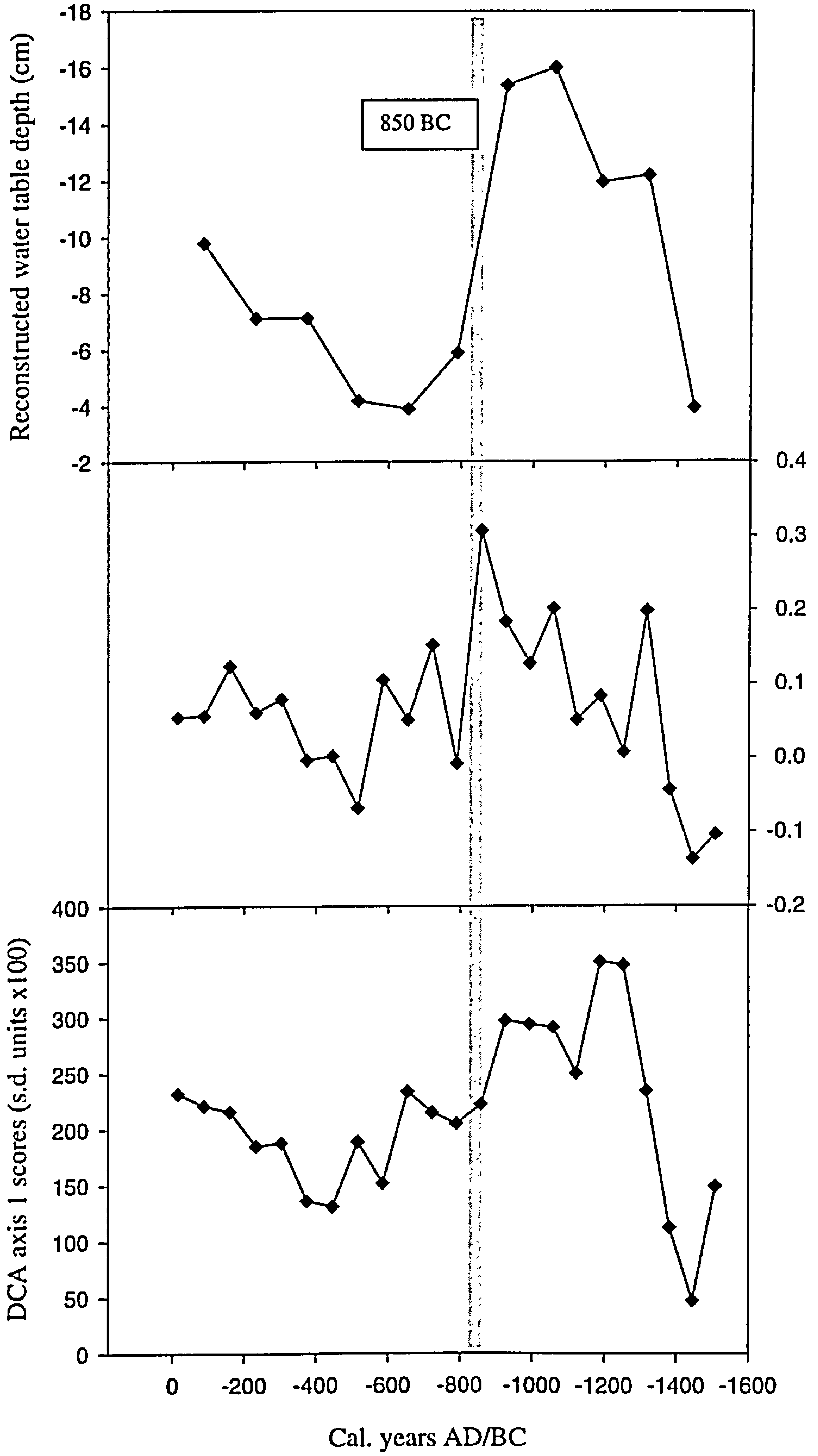
The idea of an ecological threshold, with respect to climatic tolerance, suggests that any climatic change will cross some threshold, but a change in vegetation may not necessarily take place depending on the threshold which has been crossed. This suggestion also highlights the potential lag in vegetation changes with respect to climatic changes (*cf.* Chambers *et al.*, 1997a), although mire taxa, especially *Sphagnum*, can respond very quickly to a change in environment. Conway (1948) suggested that although broad agreements would be expected for any major climatic change, the crossing of a critical threshold would be expected at slightly different times over northwest Europe, owing to differences in latitude and proximity to the Atlantic.

An example of passing across an ecological threshold, and how this can be interpreted with respect to a climatic phase, can be taken from Temple Hill Moss, from the period 1600 BC - 0 AD/BC (Figure 7.1). The main problem associated with this example is what precise date to assign to the wet phase. Initially, it appears that there is a shift to a wet phase around *ca.* cal. 850 BC in the testate amoebae and humification data, shown by the large change in direction of the records at these times. There is also a slight variation in the DCA record at this time, although it is under one standard deviation and therefore probably not significant. A climatic wet shift has been documented in many sites around northwest Europe around 850 BC (van Geel *et al.*, 1996), and thus the data from Temple Hill Moss would appear to fit the known model. Analysing the data in more detail, however, suggests that the change in humification fluctuates after *ca.* cal. 850 BC, becoming initially drier again, only to show lower values again by *ca.* cal. 500 BC. The reconstructed water table depths from testate amoebae stay at low levels throughout *ca.* cal. 850-500 BC, but were sampled at a lower resolution, and therefore may miss more subtle fluctuations from the record. Most interestingly, the DCA record actually starts declining from *ca.* cal. 1200 BC, with minor fluctuations, until the wettest phase between *ca.* cal. 600-400 BC. Care must therefore be taken when attempting to correlate the climatic deterioration with the 850 BC event (*sensu* Baillie, 1991).

The plant macrofossil changes associated with the change in DCA correspond to the decrease in *Sphagnum imbricatum* from 172-148cm (*ca.* cal. 1250-850 BC), being replaced by *Sphagnum* section *Acutifolia*, until *Sphagnum* declines after *ca.* cal. 400 BC. Evidence from the humification and reconstructed water table depths suggest that the rapid decline of *Sphagnum imbricatum* at 850 BC is associated with a wet shift. However, as the initial growth of *S. imbricatum* during this phase occurred under dry climates (according to the humification and reconstructed water table depths), then presumably as *S. imbricatum* was declining between *ca.* cal. 1250-850 BC so was the climate. This then raises the question, at which point was the threshold crossed?

This is a hard question to answer, as the levels of *Sphagnum imbricatum* can be seen to decline steadily over *ca.* 550 years (from *ca.* cal. 1250-700 BC, see Figure 6.22), over which time the humification steadily rises, and water table depth falls from -12cm to -16cm. Interpreting these data in the same manner as above, suggests that a drying/warming climate appears to be responsible for the steady decline in *Sphagnum imbricatum*, or maybe there are lags in the vegetation/climate system, or thresholds

Figure 7.1 Evidence from plant macrofossil DCA axis 1 scores, humification, and reconstructed water table depths from testate amoebae at Temple Hill Moss from 1600 BC - 0 AD/BC. This diagram highlights the problems with dectecting synchronous climatic phases from all of the proxy records.



related to the growth of *Sphagnum imbricatum* are being crossed due to a highly variable climate. The complicated ecology of *Sphagnum imbricatum*, which can produce difficulties when attempting to model its response to climatic change using DCA reconstructions, is discussed further in section 7.3.1.

In summary, the clear initial decline around *ca.* cal. 850 BC in the reconstructed water table depth, humification record, and decline in *Sphagnum imbricatum* suggests a climatic deterioration. The subsequent fluctuations in humification and DCA records suggest a further wet phase until around *ca.* cal. 500 BC, and hence a climatic deterioration has been inferred between *ca.* cal. 850-500 BC at Temple Hill Moss.

7.1.2 Coherent phases of change between the peat records

The site synthesis data from Chapter 6 has been summarised in Tables 7.1 and 7.2, and is shown graphically in Figures 7.2 and 7.3. Certain phases of climatic deterioration can be identified in Figure 7.2, including around *ca.* cal. AD 1700-1800 at Temple Hill, Craigmaud, Mallachie and Ben Gorm Moss. These sites may have been responding to the later stages of the Little Ice Age (LIA), the coldest conditions of which have been suggested to be between AD 1570 and 1730 (shown by a wet/cold phase at Shirgarton Moss around *ca.* cal. AD 1550), and in the nineteenth century in the Northern Hemisphere (Bradley and Jones, 1993). Barber *et al.* (1999) identified coherent wetter/colder phases of climate from a montane bog in the Cairngorms (Moine Mhor) and a lowland raised bog in Northern Ireland (Fallahogy) between cal. AD 1680 and 1850. Not all individual records, however, show the same patterns, and changes in the general circulation will result in differing patterns (Bradley and Jones, 1993), as has been observed in the instrumental records for the present century (Briffa and Jones, 1993). This pattern is demonstrated in the peat based proxy record from Craigmaud Moss, which shows distinct wet/cold phases around *ca.* cal. AD 1720 and *ca.* cal. AD 1470, although exhibiting comparatively drier/warmer conditions between *ca.* cal. AD 1530-1660. The only other site to register two climatic deteriorations within the last *ca.* 500 years is Ben Gorm Moss, which displays wet phases around *ca.* cal. AD 1570 and *ca.* cal. AD 1810. Only these two northern most sites show a two stage phase of the LIA, although this may be due more to problems with the recent peats than climatic insensitivity.

Interpreting variations in proxy records from the top layers of peat stratigraphy is troublesome, especially as these are frequently difficult to date (e.g. Oldfield *et al.*, 1995; 1997), and may contain hiatuses due to either climatic drying or anthropogenic activities. The late LIA phase has not been recorded at Langlands Moss due to suspected disturbance of the surface peats (see section 5.3.2). The plant macrofossil DCA (Figure 7.6) and humification (Figure 7.7) records at Longbridge suggest a possible climatic deterioration around this time, but are contradicted by the testate amoebae reconstruction (Figure 7.8), which is dominated by xerophilous taxa (see Figure 6.5). The stratigraphy at Shirgarton Moss around this time is dominated by *Sphagnum magellanicum*, which has a large range of moisture regimes over which it can grow. The testate amoebae or humification data from Shirgarton both fail to show a significant wet shift around this time. Although the data for Mallachie Moss suggests a significant wet phase between *ca.* cal. AD 1700-1800, the suggested hiatus within the sequence (see sections 5.3.5 and 6.10.1) casts doubt on the dating of the top of the profile, and hence climatic inferences around this time cannot be made with any confidence. There is also a potential problem with the detection of a low peak in the humification signal at Ben Gorm around *ca.* cal. AD 1810, which may have been due to incomplete decay of the acrotelmic peats, giving an erroneous signal.

Another 'band of agreement' between the sites signifying a wet phase occurs between *ca.* cal. AD 900-1100. The only sites which do not appear to have been sensitive to a significant shift in the proxy records are Temple Hill Moss and Craigmaud Moss, the two most eastern sites. The earliest detection of a climatic signal around this time is from Longbridge Moss which records a change at *ca.* cal. AD 870. Although the wet phases identified in Table 7.1 and Figure 7.2 suggest periods when the climate was definitely wetter, the phase shifts are not always clear due to the complexity of determining exactly when the threshold had been crossed (see section 7.1.1). Palaeoecological evidence for a climatic deterioration around this time has also been identified in Scotland by Stoneman (1993), Chambers *et al.* (1997) and Anderson (1998), as well as in northern England by Barber (1981), Mauquoy and Barber (1999a), Charman *et al.* (1999) and Hughes *et al.* (submitted). Hence, changes around this time can be documented as broadly synchronous in northern Britain, and possibly in northwestern Europe, according to the data of Aaby (1976) and Nilssen and Vorren (1991), although these shifts appear to be slightly later than in northern Britain.

Good evidence exists from four of the sites (Longbridge, Shirgarton, Craigmaud and Ben Gorm) for a dry climatic phase between *ca.* cal. AD 1200-1400, with Temple Hill Moss displaying a drier phase between *ca.* cal. AD 920-1160, which coincides with the wet phase discussed above. This may be due to the easterly location of Temple Hill, as the other eastern site, Craigmaud Moss, also fails to pick up the wet phase between *ca.* cal. AD 900-1100, suggesting possible climatic differences between the eastern and more western sites. There is plant macrofossil and humification evidence to suggest a drier phase between *ca.* AD 1200-1400 at Langlands Moss, although the testate amoebae data imply a comparatively high water table.

A climatic wet phase can be detected at four sites (Langlands, Temple Hill, Mallachie and Craigmaud) around *ca.* cal. AD 500-600. This was not detected in any of the other proxy records from Scotland discussed in Chapter 2 (see table 2.1), although this event has been identified in many peat based proxy climatic records from northern England and northwest Europe, notably from three sites by Blackford and Chambers (1991). Another wet phase around *ca.* cal. 750-950 BC can be identified from every site, which corresponds well with the other records from northwest Europe (van Geel *et al.*, 1996; and see Table 2.1).

Other notable wet phases identified from the sites occur between *ca.* cal. 1400-1240 BC, 1630-1560 BC, 1800-1900 BC, and around *ca.* cal. 2300 BC, which agree well with other phases identified in the literature for Scotland (see Table 2.1). The dry phases appear to be more site specific, although there is a suggestion of a consistent dry phase between *ca.* cal. 400-200 BC in the three most northern sites.

Table 7.1 Wet phases for each site recorded by at least two of the proxy indicators from the site synthesis.

Site	AD					BC								
Longbridge Moss		1350	870			250		940	1380			2250		*
Langlands Moss			900	600	300		500	930	1240	1560	1900			
Temple Hill Moss	1700-1800			500-600				850-500		1450	1900	2300		3350 3900 4700
Shirgarton Moss		1550	1050		340 60			800	1340	1570				
Mallachie Moss	1850		950-1040	500		10		850	1350		1700	2200	2500	
Craigmaud Moss	1720	1470		600				750						
Ben Gorm Moss	1810	1570	1090-1150				660	900	1390	1630	1870-1810			

Table 7.2 Dry phases for each site recorded by at least two of the proxy indicators from the site synthesis.

Site	AD					BC							
Longbridge Moss		1250		430-680			750		1570				
Langlands Moss				520		110		1040		1700			
Temple Hill Moss		920-1160			100			1250-920		1750		2900	4400 4900
Shirgarton Moss		1270-1430			50			900 1100	1450				
Mallachie Moss	1420-1630			350		390	860		1530		2350	2700	
Craigmaud Moss	1530-1660	1210-1340	770	450		440-180							
Ben Gorm Moss		1270	600-840			420-AD 10 540	780			1690 1990			

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[*] These bold lines indicate the age limit of the palaeoecological records

Figure 7.2 Climatic wet phases from all three proxy records at each site.

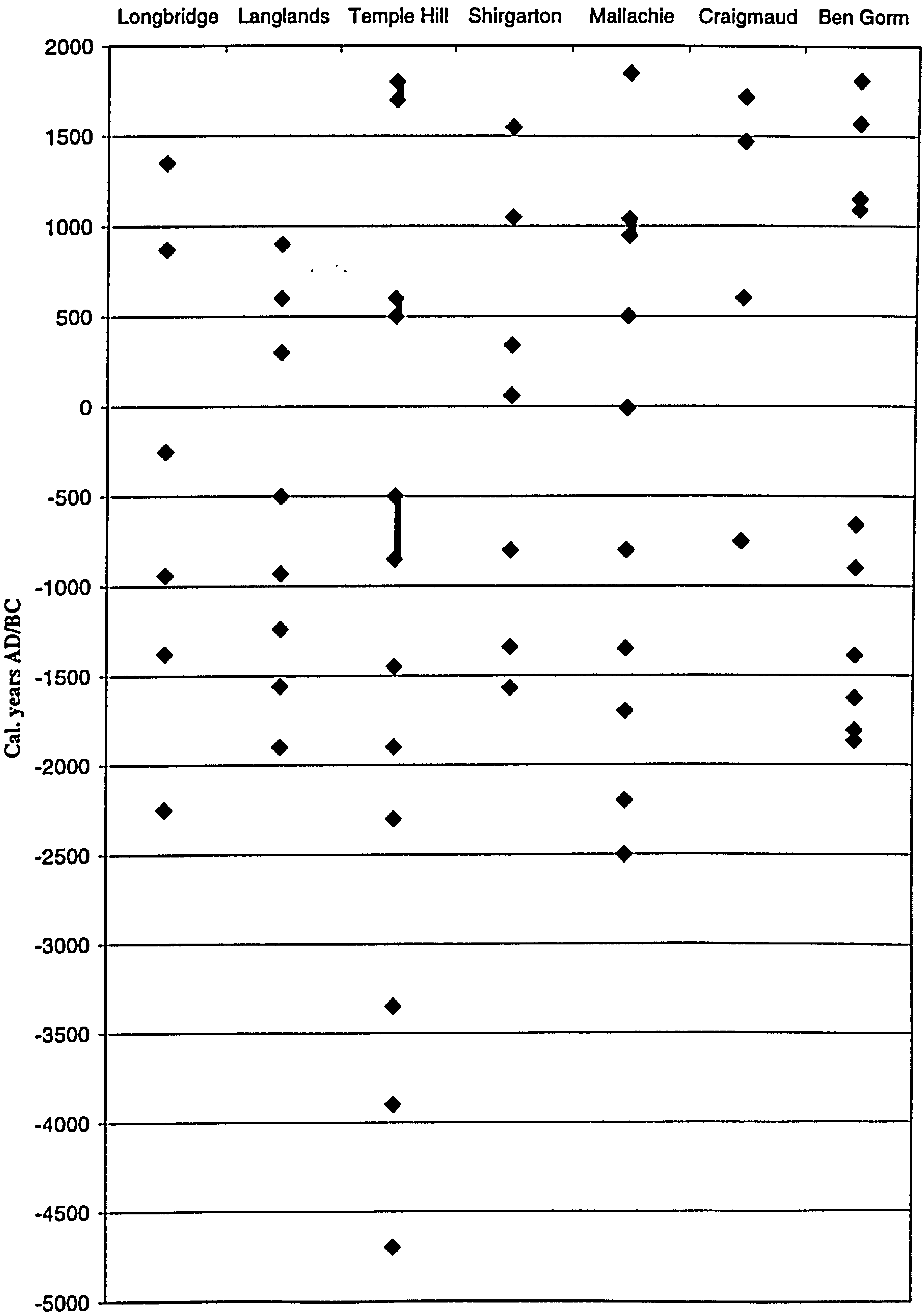
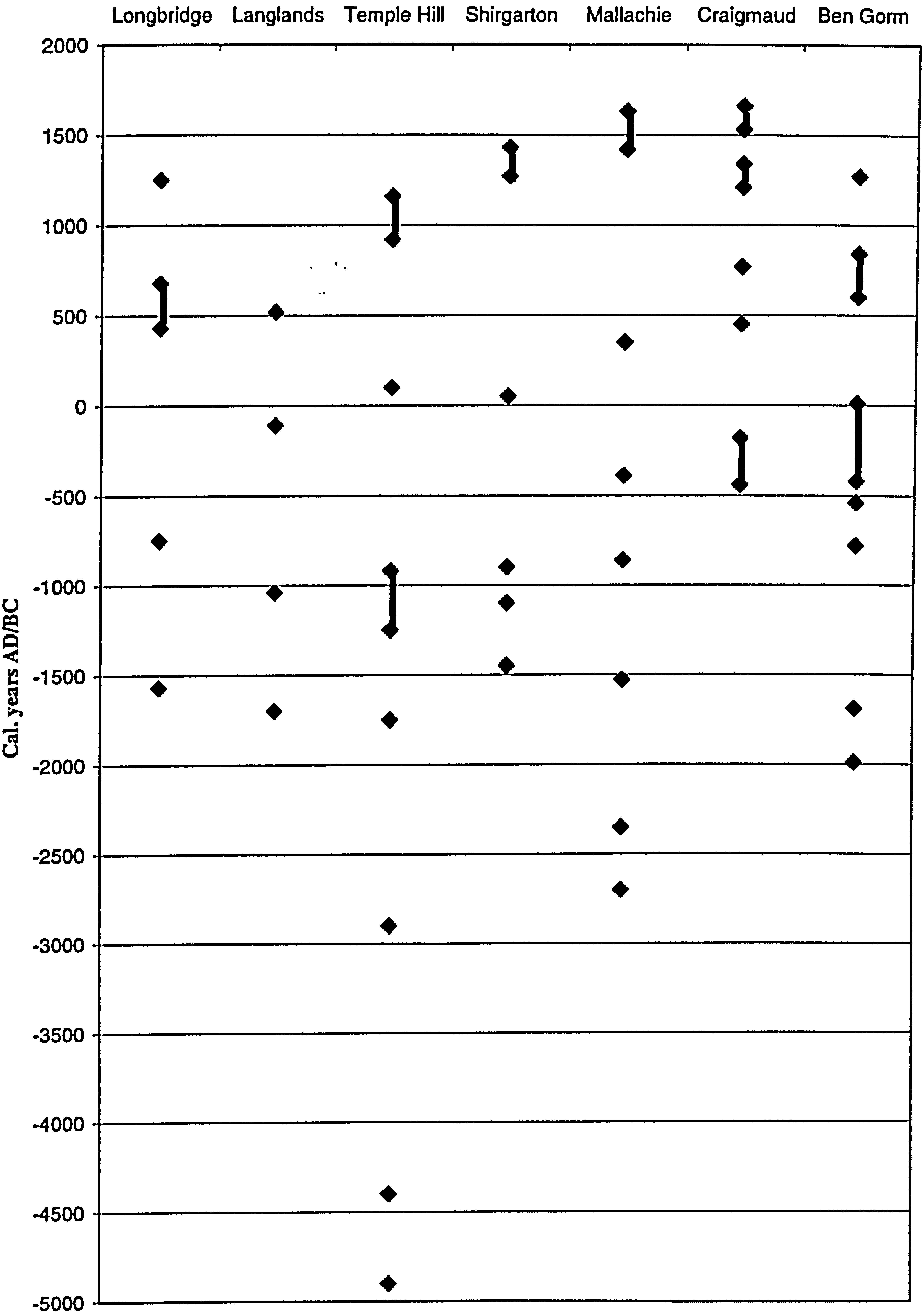


Figure 7.3 Climatic dry phases from all three proxy records at each site.



There has been little independent evidence to corroborate the possible climatic change around *ca.* 4000 BP which has been suggested as being responsible (possibly in conjunction with human activity) for the initiation of blanket peats and decline of *Pinus sylvestris* in Scotland (e.g. Birks, 1977; 1996). Recent evidence has been presented by Anderson *et al.* (1998) for a climatic deterioration in northwest Scotland between cal. 3900-3500 BP (1950-1550 BC). Supporting evidence from this study is found at Ben Gorm Moss, also in the northwest of Scotland, which provides evidence for climatic deteriorations around *ca.* cal. 1870-1810 BC and *ca.* cal. 1630 BC. There is also an inferred wet phase around *ca.* cal. 1700 BC at Mallachie Moss, with Langlands Moss and Shirgarton Moss displaying wet phases within 50 years of *ca.* cal. 1600 BC, and another wet phase at Langlands Moss and Temple Hill Moss around *ca.* cal. 1900 BC. The 'pluvial phases' identified from pine stumps in the Cairngorms by Dubois and Ferguson (1985) are well represented at Mallachie Moss. Dubois and Ferguson (1985) identified pluvial phases around cal. 1600 BC and cal. 2850-2430 BC which correspond with the wet phases around *ca.* cal. 1700 BC and *ca.* cal. 2500 BC at Mallachie. There is evidence from three sites (Longbridge Moss, Temple Hill Moss and Mallachie Moss) for a climatic deterioration between *ca.* cal. 2500-2200 BC, which has been recognised in many sites from northwest Europe, but as noted in Chapter 2, has not previously been identified in Scotland. Climatic deteriorations can therefore be determined from the sites in this project at around 4000 BP, although some of the sites exhibit phase changes before 4000 BP, with some coming later. Pine pollen was counted in the testate amoebae analyses in order to try to ascertain if the climatic deterioration (and Hekla-4 tephra, *cf.* Blackford *et al.*, 1992; Hall *et al.*, 1994b) preceded or postdated the pine decline. However, sufficient grains of pine pollen were not encountered in the analyses to enable any hypotheses to be tested rigorously, and it is suggested that this could be attempted for future work.

7.1.3 Tephra isochrones - precise correlations

The Glen Garry isochrone occurs at each of the sites with the exception of Longbridge Moss. The palaeoclimatic inferences made from the proxy records when the isochrone was deposited, are shown in Figures 7.4(a-c), and can be correlated precisely between sites. Similar palaeoclimatic reconstructions are shown for the Hekla 4 isochrone in

Figure 7.4a One sample before the Glen Garry tephra isochrone and corresponding climatic reconstructions

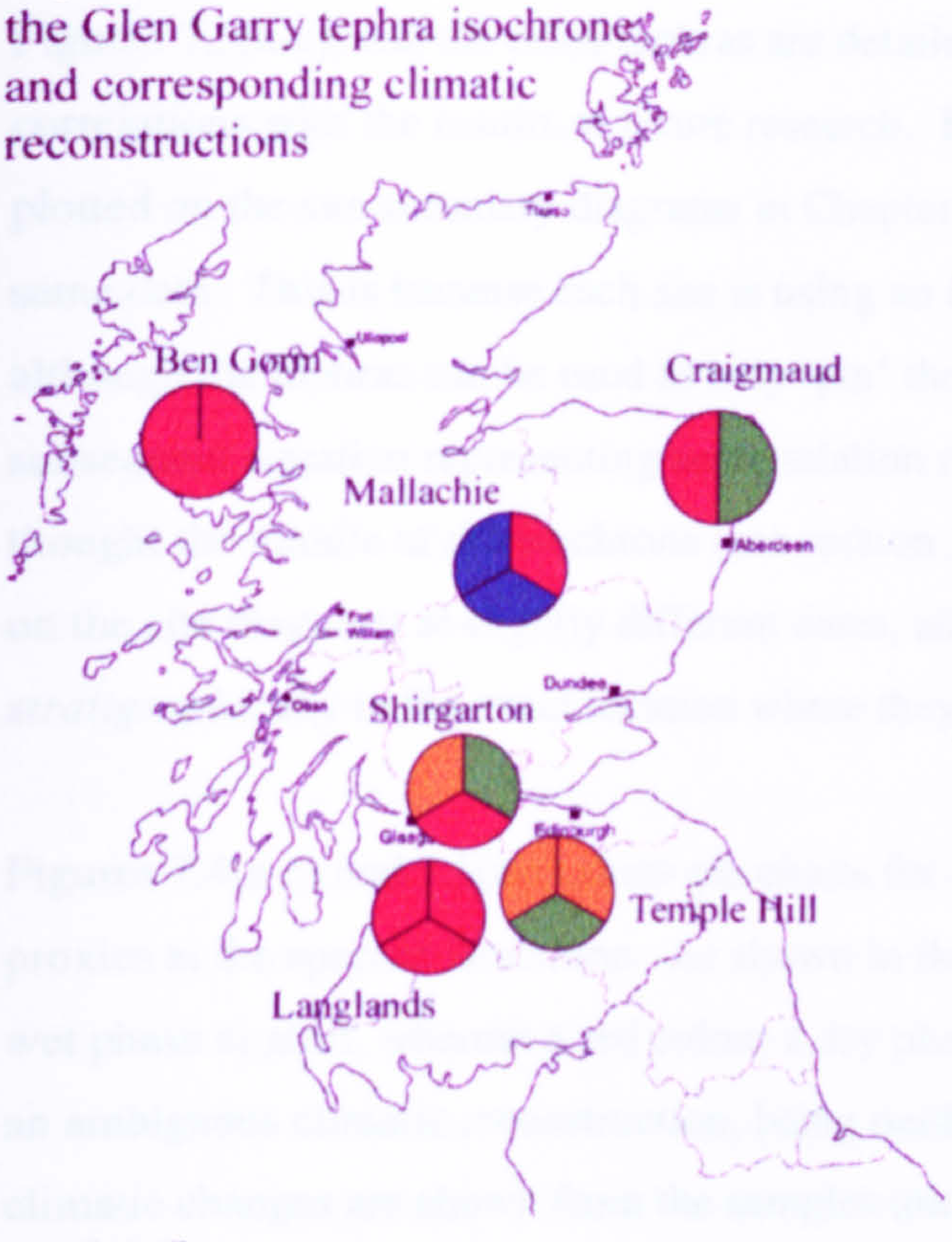


Figure 7.4b The Glen Garry tephra isochrone and corresponding climatic reconstructions

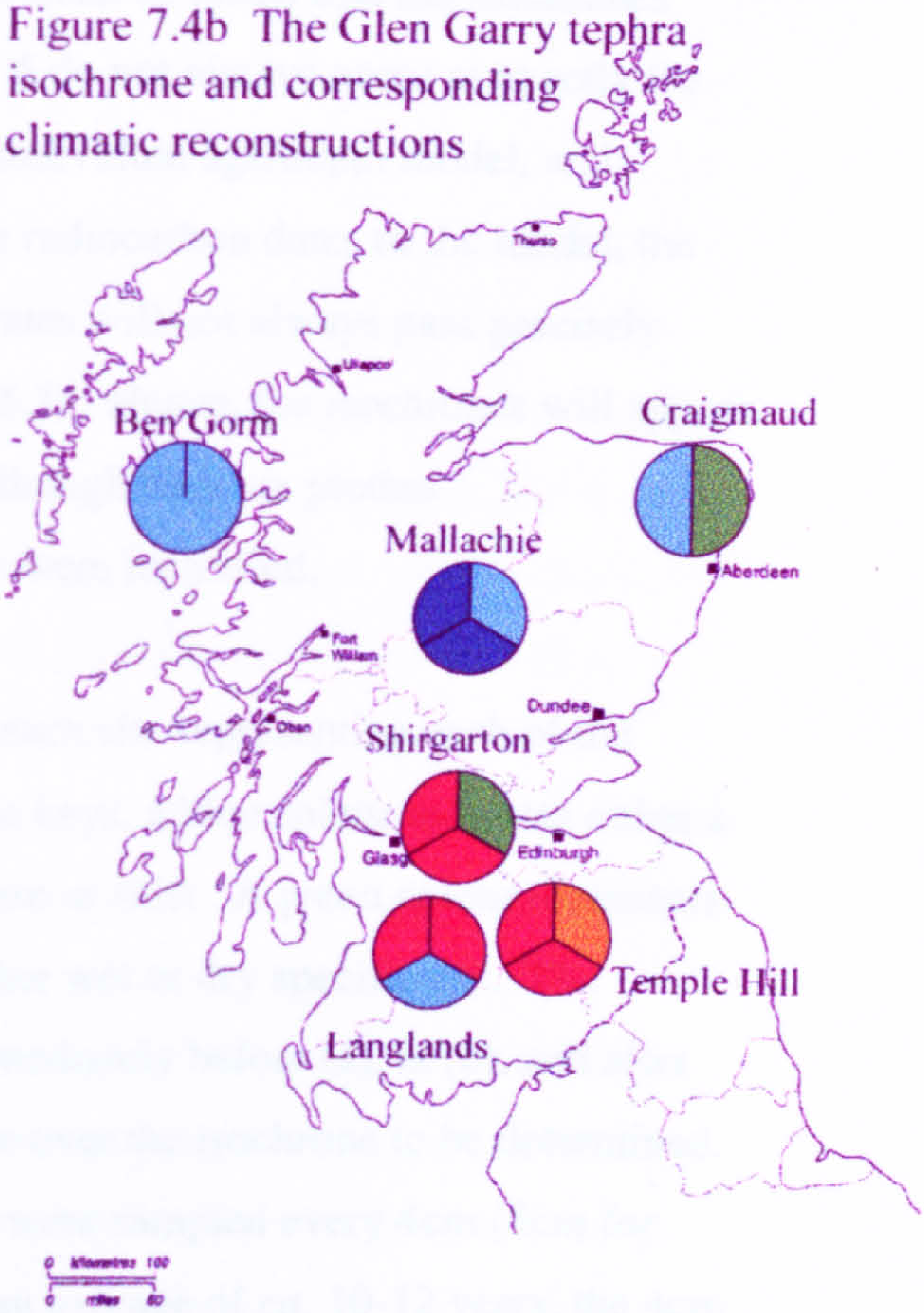
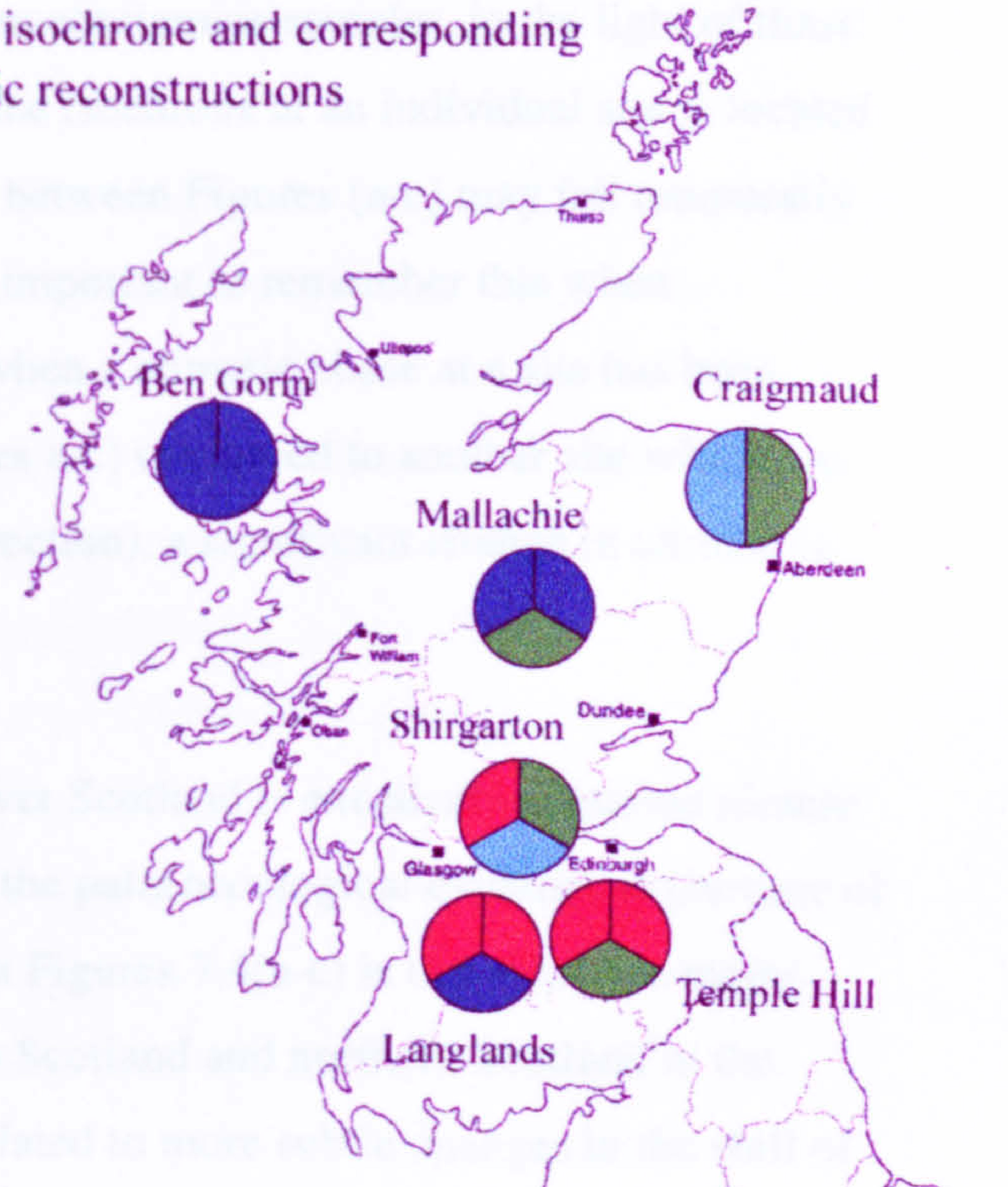


Figure 7.4c One sample after the Glen Garry tephra isochrone and corresponding climatic reconstructions



Key

TA=testate amoebae
DCA=detrended correspondence analysis
Hum=humification

■ wet phase
■ wet shift
■ dry phase
■ dry shift
■ stable climate

For Craigmaud: right half is DCA, left half is Hum
For Ben Gorm: all Hum

Figures 7.5(a-c), and the other tephras are detailed later in the text to allow precise correlations with the results of future research. It must be noted that the isochrones plotted on the site summary diagrams in Chapter 6 do not always occur at exactly the same date. This is because each site is using an individual age/depth model, and although the tephras can be used to help 'pin' the radiocarbon dates to the model, the subsequent equation representing accumulation rates will not always pass precisely thought the middle of the isochrone (see section 5.3). Hence, the isochrones will appear on the site diagrams at slightly different dates, although they are plotted *stratigraphically* in the exact location where they were identified.

Figures 7.4(a-c) and 7.5(a-c) show pie charts for each site representing each of the proxies at the specific isochrone. As shown in the keys, a blue colour indicates either a wet phase or shift, whereas a red colour a dry phase or shift. A green colour represents an ambiguous climatic reconstruction, being neither wet or dry specifically. The climatic changes are shown from the samples immediately before (a), at (b), and after (c) the isochrone, enabling the direction of change over the isochrone to be determined. Even though the proxy reconstruction techniques were sampled every 4cm (8cm for testate amoebae), with each sample representing an average of *ca.* 10-12 years, the 4cm interval will thus represent, on average, 40-50 years. Future work is expected to be done at a higher resolution, for example 1cm contiguous samples, in the light of these promising results. Depending on whether the isochrone at an individual site is located near a sample or not, the records displayed between Figures (a-c) may fall temporally close together or a few decades apart. It is important to remember this when interpreting Figures 7.4 and 7.5, although when a climatic phase at a site has been shown to change over the isochrone (Figures a-c) compared to another site which has not changed (or changed in the opposite direction), a significant change in climate is considered.

As coverage of the Glen Garry isochrone over Scotland is excellent, a detailed picture of climatic change can be ascertained from the palaeoecological evidence at the time of this tephra. What is immediately clear from Figures 7.4(a-c) is that there are major differences between the climate of southern Scotland and northern Scotland in the periods examined. This may possibly be related to more subtle changes in the shift of the polar front, as described in Chapter 2. It can therefore be deduced that climatic changes in northern and southern Scotland occurred asynchronously. The sites in central/southern Scotland (Langlands, Temple Hill and Shirgarton Moss) show dry climates and dry shifts in the samples before the Glen Garry isochrone, and remain dry at and after the isochrone. The exceptions to this pattern are the humification record at Langlands Moss, which documents a wet shift at the isochrone, and the humification record at Shirgarton which exhibits a wet shift just after the isochrone. Although the reconstructed water table level from testate amoebae data is comparatively low (-4.7cm) at Shirgarton, this is relatively dry compared to the rest of the stratigraphy, and is represented by a peak in *Difflugia pulex* (Figure 6.34), which has no modern analogue and is poorly modelled. This lowering of the water table also corresponds with an increase in humification, and is thus considered comparatively dry. The testate amoebae and DCA records at Temple Hill Moss show a dry shift before the isochrone, resulting in a dry phase after the isochrone. In contrast, Ben Gorm Moss exhibits a wet shift through the isochrone, being dry just before, to wet in the sample after the isochrone. The same pattern of change occurs in the DCA data from Mallachie Moss, with the other proxy records documenting a wet phase throughout. Craigmaud Moss has an ambiguous DCA record, being dominated by UOM around this time, although the humification data displays a clear wet shift around the isochrone. It therefore appears that different climatic changes were affecting northern and central/southern Scotland at the time of deposition of the Glen Garry tephra.

A similar pattern of climatic change can be observed for the Hekla-4 isochrone (Figures 7.5a-c). Although only four sites are represented, there is a clear change at Longbridge Moss from the sample before the isochrone representing a dry phase (testate amoebae water level reconstruction and DCA), although the humification record documents a wet phase throughout the samples, to a wet shift at and after the isochrone. This pattern is mirrored by the humification record at Craigmaud Moss. In the Cairngorms, Mallachie Moss displays a wet climate immediately before the isochrone, but becomes comparatively drier through and immediately after the isochrone. The humification record from Ben Gorm Moss is very dry before and during the isochrone, although shows a wet shift immediately after Hekla-4 deposition.

Figure 7.5a One sample before the Hekla-4 tephra isochrone and corresponding climatic reconstructions

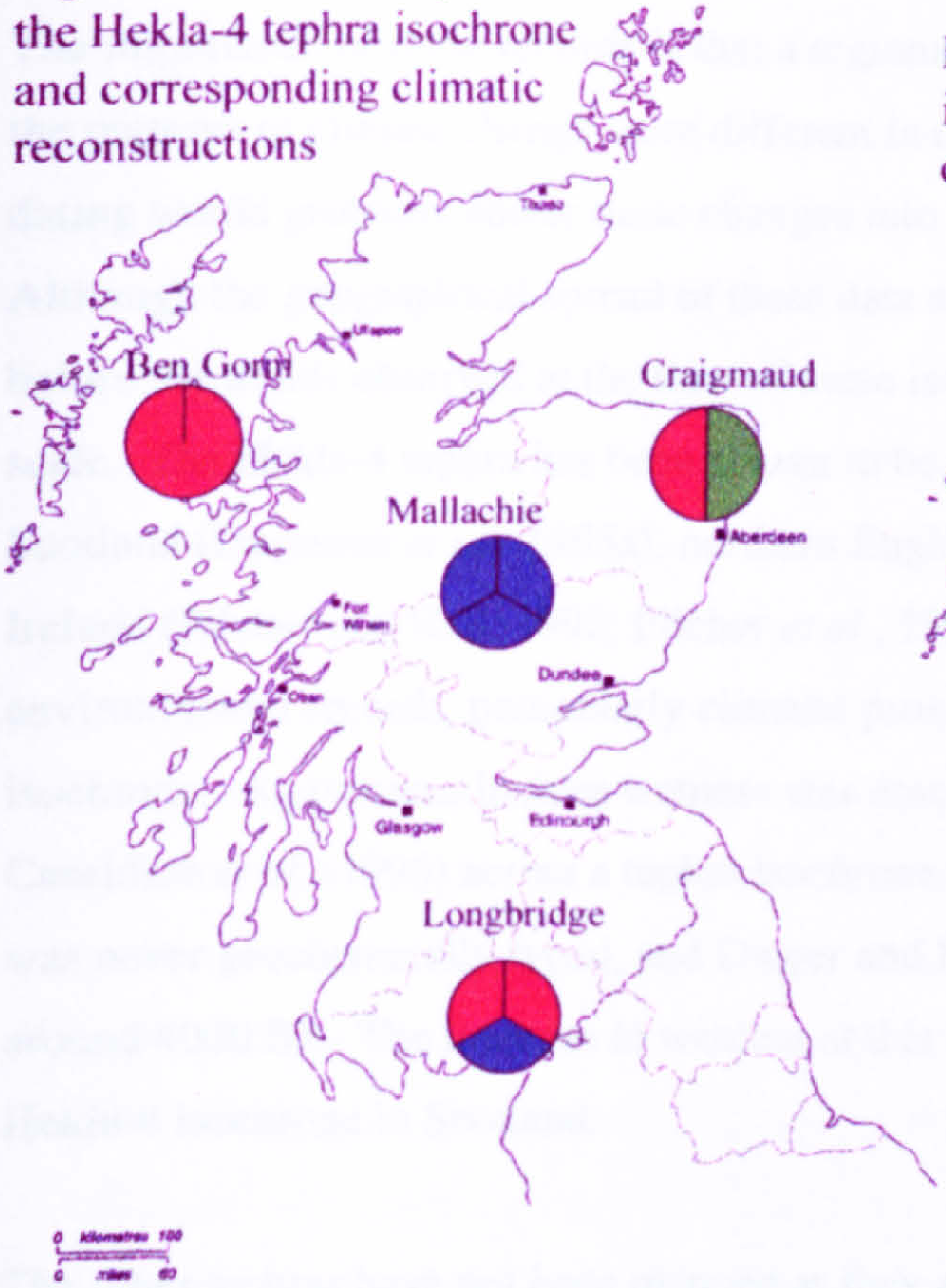


Figure 7.5b The Hekla-4 tephra isochrone and corresponding climatic reconstructions

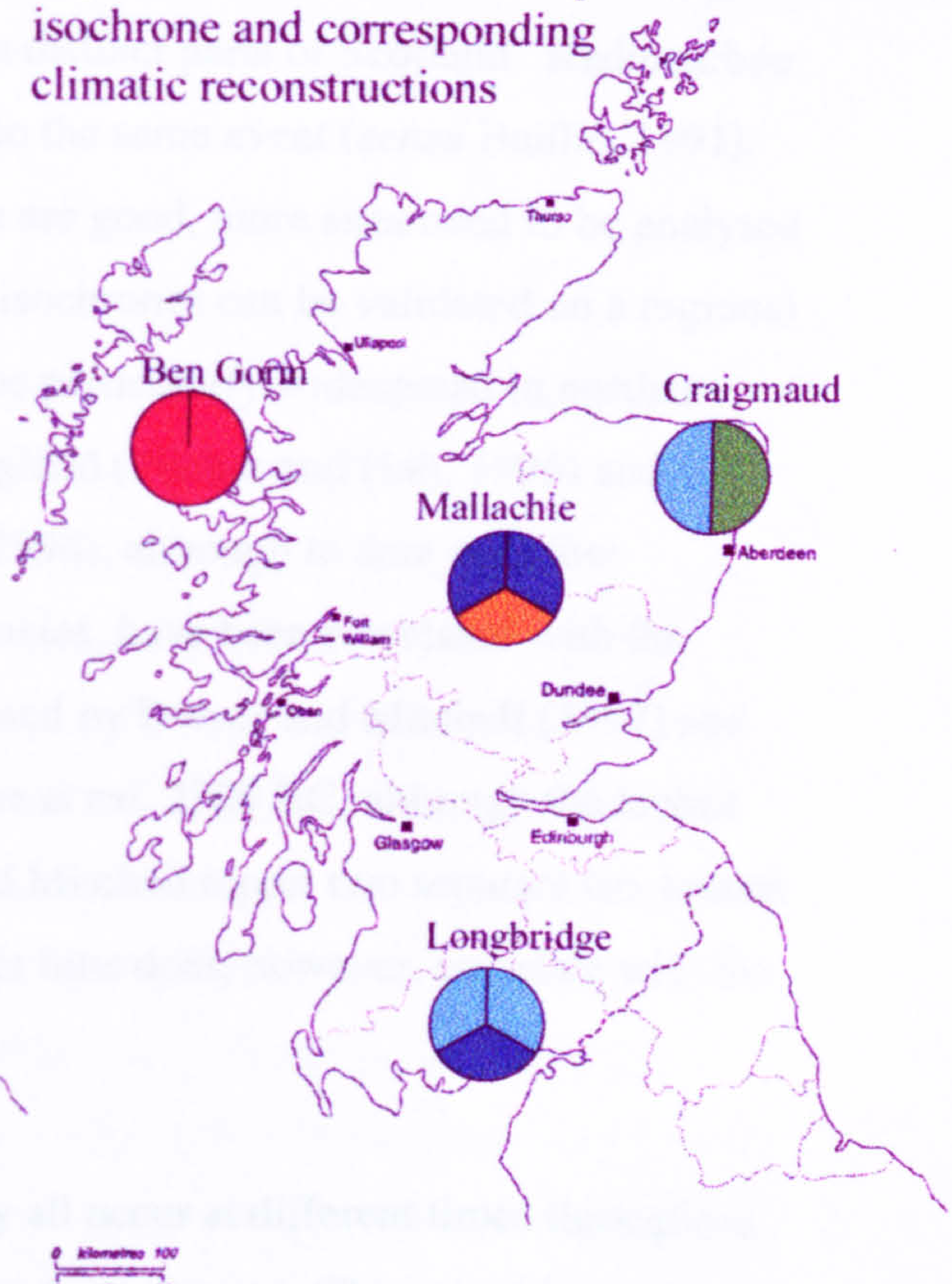
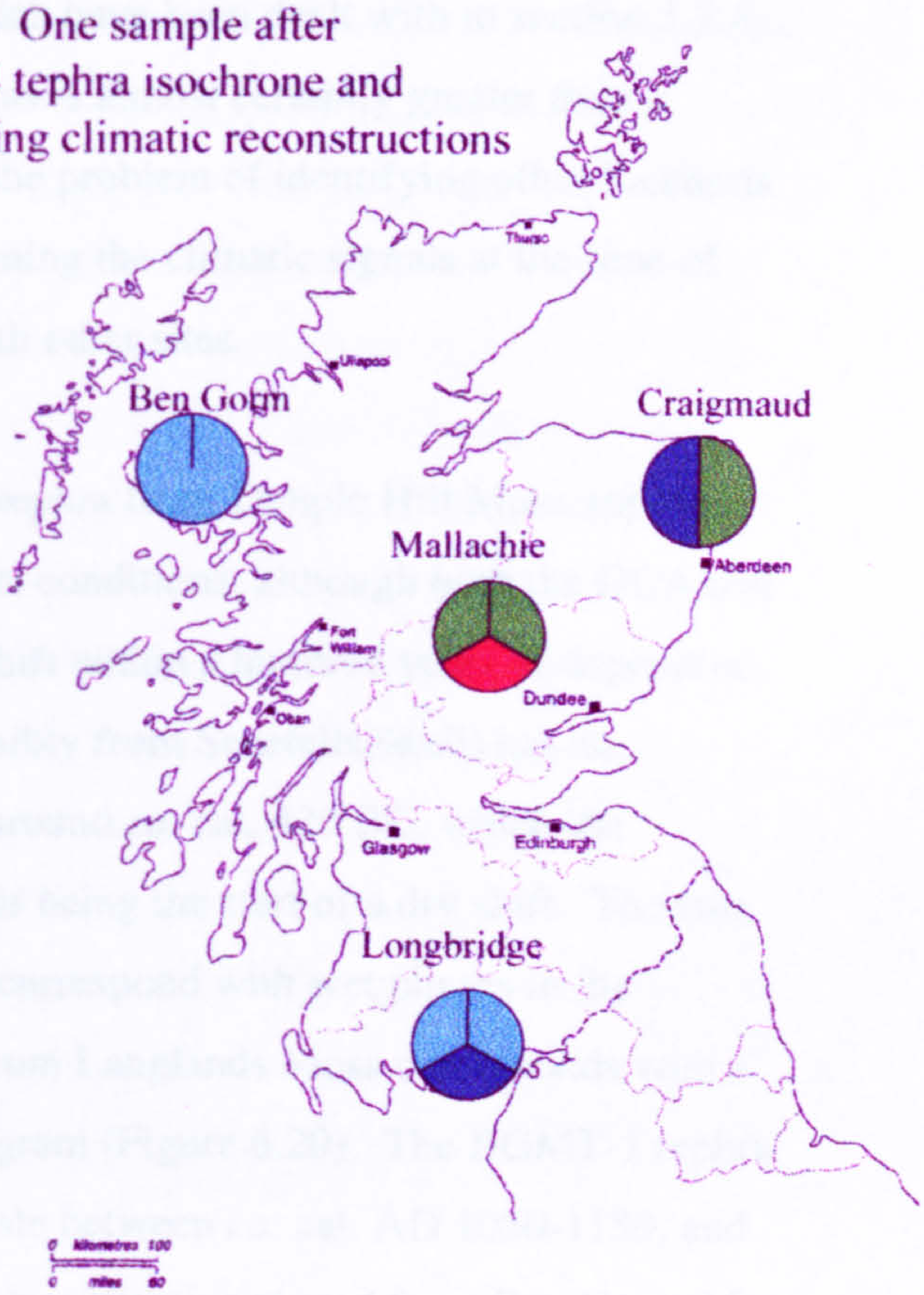


Figure 7.5c One sample after the Hekla-4 tephra isochrone and corresponding climatic reconstructions



Key

TA=testate amoebae
DCA=detrended correspondence analysis
Hum=humification

🟦 wet phase
🟦 wet shift
🟥 dry phase
🟧 dry shift
🟩 stable climate

For Craigmaud: right half is DCA, left half is Hum
For Ben Gorm: all Hum

The importance of these records is that a regional picture can be presented, showing that the patterns of climate change were different in distinct parts of Scotland. Radiocarbon dating would probably smear these changes into the same event (*sensu* Baillie, 1991). Although the geographical spread of these data are good, more sites need to be analysed before the trends observed at the time of these isochrones can be validated on a regional scale. The Hekla-4 tephra has been shown to be particularly widespread in northern Scotland (Dugmore *et al.*, 1995a), northern England (Pilcher and Hall, 1996) and Ireland (Pilcher and Hall, 1992; Pilcher *et al.*, 1996), although to date very few environmental records, particularly climatic proxies, have been correlated with the isochrone. An increase in mire wetness was noted by Dwyer and Mitchell (1997) and Caseldine *et al.* (1998) across a tephra isochrone at cal. 2300 BC, although the tephra was never geochemically typed, and Dwyer and Mitchell report two separate isochrones around 4000 BP. The increase in wetness at this time does, however, correlate with the Hekla-4 isochrone in Scotland.

The other tephras have not been mapped as they all occur at different times throughout the Holocene, with none being identified at more than one site. This raises not only questions as to tephra dispersal (most of which have been dealt with in section 5.2.4), but suggests that the coverage of these tephras is almost certainly greater than is represented in this thesis. This only leaves the problem of identifying other locations where the tephras can be found, and determining the climatic signals at the time of deposition, to enable precise correlations with other sites.

Starting in chronological order, the Lairg A tephra from Temple Hill Moss appears to have been deposited under comparatively wet conditions, although both the DCA and humification reconstructions suggest a dry shift within a hundred years of deposition. The BGMT-3 tephra identified in Skye (possibly from Snaefellsjökull) has an interpolated date from the age/depth model around *ca.* cal. 420 BC, which the humification data from Ben Gorm suggests as being the start of a dry shift. The two historic tephras identified in this thesis both correspond with wet phases in the palaeoclimatic records. The AD 860 layer from Langlands Moss corresponds with a wet phase, as shown in the site summary diagram (Figure 6.20). The BGMT-1 tephra (possibly Beinn Eighe) has an interpolated date between *ca.* cal. AD 1050-1180, and correlates with a significant wet phase in the humification record from Ben Gorm Moss.

These historic isochrones in particular are vital for precisely dating and correlating historic climatic changes, in order for any correlations with solar variability or volcanic activity to be ascertained. There is thus an urgent need for precise dates to be assigned to all of these tephra isochrones, either by correlation with historic volcanic events, or by high precision wiggle match dating for the older isochrones. Only then can tephrochronology move on from precise correlations between sites to being able to precisely date climatic events archived within the Holocene strata.

7.1.4 Summary of the palaeoclimatic reconstructions

Before invoking a broad, dramatic climatic shift, it is always desirable to have sources of evidence from many regions (Adams *et al.*, 1999). In this case, seven sites have been analysed for palaeoclimatic changes throughout Scotland, each one differing in geographical and climatological characteristics. The establishment of synchronous climatic changes at different sites thus implies that a regional climatic signal can be ascertained, although defining synchroneity on a radiocarbon timescale is not without its problems (see sections 7.1.3 and 7.4). All the sites document a climatic deterioration between *ca.* cal. 940-750 BC. This may in fact be the same event recorded at slightly different times (van Geel *et al.*, 1996), due to either differing thresholds in the ecology of the individual mires, or inaccuracies in the radiocarbon chronologies. Other events are well documented in the majority of sites. They may not have been recorded at other sites due to insensitivities within the mire ecosystem, or a more localised effect, especially if the mires were responding to more spatially incoherent precipitation patterns (Barber *et al.*, 1999). Various phases of the LIA and MWP are well represented at the majority of sites, (the exception being Langlands Moss due to probably human interference in the last 1000 years of the record). However, there are relatively large temporal discrepancies between the LIA and MWP phases, possibly representing a more localised signal (Bradley and Jones, 1993). What is important to note is that the identification of these climatic fluctuations does provide evidence for a correlation of these proxy based records with independently derived climatic indices, suggesting the other climatic fluctuations within the proxy records represent significant changes and not just ecological noise.

7.2 Comparison and value of the three proxy records

The three proxy records have been shown to correlate well over the last *ca.* 5000 years (*ca.* 7500 years at Temple Hill Moss), providing high resolution reconstructions of Holocene climatic change in Scotland. There are, however, differences between the records, relating to the magnitude of events, notably for the non-linear reconstructions (DCA reconstructions and humification), whereas the modelling of the testate amoebae transfer function produces linear reconstructions, from which the magnitude of change can be compared between sites (Charman *et al.*, 1999). *Caveats*, however, do exist, which are discussed below.

7.2.1 Comparison of plant macrofossil DCA

The DCA plots for each site on which analyses were undertaken are shown in Figure 7.6. What is important to note from the reconstructions shown in Figure 7.6 is the different scales from the non-linear models. The sites with most variability are Temple Hill Moss and Shirgarton Moss which show axis 1 scores plotted over a range of 4 s.d. units, suggesting a complete species turnover in composition of samples (ter Braak and Looman, 1995), and raising the sensitivity of the model. Although there appear to be large fluctuations in the DCA from Langlands Moss, the range of changes in sample composition is low, as only just over 2 s.d. units are represented. There also appears to be a greater sensitivity of the southwestern sites (notably Longbridge Moss and Langlands Moss) compared with the more northern sites, which is possibly related to the dominance of a single species (either *S. imbricatum* or *S.*s. *Acutifolia*) in the more northern sites. The greatest magnitude of change, however, occurs at Temple Hill Moss between *ca.* cal. 1450-1250 BC, which is represented on the DCA axis 1 scores by a change of 3 s.d. units. This phase is represented by a stratigraphy dominated by *S.*s. *Cuspidata* being replaced by *S.*s. *Acutifolia*, and eventually *S. imbricatum* (Figure 6.22). Changes of a similar magnitude can be seen at Longbridge Moss in the large fluctuations associated with hollow/pool taxa, and at Mallachie Moss when the UOM/Monocot layer is superseded by Sphagna around 28cm depth (Figure 6.40). As the ordinations are non-linear, however, it is difficult to suggest whether these phases are the most extreme climatic fluctuations from the individual records.

Figure 7.6 Plant macrofossil DCA reconstructions for all sites. All scales are axis 1 scores in s.d. units x100.

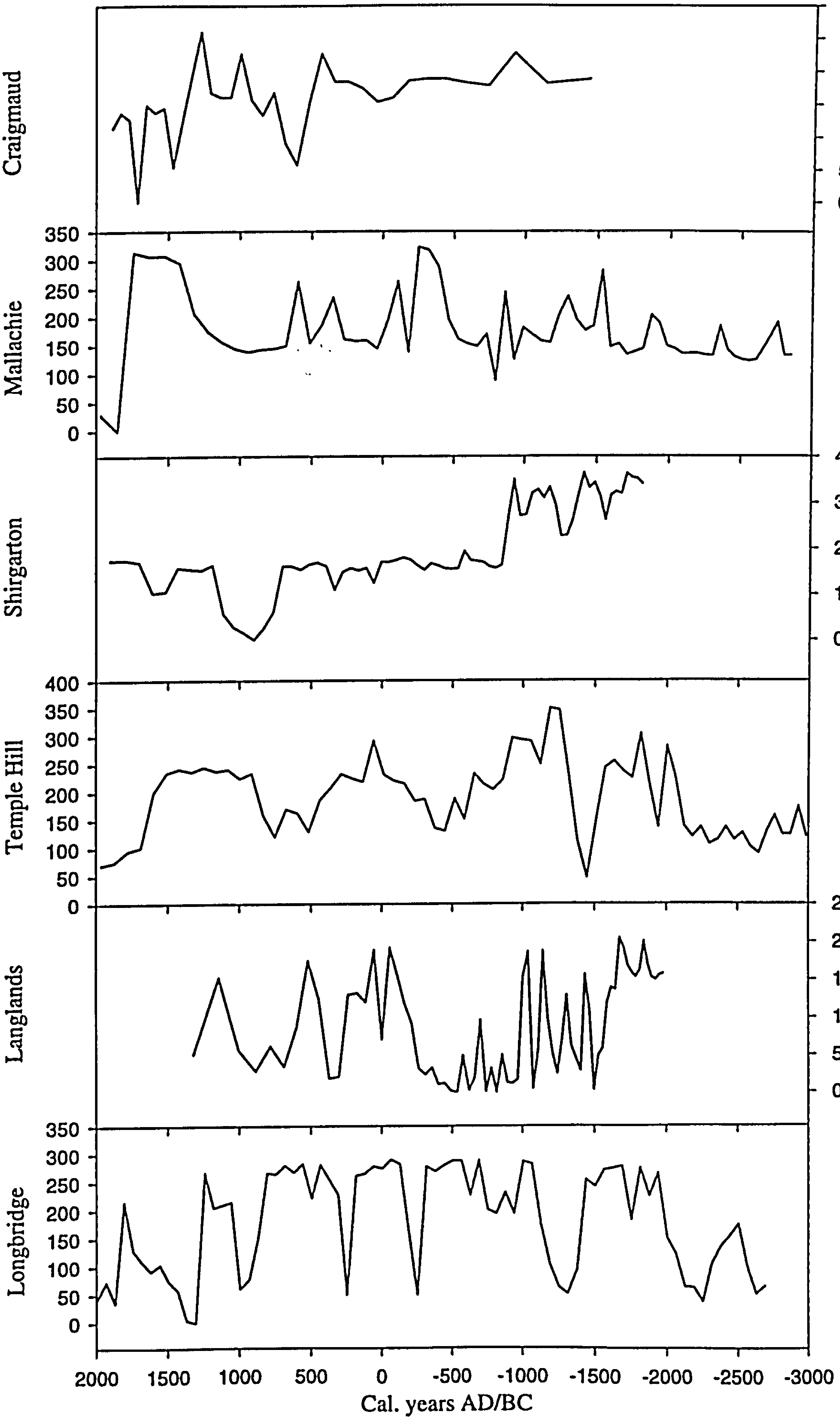


Figure 7.7 Humification plots for all sites. All scales are the residuals from the detrended absorbance values.

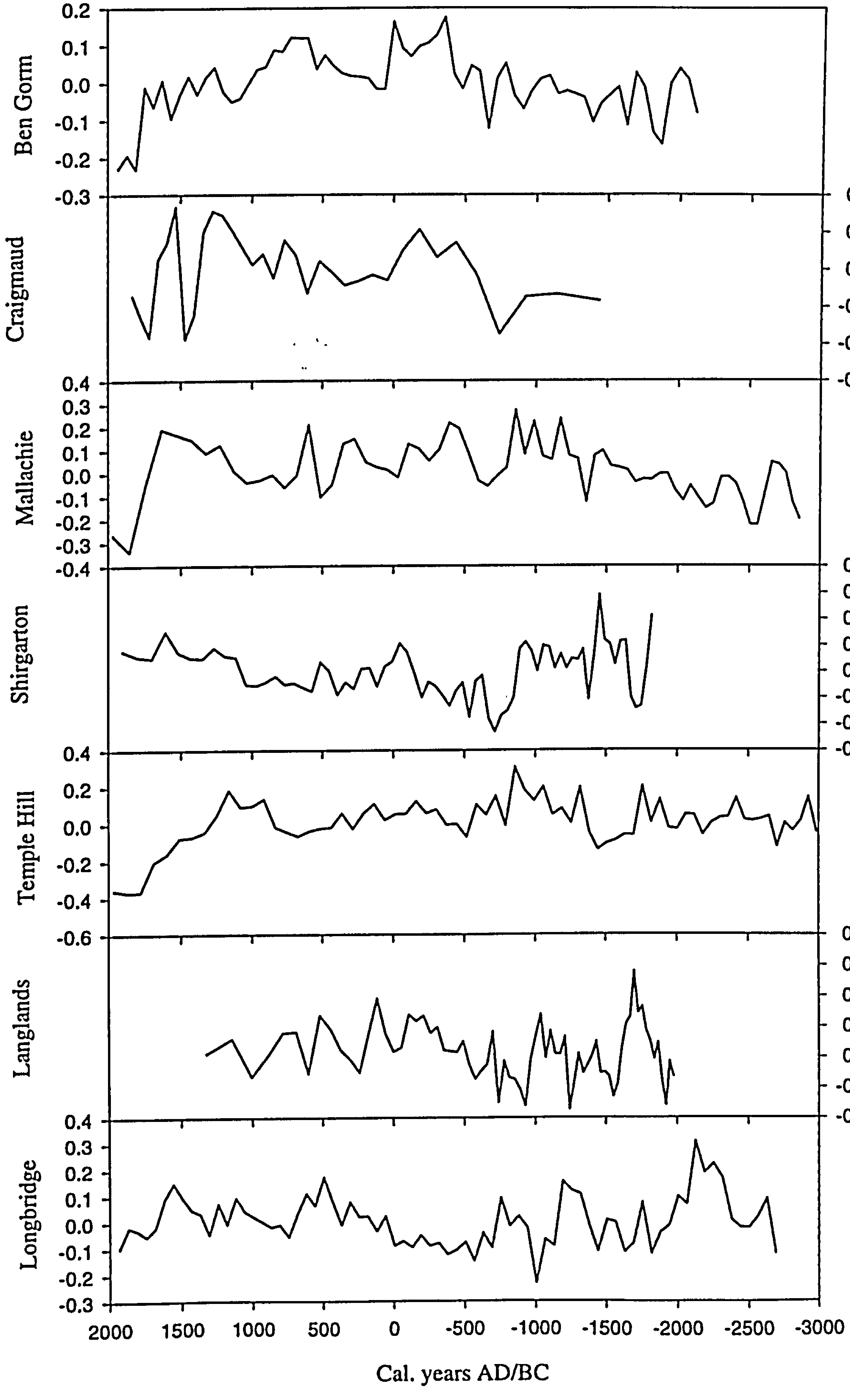


Figure 7.8 Reconstructed water table depths (mean values) from testate amoebae data for all sites. All scales are in cms.

In general the DCA ordinations have been shown to model the changing taxa from the plant macrofossil records accurately, as the moisture gradients from each ordination agree with taxonomic inferences, and the data correspond well with the other proxy reconstructions. It may be suggested, therefore, that the data from similar sites could be modelled using the same DCA, and if similar climatic changes are shown by both sites from the same ordination then the taxa (and sites) are reacting to the same forcing mechanisms. The initial problem with this hypothesis was that there were no two sites which appeared to have similar climatic controls. The present day climatic data (Chapter 4) suggested that although some sites may have similar precipitation regimes, thermal differences were present. However, the sites in southern/central Scotland were tested, as these four sites (Longbridge Moss, Langlands Moss, Temple Hill Moss and Shirgarton Moss) all contained similar taxa, notably *Sphagnum imbricatum*, which was absent from the more northern sites.

The ordination of the four sites produced axis 1 scores which did not represent a moisture gradient, although the axis 2 scores were more indicative of a wetness gradient. It is possible that the axis 1 scores were responding to intrasite environmental variability. The axis 2 scores, however, did not differ greatly from the original individual DCA's. The main change was in the magnitude of specific phases, which necessitates the use of other proxies to identify whether the phases are significant. Ordinations were also attempted using just two sites, although the resulting reconstructions were always very similar to one of the original individual DCA's, while the variability was often damped in the other record. The main problem was with *Sphagnum imbricatum* and *Sphagnum* section *Acutifolia* due to their moisture tolerances varying between sites. This approach to DCA modelling was therefore thought not to be particularly useful, as the resulting ordinations were less sensitive to climatic changes than the original models. It can be argued that the composite approach treats all the data in the same way, but the evidence presented here suggests that the species/sample scores are not comparable between sites and some of the valuable information from the individual plant macrofossil records can be lost using this approach. The original data has thus been retained for this discussion.

7.2.2 Comparison of humification records

The humification records for all seven sites can be seen in Figure 7.7. There are some coherent phases of change between sites, although there are also some notable discrepancies. One example at *ca.* cal. 1750 BC shows Langlands Moss and Ben Gorm Moss to be in a dry climatic phase, while Shirgarton Moss is clearly experiencing a wet phase. This could be due to inaccurate chronologies, which is where the precision of tephra isochrones is vital, as described above. It could also represent real differences between the sites, possibly reacting to non-coherent precipitation regimes (Barber *et al.*, 1999).

It is encouraging to note that the humification reconstructions display variability when the stratigraphy is dominated by a single taxon for considerable periods of time, enabling high resolution palaeoclimatic reconstructions to be produced. A good example can be found at Shirgarton Moss, where between 200-130cm the stratigraphy is dominated almost exclusively by *Sphagnum imbricatum*. During this period (*ca.* cal. 850-50 BC) there are numerous fluctuations in the humification record, suggesting that the humification signal is independent of fossil plant species compositions. Other evidence, however, suggests that the humification curve may respond to specific macrofossils, as documented at Longbridge Moss (section 6.5.7). Large variations in magnitude can be seen in all the fluctuations from the humification record, although the largest changes do not always correspond with the biggest changes in the DCA records, reflecting the non-linearity of both reconstructions. Notable exceptions, however, are the wet phase at Langlands Moss around *ca.* cal. 1560 BC, and the wet phase around *ca.* cal. 1470 BC at Craigmaud Moss, which reflect the highest magnitude of change at both sites for both the DCA and humification records.

7.2.3 Comparison of testate amoebae reconstructions

The testate amoebae mean water table reconstructions can be seen in Figure 7.8 for each site on which the analyses were undertaken. What is immediately apparent from Figure 7.8 is the low variability recorded at Longbridge Moss, with mean reconstructed water table depths fluctuating between only 0 to –7.5cm, compared with the other sites which display much more variability. This is due to the dominant taxa within the assemblages at each of the sites, and the related modelling of each taxon by the transfer function. As

Longbridge Moss is dominated by *Difflugia pulex*, which is poorly represented in the transfer function (see section 6.6.4), the reconstructed water table values are damped. The modelling of *Difflugia pulex* appears to be complex, and more research into modern samples and fossil assemblages is needed. Hendon (1998) suggested that they have a cosmopolitan distribution, but that in ordination plots they occur with drier taxa, specifically *Hyalosphenia subflava*. However, data from the five sites studied here suggests that they can also occur with relatively high abundances of *Amphitrema flavum*. At Langlands Moss, where *Difflugia pulex* is abundant, it is generally replaced by *Hyalosphenia subflava*, rather than being associated with it. Thus, maybe the original cosmopolitan distribution of *Difflugia pulex*, as suggested by Hendon (1998) is accurate, although it may also have a bimodal distribution similar to *Sphagnum imbricatum*.

Other sites, although they might be dominated by *Difflugia pulex*, have high abundances of other 'indicator' species for which the ecologies are well known (Table 7.3), which define specific high or low water tables, and generate a more sensitive climatic record.

Table 7.3 The dominant taxa from testate amoebae analysis at each site.

Site	Dominant species
Longbridge Moss	*Difflugia pulex, Amphitrema flavum*
Langlands Moss	*Difflugia pulex, Hyalosphenia subflava, Amphitrema flavum*
Temple Hill Moss	*Amphitrema flavum, Hyalosphenia subflava*
Shirgarton Moss	*Amphitrema wrightianum, Amphitrema flavum, Difflugia pulex*
Mallachie Moss	*Amphitrema flavum*

The magnitudes of change within the water table depth reconstructions appear to be largest when the assemblages change from containing relatively hygrophilous taxa to one dominated by *Hyalosphenia subflava*, often by as much as a 16cm drop. This may reflect the poor modelling of this species, which is discussed in detail in Hendon (1998). However, this large magnitude of change is subject to the large error margins on the transfer function. Much of the variability of the reconstructed water table depths, especially from Longbridge Moss, is within the sample prediction errors. These error margins need to be reduced before any precise correlations can be undertaken between sites (this is currently being addressed: D. Charman, pers. comm.). This problem notwithstanding, the variation between sites can be considered from Figure 7.8. It is noticeable that the largest fluctuations from each site do not occur at the same times. This may be a function of site sensitivity, but is also related to the variable modelling between taxa, especially those which have poor modern analogues.

One of the problems with modelling water table depths from fossil testate amoebae assemblages is related to the preservation rates of fossil amoebae, and possible differential selectivity. Louiser and Parkinson (1981) found differential species preservation in mineral soils, and some evidence exists to suggest that it may also occur in certain peats (Steinecke, 1927; Tolonen, 1968 - both described in Charman, 1999). This potential problem may affect species richness, which in turn could affect a water table depth reconstruction. This problem is also related to taxonomy, as different schemes will relate directly to species diversity. This is discussed in detail by Charman (1999). The biodiversity of the five cores varies between sites, with the lowest average number of species per sample being Mallachie Moss (10.6 in a range from 12.9-10.6), which correspondingly has generally low levels of variability from the reconstructed water table depths after 0 AD/BC when biodiversity is at its lowest. The site with the highest average number of species, Longbridge Moss, correspondingly shows a high degree of variability but over a narrow range of reconstructed water table depths, probably due to the influence of *Difflugia pulex*, as discussed above. It is therefore conceivable that the taxonomic system used, coupled with the derived transfer function from a specific taxonomy, is influential in the resulting modelling of fossil assemblages (*cf.* Charman, 1999). These problems notwithstanding, the testate amoebae water table depth reconstructions agree very well with the other proxy records, and together form the basis for detailed palaeoclimatic reconstructions.

7.2.4 Summary of the three proxies

The preceding discussion recognised the importance of all three palaeoclimatic proxy reconstructions. Without the combined results of three methods, the exercise would have been more ambiguous, as it is possible to draw a straight line between two points, but not between three. There is thus not one *best* proxy, although the importance of the plant macrofossil record cannot be overstated. There are certainly problems with using the macrofossil record alone for palaeoclimatic reconstructions, as discussed above, although each of the other proxy reconstructions always needs to be compared with the plant macrofossils. The methods behind peatland palaeoclimatic reconstructions have come a long way since Barber (1981), notably with respect to the data modelling, and the climatic reconstructions can now be inferred with increasing confidence.

7.3 The role of climate change in species replacement

Two species are discussed, the role of *Sphagnum imbricatum* in palaeopeatlands from the plant macrofossil record, and the decline in *Amphitrema wrightianum* from the testate amoebae record in response to Woodland's (1996) comments.

7.3.1 The role of *Sphagnum imbricatum* in peatland development and climatic change

Sphagnum imbricatum was found at four of the sites studied, all of which are in the southern part of Scotland. The lack of *S. imbricatum* at the northern and more easterly sites reflects its oceanic distribution, although it has been located on a small raised bog near Mallachie Moss (M. Proctor, pers. comm.), which would make an interesting stratigraphical comparison between the two sites. The abundance of *S. imbricatum* can be compared between sites, in order to reveal patterns of dominance, fluctuations in growth, and eventual extinction (Figure 7.9 and Table 7.4).

Figure 7.9 *Sphagnum imbricatum* percentages against time for all sites where the species was identified.

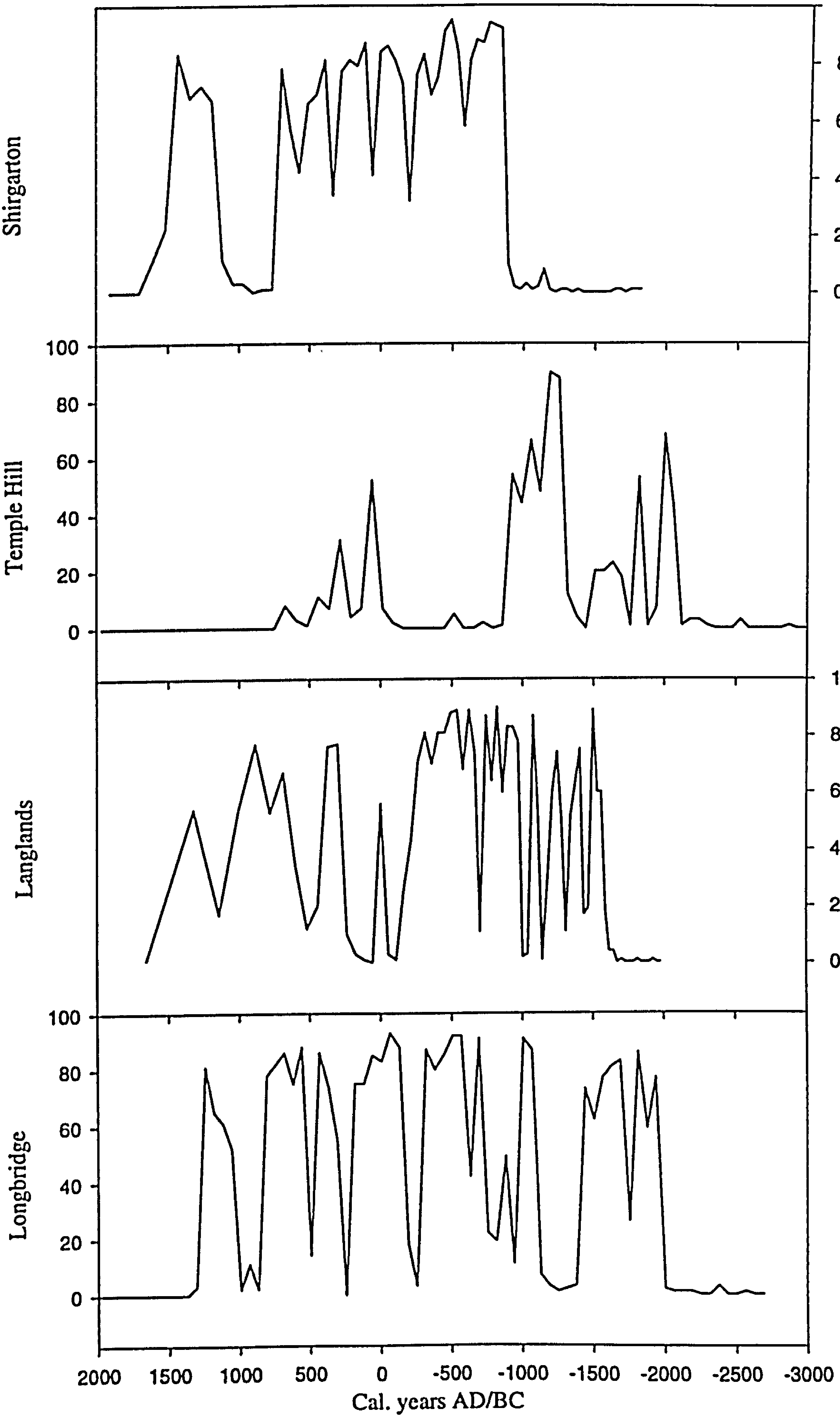


Table 7.4 The dates of *Sphagnum imbricatum* growth and extinction from four sites in southern and central Scotland. All dates are individual radiocarbon dates (2 sigma range), see Table 5.17 for details.

Site	Start of growth (cal. BC)	Extinction (cal. AD)	Growth time (years)
Longbridge Moss	2130-1780	1030-1230	2810-3360
Langlands Moss	1670-1420	890-1030	2310-2700
Temple Hill Moss	2890-2600	20-230	2620-3120
Shirgarton Moss	910-760	1220-1390	1980-2300

The onset of growth occurs at different times at all sites, which has also been found in other Cumbrian and Border mires (e.g. Stoneman, 1993; Mauquoy and Barber, 1999a). *S. imbricatum* is only present for three short periods of time at Temple Hill Moss, with the decline of the second phase of growth at *ca.* cal. 850 BC being attributed to a climatic deterioration, according to the humification and reconstructed mean water table depths (see section 7.1.1). The onset of *S. imbricatum* growth at Shirgarton Moss occurs at the same time (Figure 7.9), and is also attributed to a climatic deterioration from the humification and reconstructed water table data. These two separate events, initiated by the same climatic deterioration, highlight the complex nature of this species, which is related to its ability to grow under a range of moisture conditions.

There are numerous variations in dominance of *S. imbricatum* at each of the sites in Figure 7.9, ranging from the short lived phases at Temple Hill Moss, to the almost complete dominance at Shirgarton Moss. At Shirgarton *S. imbricatum* rarely drops below 40% in abundance once it has established its presence on the mire, apart from a major decline between *ca.* cal. AD 770-1120 where *S. papillosum* and *S.*s. *Cuspidata* replaced *S. imbricatum*, due to a climatic deterioration, as recorded by all three proxies. Variations in dominance are greater at the other three sites. The pattern of *S. imbricatum* supremacy at Langlands Moss fluctuates with the competition of *S.*s. *Acutifolia*, which appears to be cyclic between *ca.* cal. 1650-860 BC, and appears to be linked with climatic change according to the data from the three proxies (see Figure 6.20). A similar pattern relating to climatic changes can be seen in the *S. imbricatum* curve from Longbridge Moss, although it had to compete with *S. papillosum* and *S.*s. *Cuspidata* as well as *S.*s. *Acutifolia*.

The extinction of *Sphagnum imbricatum* from many peat bogs in Britain has long perplexed palaeoecologists. Various theories have been put forward regarding this event, as referred to in Section 6.1.20, the most popular of which are the climatic and anthropogenic (direct or indirect) mechanisms. The most recent paper on the decline of *Sphagnum imbricatum* (Mauquoy and Barber, 1999b) suggests climatic deterioration played a major role, following the ideas of Barber (1981) and Stoneman *et al.* (1993). Barber (1981) suggested that *S. imbricatum* may not have been able to adjust rapidly enough from the compact hummock ecad to the lax lawn ecad in response to climatic forcing and interspecific competition on Bolton Fell Moss. Barber identified a change from *S. imbricatum* dominated communities to communities dominated by *S. magellanicum* via a S.s. *Cuspidata* or *S. papillosum* stage in 16 out of 21 profiles examined on Bolton Fell Moss. Similar results were presented by Stoneman (1993) for ombrotrophic mires in Cumbria and the Scottish Borders. Mauquoy and Barber (1999b) suggested that specific wet shifts were associated with the decline of *S. imbricatum* from four ombrotrophic mires in Cumbria (Raeburn Flow, Bell's Flow, Coom Rigg Moss, and Felecia Moss). They suggested interspecific competition was responsible for the decline at Bolton Fell Moss and Walton Moss, as *S. imbricatum* appeared to have been growing on a hummock microform which became colonised by *S. magellanicum*, which was already dominant in the surrounding lawns and hollows. The decline in *S. imbricatum* at the six bogs studied by Mauquoy and Barber (1999b) spanned the time period cal. AD 1030-1460 (2 sigma age ranges), which agrees with the dates from Longbridge Moss and Shirgarton Moss (Table 7.4). The date at Langlands Moss (cal. AD 890-1030: SRR-6331) is at the edge of this range, and is not inconsistent with an earlier decline at Felicia Moss, dated to cal. AD 670-880 (Mauquoy and Barber, 1999b). The decline at Temple Hill Moss (cal. AD 20-230: SRR-6336) obviously occurs much earlier than at other sites.

The plant macrofossil records from the four sites in this thesis help to explain the disappearance of *S. imbricatum*, although evidence is also presented from the humification and reconstructed water table depths from testate amoebae (*cf.* Mauquoy and Barber, 1999b). Longbridge Moss and Shirgarton Moss show that *S. imbricatum* was replaced initially by *S. papillosum*, which was then either partly or totally replaced by *S. magellanicum* respectively. At Temple Hill Moss, *S. imbricatum* was replaced by

S.s. Acutifolia, and at Langlands Moss, *S. imbricatum* was replaced by *S.s. Acutifolia* and *S.s. Cuspidata*. The high levels of *S.s. Acutifolia* may indicate the swamping of *S. imbricatum*, as Boatman (1983) found that *S. capillifolium* var. *capillifolium* could grow 12cm above a lawn microform in four years.

A common feature of the *S. imbricatum* record at the four sites in this thesis is a two stage decline. After the initial decline in *S. imbricatum* at Shirgarton Moss at *ca.* cal. AD 770, it was able to re-establish itself at *ca.* cal. AD 1120, until its eventual demise around cal. AD 1220-1390 (SRR-6341). This two stage decline can be seen at each of the other sites in Figures 7.10(a-d), which also include the changes in humification and *Amphitrema wrightianum* abundance, a hydrophilous testate which also becomes extinct towards the surface (see section 7.3.2).

A high abundance of *S. imbricatum* existed at Longbridge Moss until *ca.* cal. AD 800 (Figure 7.10a), when they declined dramatically, being replaced by *S. papillosum*. Other evidence suggests a climatic deterioration was responsible for this, as levels of humification were low and the hydrophilous *Amphitrema wrightianum* were abundant. The levels of humification, however, decreased before the *S. imbricatum* decline, possibly implying that *S. imbricatum* was able to withstand a certain amount of moisture increase, until it was outcompeted by the faster growing *S. papillosum* (*cf.* Flatberg, 1986). By *ca.* cal. AD 1000 the levels of *A. wrightianum* had declined, and humification had increased, enabling *S. imbricatum* to become abundant once more. However, *S. imbricatum* only lasted another *ca.* 300 years at Longbridge, as it was finally replaced by *S. papillosum*, which coincided with lowered levels of humification and an increase in abundance of *A. wrightianum*, suggesting another climatic deterioration. *S. magellanicum* colonised the mire shortly after this final decline, and probably exploited the vacant niche left behind by *S. imbricatum*, ensuring that it could not return. Both the humification and testate amoebae records suggest that the last two phases of *S. imbricatum* growth at Longbridge Moss were of the compact hummock form. Humification levels were high, and the testates were dominated by *Difflugia pulex*, which has been shown to be associated with more xerophilous species (Hendon, 1998). A second core was analysed (Langdon, unpublished results) from Longbridge

Figure 7.10a Longbridge Moss: The extinction of *Sphagnum imbricatum*, showing levels of *S. imbricatum*, humification, and *Amphitrema wrightianum*.

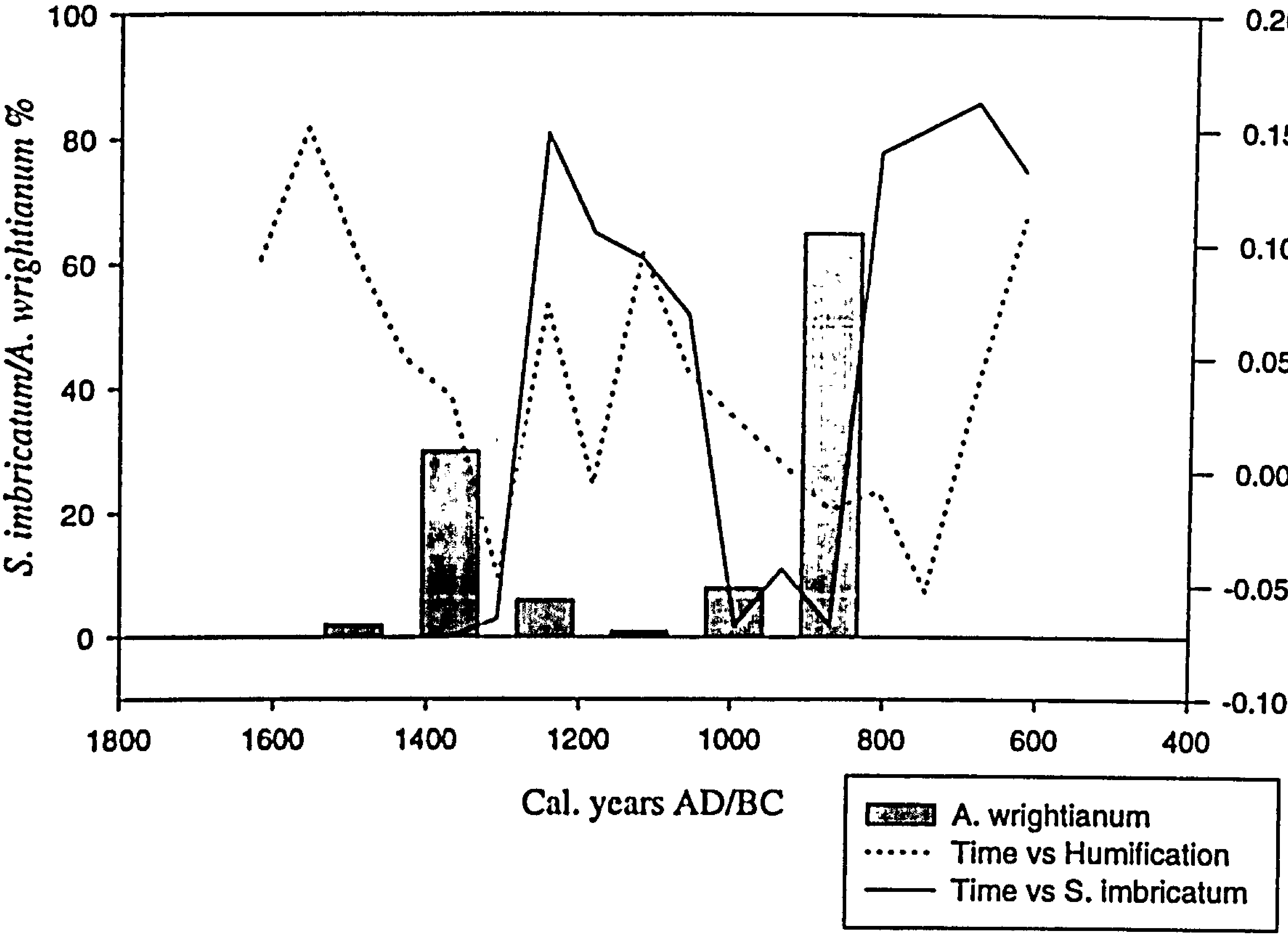


Figure 7.10b Langlands Moss: The extinction of *Sphagnum imbricatum*, showing levels of *S. imbricatum*, humification and *Amphitrema wrightianum*.

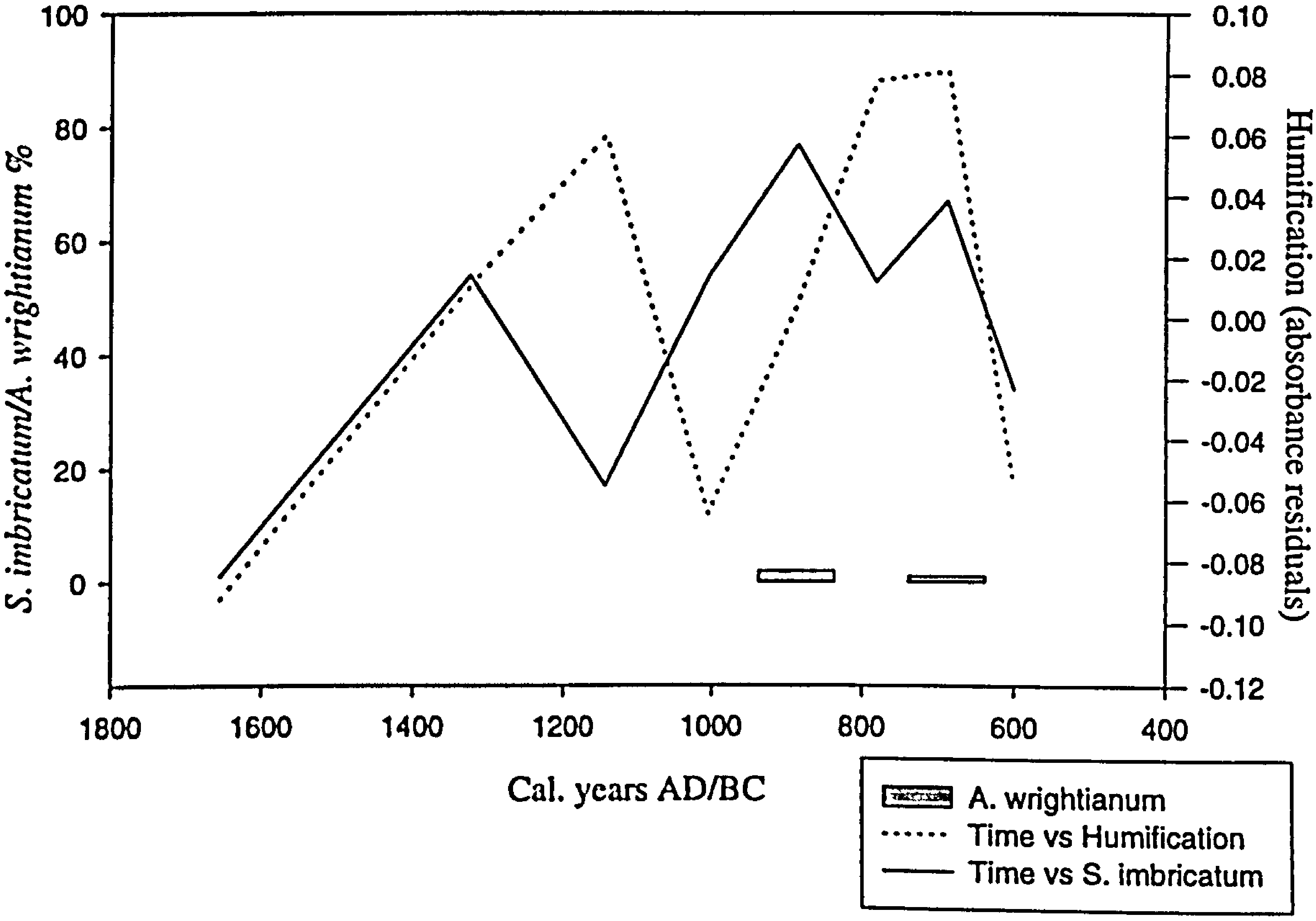


Figure 7.10c Temple Hill Moss: The extinction of *Sphagnum imbricatum*, showing levels of *S. imbricatum*, humification, and *Amphitrema wrightianum*.

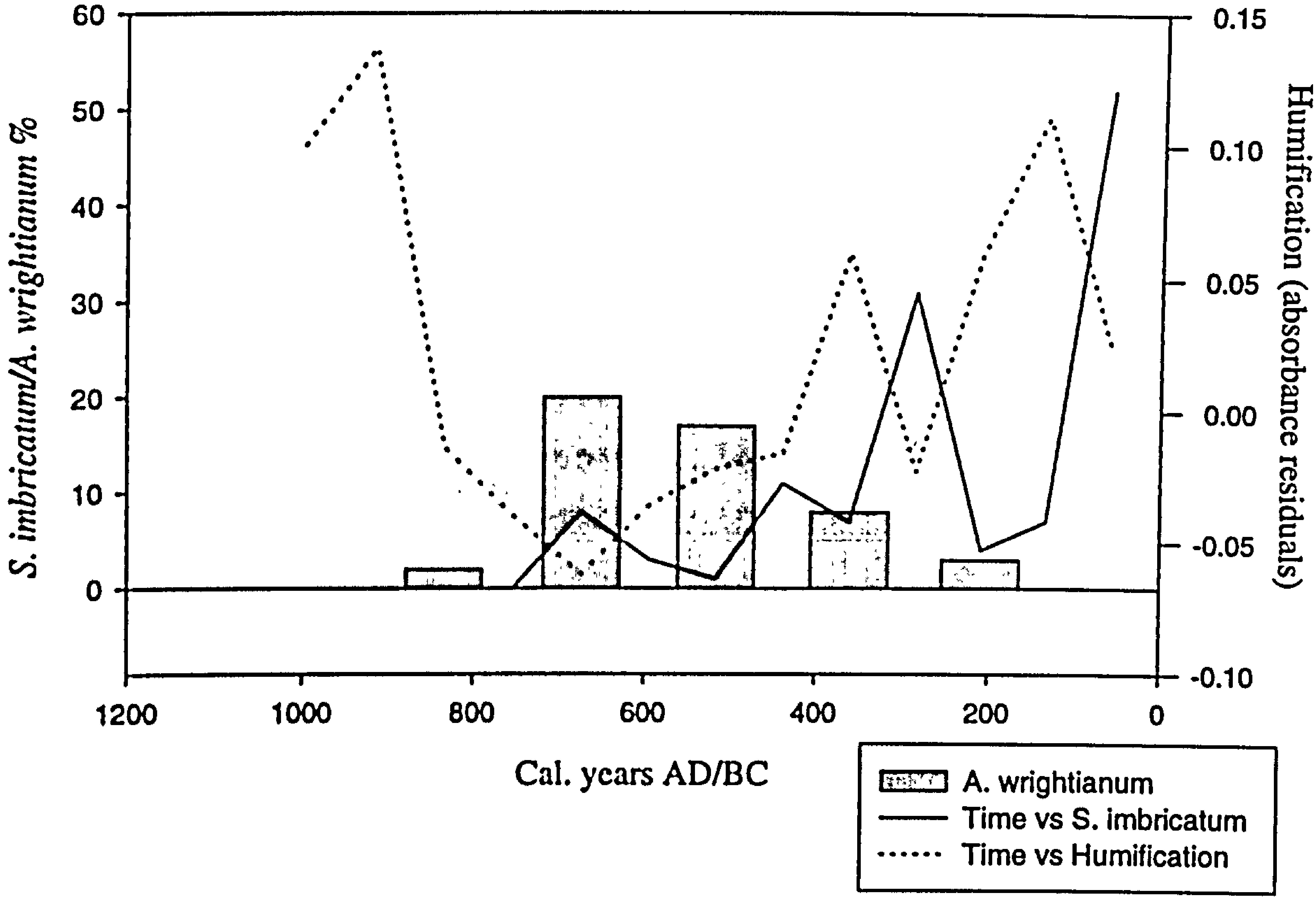
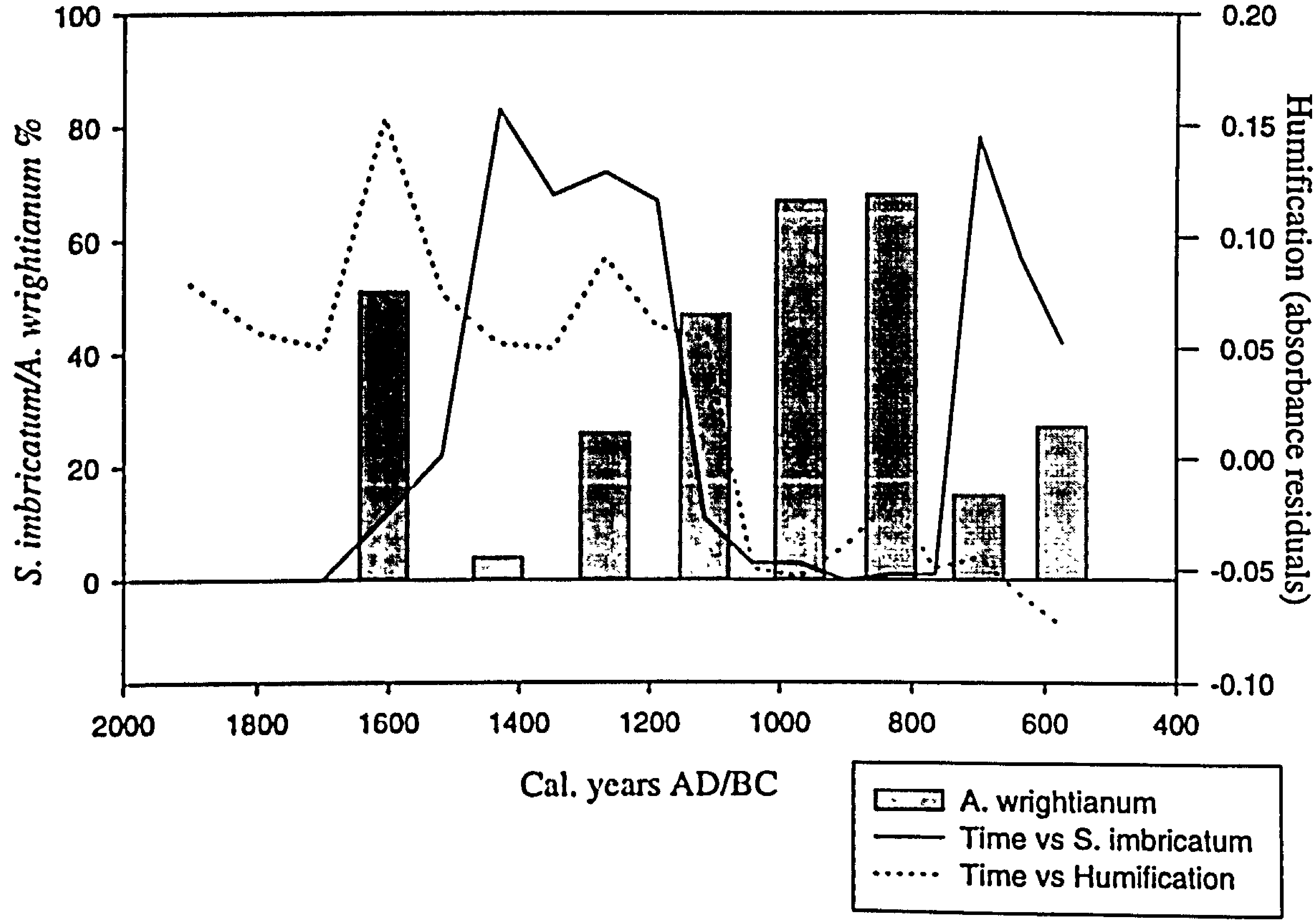


Figure 7.10d Shirgarton Moss: The extinction of *Sphagnum imbricatum*, showing levels of *S. imbricatum*, humification, and *Amphitrema wrightianum*.



Moss which also showed a two stage *S. imbricatum* decline, with the final phase of growth being restricted to the compact hummock form, as it coincided with a phase of *Hyalosphenia subflava* dominance.

The *S. imbricatum* decline at Shirgarton Moss is similar to that at Longbridge Moss, with notably high levels of *A. wrightianum* present when *S. imbricatum* declines (Figure 7.10d).

The decline of *S. imbricatum* at Langlands Moss is shown in Figure 7.10b. The *S. imbricatum* curve follows the humification fluctuations, although there is a lag in the first decline of *S. imbricatum*, similar to Longbridge Moss. Unfortunately very low concentrations of testates at the first *S. imbricatum* decline meant that no sampling could be undertaken. The climatic inferences thus only rely on the humification and plant macrofossil records, although they do suggest a climatic deterioration at the time of local extinction for *S. imbricatum*.

The pattern of local extinction of *S. imbricatum* at Temple Hill Moss shows an initial decline between *ca.* cal. AD 100-200, being replaced by *S*.s. *Acutifolia*, and is associated with high levels of humification (Figure 7.10c), indicating a climatic shift towards dryness or increased continentality, as was suggested by Godwin and Conway (1939). *S. imbricatum* then recovered, and humification levels declined, although low levels of humification, and a steady increase in the hydrophilous *A. wrightianum* suggest a climatic deterioration occurred which ultimately led to the local extinction of *S. imbricatum*.

Other theories put forward to explain the widespread decline of *S. imbricatum* have been reviewed by Stoneman *et al.* (1993) and Mauquoy and Barber (1999b). Mauquoy and Barber suggested that the theory of deposition of agriculturally derived dust and charcoal (van Geel and Middeldorp, 1988) is unlikely to explain the phenomenon, as hummocks retain more deposited elements than hollows (Clymo *et al.*, 1990), which should lead to *S. imbricatum* on hummocks being replaced first as *S. magellanicum* is better adapted to grow in such conditions. Barber (1981), however, noted that *S. imbricatum* is found to persist on hummocks while becoming extinct on lawn microforms. Mauquoy and Barber (1999b) do, however, stress that the importance of nutrients in such situations is far from clear.

Burning, and other anthropogenic activities have also been suggested as being responsible for the decline of *S. imbricatum* (Pearsall, 1956; Piggot and Piggot, 1963). All the mires studied are structurally intact, with only peripheral cuttings and drainage ditches, with the possible exception of Langlands Moss (see section 5.3.2). Both Stoneman (1993) and Mauquoy and Barber (1999b) were unable to find a correlation between charcoal and the demise of *S. imbricatum*. Similarly, there is no such correlation at any of the sites in this thesis, with the possible exception of Longbridge Moss which has small amounts (2 on the 1-5 scale) of macroscopic charcoal at 48cm.

The balance of evidence therefore suggests that a climatic deterioration did occur at the time of local *S. imbricatum* extinctions. It also appears that the final phase of *S. imbricatum* growth was of the compact hummock ecad, certainly at Longbridge Moss and possibly at Shirgarton Moss, corroborating the ideas of Barber (1981). It is suggested that these climatic deteriorations, probably coupled with interspecific competition (notably *S. papillosum*) were responsible for the local extinctions of *S. imbricatum* at these four sites in Scotland.

7.3.2 The decline of *Amphitrema wrightianum*

Although present in fossil assemblages, *Amphitrema wrightianum* was not present in any of the surface moss assemblages studied by Woodland (1996), and is found to decline significantly towards the peat surface in other British mires (e.g. Mauquoy, 1997; Hendon, 1998; Charman *et al.*, 1999; McMullen, in prep). In Britain *A. wrightianum* has been found in bog pools, although was found to be less abundant than *A. flavum* (Heal, 1961), and has been found at low to intermediate depths in *Sphagnum recurvum* (Heal, 1962). *A. wrightianum* is only found in the wettest parts of southern Finnish mires (Tolonen *et al.*, 1992), which compare well with respect to the moisture content characterised by testate amoebae of British mires (Woodland, 1996).

Woodland (1996) stated that the reason for the *A. wrightianum* decline is uncertain, although she suggested that *A. flavum* may be outcompeting *A. wrightianum* in Britain as both species thrive in similar hydrological conditions (Heal, 1962; Tolonen *et al.*, 1992). Woodland (1996) and Hendon (1998) both noted that the decline of *A. wrightianum* occurred within the top 15cm of peat cores from Cumbrian sites, suggesting that the species demise may be due to an increased deposition of pollutants. Woodland concluded her observations on *A. wrightianum* by stating that further work was required on the timing of the decline of *A. wrightianum* and the replacement of *A. wrightianum* by *A. flavum*, as well as the present day spatial distribution of *A. wrightianum* with respect to industrial areas in Britain. The results from this thesis provide more detail on some of these issues. Present day samples were taken only from one site, Longbridge Moss, but no samples contained *A. wrightianum*, even those taken from pools.

The timing of the *A. wrightianum* decline can be seen in Table 7.5. Only four of the five sites which were subjected to testate amoebae analysis are represented (Longbridge Moss, Temple Hill Moss, Shirgarton Moss, and Mallachie Moss), as there were very few counts containing *A. wrightianum* at Langlands Moss (see Figure 6.16). Two of the sites (Temple Hill Moss and Mallachie Moss) show a main decline and then a later count which contained less than five individuals.

Table 7.5 The dates of the *Amphitrema wrightianum* decline from four sites in Scotland. All dates are interpolated from the age/depth models from Chapter 5.

Site	Main species replacing *Amphitrema wrightianum*	Date of *Amphitrema wrightianum* decline
Longbridge Moss	*Cyclopyxis arcelloides* type	AD 1370-1500
Temple Hill Moss	*Amphitrema flavum*, *Assulina muscorum*	AD 670-830; one occurrence at AD 1160
Shirgarton Moss	*Arcella discoides* type, *Difflugia pulex*	AD 1610-1800
Mallachie Moss	*Amphitrema flavum*	AD 1040-1230; one occurrence at AD 1850

It is immediately obvious from the testate amoebae diagrams in Chapter 6 that the *A. wrightianum* decline is not restricted to the top 15cm. The data in Table 7.5 show that the range of dates for the main decline of *A. wrightianum* is between *ca.* cal. AD 670-1800, suggesting that there may be different causes at each site. There are also different species which replace *A. wrightianum* at each site, although in the two instances when *A. flavum* displaces *A. wrightianum* (at Temple Hill Moss and Mallachie Moss), *A. flavum* is the dominant species at that site.

Mauquoy (1997) studied testate amoebae on seven sites in Cumbria, and although the interpolated dates of the decline of *A. wrightianum* were not recorded, most postdated *ca.* AD 1400. Most of the sites studied were dominated by assemblages of either *Amphitrema flavum*, *Amphitrema wrightianum*, *Arcella discoides* type, or *Hyalosphenia subflava*. The majority of sites recorded a decline in *A. wrightianum* in the top 30-40cm, although this depth was variable, with the decline occurring in the top 20cm at Bell's Flow, and *A. wrightianum* being present (at abundances greater than 5%) in the surface sample of Coom Rigg Moss. Charman *et al.* (1999) found that *A. wrightianum* was replaced by *A. flavum* in the top 15-20cm in the two other profiles from Coom Rigg Moss. The role of climate change in the decline of *A. wrightianum* is considered to be negligible, as most of the declines appear to have occurred in the LIA, when conditions should be suitable for *A. wrightianum*. On sites where the climate became drier, *A. wrightianum* may have declined as its optimum habitat is in bog pools. It could therefore be due to natural competition, possibly aided by changes in pH as a response to increased atmospheric pollution over the last few centuries, but more data is needed to confirm any of these hypotheses.

7.4 Chronologies and periodicities

The radiocarbon dates used in this project, assisted by tephras, SCP's, and pine pollen at specific sites, have provided a framework for the age/depth models used to interpolate dates throughout each sequence. The accumulation rates inferred from the dates for each sequence are clearly not precise, and will differ due to the variable rates of decay between hollow/pool and hummock microforms (Johnson *et al.*, 1990; Rochefort *et al.*, 1990; Johnson and Damman, 1991; Hogg, 1993). The bulk radiocarbon dates may also have individual problems, as an 8cm depth sample may actually be composed of a hollow and a hummock microform which can have very different accumulation rates, resulting in an ambiguous date. However, on the larger scale, the general rate of accumulation of bogs tends to remain the same over millennia (Clymo, 1991), and it is thus assumed that the majority of these high resolution variations can be averaged out to give generalised age/depth models.

The only way to possibly resolve the more subtle changes in accumulation rates would be to date large sections of the profile by 'wiggle matching', which would be an extremely costly exercise, although the precision of the data is likely to be improved. The analysis of SCP's was a useful technique at Shirgarton Moss for interpreting the age of the surface peats. Time precluded their use at other sites, but the increases in pine pollen at specific sites, even though the resolution was low, served as a useful indicator that the mires were intact.

The periodicities identified in section 6.13 can be subjected to interpretation, with the assumption that any errors in the chronological controls are comparatively small. The periodicities identified from five of the sites (no significant periodicities were discovered at Shirgarton Moss and Ben Gorm Moss) can be seen in Table 7.6.

Table 7.6 The periodicities in years discovered from spectral analyses of the humification and DCA axis 1 scores at five raised bogs in Scotland.

Site	DCA axis 1 scores	Humification
Longbridge Moss	520	470
Langlands Moss	580, 316	575, 370, 315
Temple Hill Moss	1100	1100
Mallachie Moss	920, 120, 100	365
Craigmaud Moss	280	400, 350, 280

It is apparent from Table 7.6 that both the DCA axis 1 scores and humification data can reveal more than one periodicity, reflecting the variability within each proxy reconstruction. What is most encouraging to note is that some of the periodicities can be seen in both records, emphasising an independent check on the period, and suggesting that it is not random. The periodicity at Craigmaud Moss of 280 years is very similar to the periodicity identified by Aaby (1976), who documented a cycle of 260 years from raised mires in Denmark. Aaby also identified a double period of 520 years in his data, which matches significant periods around 520 years and 580 years from Longbridge Moss and Langlands Moss. Therefore, it is possible that evidence is present for the same forcing mechanism at each of these sites, as well as in Denmark, and may thus be related to oceanic forcing controlling the climate over the Atlantic which affects northwest Europe (*cf.* Barber *et al.*, 1994b).

Another interesting periodicity shown in Table 7.6 is the 316 year cycle, identified in the DCA axis 1 scores and humification records from Langlands Moss, which may be related to the 260 years cycle found by Aaby. However, one of the most significant periods is the 1100 year cycle identified at Temple Hill Moss, which has also been identified at Walton Moss, Cumbria by Hughes *et al.* (in press) and may be connected with the Bond cycles (Bond *et al.*, 1997). Most of the other sites registered a millennial scale cycle, although the length of each record (with the exception of Temple Hill Moss) was only of the order of 4000-5000 years, and thus these cycles cannot be considered significant until the palaeoecological records are extended to include at least 7 complete cycles. However, the 1100 year cycle has now been identified in two of the longest records analysed by the Southampton group (Temple Hill Moss, and Walton Moss – see Hughes *et al.*, In Press), highlighting the need for further analyses of other long terrestrial sequences.

7.4.1 Oceanicity vs continentality: site sensitivity

One of the potentially interesting areas for site comparisons in this project is between the bogs situated in the west, and those in the east, possibly highlighting differences in oceanicity or continentality. Birse (1971) divided Scotland into three major zones, based on thermal characteristics, which were essentially as measure of degrees of oceanicity in Scotland (as described in Chapter 4). Oceanicity can have an affect on the sensitivity of species, as instead of enhancing species diversity through amelioration of potentially stressful environments, it tends to create habitats which are hostile to many species, often as a direct response to the reduced evaporative power of oceanic climates (Crawford, 1997).

There has been a suggestion earlier in the discussion (from the DCA analyses, section 7.2.1) that the southwestern sites are more sensitive to climatic changes than the more northern sites. Sensitivity is defined here as the number of climatic changes registered by each mire. There is certainly a loss of sensitivity at Craigmaud Moss before

0 AD/BC, probably as a result of Sub-boreal dryness (Godwin, 1954). It is also noticeable that Temple Hill Moss (one of the more eastern sites) reacts at the same time as most other sites at some climatic deteriorations, although there are also times when many other sites display a wet shift and Temple Hill Moss does not (see Table 7.1). This may well be related to site sensitivity and geographical location.

The sites which have present day summer water deficits (Chapter 4), notably Longbridge Moss and Craigmaud Moss, do not appear to be any less sensitive than the other sites. Conversely, the site with the highest effective precipitation by over 100% compared to the other sites, Ben Gorm Moss, displayed a humification record which appears to be as sensitive to climatic change as any of the other humification records. It was originally thought that this hyperoceanic environment may be insensitive to climatic changes, although this research, and the work of Anderson (1998) suggests otherwise. There is thus the potential for a large number of water shedding blanket mires in western Scotland to be analysed using the above methods.

Chapter 8. Conclusions and future research

- A multiproxy approach is essential in order to determine climatic changes from peat stratigraphy. Plant macrofossil, colorimetric humification, and testate amoebae analyses have enabled clearly defined climatic changes to be identified from seven ombrotrophic bogs in Scotland.

- The multiproxy evidence must be examined in detail, and in combination, in order to ascertain common periods of wetter/cooler or drier/warmer climates. A synthesis of site data is essential as individual proxies can be misleading. It is possible for the DCA to model taxa ambiguously; there can occasionally be a species signal in humification data, and the transfer function for testate amoebae has poor modern analogues for certain species, tending to represent some xerophilous species inadequately. However, a multiproxy approach reduces the uncertainty in any interpretation.

- Significant wet phases were identified from a synthesis of the three proxy records at the majority of sites around *ca.* cal. AD 1350-1550, AD 870-1150, AD 500-600, 750-940 BC, 1240-1390 BC, 1450-1630 BC, 1700-1900 BC, and 2200-2300 BC. Additional wet phases were identified at Temple Hill Moss around *ca.* cal. 3350 BC, 3900 BC, and *ca.* cal. 4700 BC. Significant dry phases could be identified around *ca.* cal. AD 920-1340, AD 350-520, and *ca.* cal. 540-390 BC, before which the dry phases become more site specific. Although these dates suggest a broad pattern across the sites, there are northern/southern, and eastern/western differences between the climatic reconstructions. The best illustration of these differences from the palaeoclimatic reconstructions comes from Temple Hill Moss, which fails to register some of the wet phases identified at many of the other sites. These differences may be due to chronological inconsistencies, but are more likely to be the result of differential forcing mechanisms.

- The ability of tephra isochrones to provide precise correlations between sites, has enabled the direct comparison of proxy climatic records, which radiocarbon dating is unable to do. The correlations at the time of the Glen Garry and Hekla-4 isochrones revealed a significant difference between climatic phases in the north and south, and west and east respectively, most notably asynchronous changes between northern and southern Scotland when the Glen Garry isochrones was deposited. It is possible that the comparatively wetter north of Scotland at the time of the Glen Garry tephra may be due to subtle fluctuations in the polar front.

- Twelve sites were initially analysed for tephra content, with fourteen isochrones being geochemically typed from seven of the sites. The most common isochrones were the Glen Garry tephra (which occurred at six out of seven sites), and the Hekla-4 isochrone (four out of seven sites). Two 'new' isochrones were discovered at Ben Gorm Moss, and the identification of the AD 860 layer at Langlands Moss (previously unidentified in Scotland from geochemical typing) offers potential for precisely correlating historical climatic phases. The Lairg layer, which was discovered at Temple Hill Moss, has not been previously identified so far south. These conclusions enable inferences to be made relating to tephra dispersal. Patchy fallout patterns are noted, which is suggested to be related to wet deposition. This is particularly noticeable with the Hekla-4 fallout pattern as there appears to be a 'gap' in central Scotland where no Hekla-4 tephra has, as yet, been identified.

- . A climatic deterioration can be associated with the extinction of *Sphagnum imbricatum* at each site where *S. imbricatum* was present within the stratigraphy. The most pronounced local extinctions were at Longbridge Moss (cal. AD 1030-1230) and Shirgarton Moss (cal. AD 1220-1390), and occurred in two distinct phases, being initially replaced by *Sphagnum papillosum*. *S. imbricatum* was able to recover for a short period, and was then replaced at these sites again by *S. papillosum*. The establishment of *S. magellanicum* soon after the demise of *S. imbricatum* may have prevented the once dominant moss from becoming re-established. Testate amoebae and humification data were fundamental in accounting for the environmental circumstances surrounding the decline of *S. imbricatum*.

- Significant periodicities have been identified in the climatic reconstructions from the DCA axis 1 scores and the humification analyses. Longbridge Moss and Langlands Moss, both situated in southwest Scotland, exhibit 520-580 year periodicities from the two climatic reconstructions, which are in general agreement with a cycle identified in Danish peat bogs (Aaby, 1976). A cycle of roughly half the period (280 years) has been identified from the most northeasterly site, Craigmaud Moss, and a millennial scale cycle has been identified at a southeasterly site (Temple Hill Moss). This evidence therefore suggests that the same forcing mechanisms are affecting the sites in Scotland, as well as in Denmark, and may be related to northwest European oceanic forcing.

8.1 Future research

- There is still a need to investigate more sites in Scotland in order to corroborate these results. Lowe (1993), Birks (1996), and Whittington and Edwards (1997) have all commented on the lack of palaeoclimatological data from Scotland, despite the presence of numerous lakes and bogs. At present, more sites in Scotland are being analysed, and it is hoped that this trend will continue.

- More work needs to be done on producing a reliable tephrochronology for Scotland, following on from Dugmore *et al.* (1995a) but on a larger spatial scale. There are plenty of geochemical analyses for the well established isochrones, and it is now time to start analysing the data as a whole, and interpreting the patterns of dispersal. It should soon be possible to produce tephra maps, identifying the general areas where specific isochrones can be located.

- There is a distinct need for high precision wiggle match dating of individual isochrones to help pin down radiocarbon chronologies. The prime candidate in Scotland would be the Glen Garry tephra which has so far only been bulk radiocarbon dated at a few sites (Dugmore *et al.*, 1995a,b; Dugmore pers. comm.). Work in Iceland is currently ongoing with respect to Holocene tephras (e.g. Larsen *et al.*, 1999), and it is hoped that this will continue, in order that the sources for certain isochrones, e.g. Glen Garry and the AD 860 layer, may be located.

- Although the testate amoebae transfer function developed by Charman and colleagues at Plymouth University has provided an excellent independent record to compare with the other proxies, more work still needs to be done in updating better modern analogues for the drier taxa. There is also the possibility that a plant macrofossil transfer function could be established. Modern analogues for *S. imbricatum* exist in Britain (Daniels and Eddy, 1990; M. Proctor, pers. comm.), and the key of McMullen (1999) for splitting *S.*s. *Acutifolia* should prove valuable.

- The next critical test of the effective precipitation archive from peat bogs is to compare it with a proxy temperature record, in order to try and ascertain which forces are driving the palaeoclimatic reconstructions. Proxy temperature reconstructions can be obtained from a recently developed transfer function utilising chironomid assemblages from lake sediments (Lotter *et al.*, 1999; Olander *et al.*, 1999; Brooks and Birks, 1999). This work is currently in progress (NERC grant GR9/04313 at PLUS: Barber and Langdon, in prep).

- Future research into non-pollen microfossils, specifically fungal spores may reveal further evidence for changes in climate, following the pioneering work of van Geel (1978). Some work is current being undertaken with respect to these microfossils (J. Blackford, pers. comm.).

- More wiggle matched dating is necessary on many peat profiles, in order to ascertain rates of change within the climate system (*cf.* Pilcher, 1991). Since the technique was first described (van Geel and Mook, 1989), comparatively few studies have been undertaken using this approach. This may be a result of the high costs associated with the procedure, but this still appears the best method of attempting to precisely correlate specific events without the presence of any tephras.

- Computer modelling of Holocene climatic changes, as well as mire development pathways, will play an important role in defining future research. Modelling mire genesis is becoming increasingly important in understanding how these ecosystems respond to changes in effective precipitation, and the scale of change involved. The modelling of Holocene palaeoclimates has now reached a stage where the modellers need precise data from the palaeoecologists, for example in the form of mean July air temperatures. It is not yet known if these measurements will be attainable from peatlands. More should be understood when the results of the bog-lake proxy records become known, and the sensitivities of the peatland ecosystems can be tested against mean July air temperatures.

These are indeed exciting times for Holocene palaeoecologists, and the future research outlined in this section aims to highlight some of the problems, the questions being asked, and possible methods for solving them.

Appendix 1. Field stratigraphy

Key

cusp = *Sphagnum* section *Cuspidata*
erioph = *Eriophorum vaginatum*
rhynch = *Rhynchospora alba*
pap = *Sphagnum papillosum*
mag = *Sphagnum magellanicum*
imb = *Sphagnum imbricatum*

Longbridge Moss (LBM)

Stratigraphy sampled on 25/2/99

LBM1 2.50m from master core on bearing of 320^0

Depth (cm)	Troels-Smith	Humification
0-8	Tb4 (cusp)	2
8-22	Sh2 Tb1 Th1 Tl+ (algal pool mud)	6
22-29	Tb2 Sh1 Th1	5
29-122	Tb3 Th1 Sh+	4
122-125	Tb2 Sh2 Th+ (algal mud)	6
125-158	Tb3 (imb)Th1 (grading to Tb4 at lower end of unit)	3-4
158-164	Sh2 Tb1 Th1 (algal pool mud)	6
164-220	Tb4 (imb) Th+	4-6
220-237	Tb3 Th1 (Th4 crioph 225-228)	5
237-249	Sh2 Th2 (rhynch) pool	6
249-300	Tb3 Th1 (Th4 crioph 268-278)	4-5

LBM2 30m from master core on bearing of 320^0
Surface veg: *S. papillosum*, *S. capillifolium*

Depth (cm)	Troels-Smith	Humification
0-22	Tb4 (pap)	1-2
22-28	Tb4 (cusp) Th++	3
28-34	Sh2 Tb1 Th1 (algal mud)	4-5
34-200	Tb4 (imb) Th++ with Tb2 Sh2 (broken down imb) at 127-133, 159-161, 174-176, 198-200	4
200-217	Tb4 (imb)	4
217-223	Tb3 Sh1 Th+	5
223-253	Tb4	5
253-255	Sh2 Tb2	5-6
255-263	Tb3 Sh1 (pool)	3
263-282	Tb3 Sh1	5
282-300	Tb3 Sh1 Th+	4

LBM3 30m from master core on bearing of 140^0
Surface veg: *Narthecium ossifragum, O. sphagni, S. papillosum*

Depth (cm)	Troels-Smith	Humification
0-34	Tb3 Th1	4
34-40	Tb4 (cusp) Th++ (pool)	2-3
40-45	Sh3 Th1	5
45-100	Tb3 (imb) Th1 (Th grades top to bottom)	4
100-132	Tb4 (imb) Th+	4
132-134	Sh4 (pool)	6
164-169	Tb4 (imb) Th++	4/5
169-188	Tb4 (imb) Th+	4
188-190	Tb2 Sh2 (pool?)	5
190-200	Tb4 Th+ Tl+	4/5
200-218	Tb2 Sh1 Th1	5
218-229	Sh2 Tb2 Th+	5/6
229-246	Th2 (Erioph) Tb2 Sh++ Tl+	5
246-273	Tb3 Sh1 Th+ (rhynch)	5
273-288	Th3 (rhynch) Tb1 Sh++	4/5
288-300	Sh2 Th1 Tb1 Tl+ (bogbean present)	6

Langlands Moss (LAG)

Sampled on 26/2/99

LAG1 2.70m from master core on bearing of 270^0
Surface veg: *H. cupressiforme, S. rubellum, Polytrichum* spp., *O. sphagni*

Depth (cm)	Troels-Smith	Humification
0-12	Sh2 Tb2	6/7
12-22	Sh2 Th1 Tb1	3
22-32	Tb2 Th1 Sh1	4
32-53	Tb3 Th1 Sh+	4
53-82	Tb4 (imb) Th+ (humification grades 3-4 at 69cm)	3-4
82-92	Sh2 Tb2 Th+	4
92-100	Tb4 (imb) Th+	3/4
100-122	Tb3 Sh1 Th+	4
122-153	Tb3 Th1 (grading to Sh1 at 143cm)	3/4
153-168	Tb3 Th1 Tl+	4
168-200	Tb4 (imb) Th+ (erioph between 185-200)	4
200-222	Tb3 Th1 (erioph)	4
222-254	Tb3 (imb) Th1 (erioph) (243cm grades to H4)	3
255-269	Tb4 (imb) Th+	3
269-300	Tb3 Th1 Tl+	4

LAG2 30m from master core on bearing of 270^0
Surface veg: *S. capillifolium*

Depth (cm)	Troels-Smith	Humification
0-20	Tb2 Sh2	6
20-39	Tb2 Th1 Sh1	4
39-60	Tb3 (imb) Th1 (erioph around 50cm) Tl+	3
60-78	Tb2 Sh1 Th1 Tl+	4
78-91	Tb2 Th2 (erioph) Sh+	4/5
91-100	Tb4 (imb) Sh+	4
100-110	Tb3 Sh1	4
110-115	Tb4 (imb)	4
115-125	Tb3 Sh1 Tl+	4/5
125-133	Tb3 (imb) Th1 (erioph)	4
133-149	Tb2 Th1 Sh1	4
149-161	Tb4 (imb) Sh+ Tl+	4
161-164	Tb3 Sh1 (pool? cusp?)	3/4
164-200	Tb3 (imb) Sh1 Th+ Tl+	4/5
200-229	Tb4 (imb) Th+	3/4
229-231	Tb3 (imb) Th1 (erioph)	4
231-243	Tb3 Th1 Sh+	4
243-247	Th2 (erioph) Tb1 Sh1	4/5
247-286	Tb3 (imb) Sh1 Tl+	3/4
286-300	Tb3 Sh1 Tl+ (Th1 erioph localised)	4

LAG3 30 metres from master core on bearing of 090^0
Surface veg: *S. papillosum, S. capillifolium*

Depth (cm)	Troels-Smith	Humification
0-35	Tb4 (pap) Th+	2
35-51	Tb4 (cusp)	1/2
51-58	Tb2 Sh2 Tl+ (pool muds)	4
58-79	Tb3 Sh1	4
79-110	Tb4 (imb)	3
110-120	Tb3 Th1 (erioph)	4
120-137	Tb2 Sh2	4
137-158	Tb4 (imb) Sh+ Th+	4
158-163	Tb3 Th1 (erioph)	4
163-254	Tb4 (imb) Th+ including following variations 213-218 Sh+ H4 218-222 H3 222-230 Sh+ H4 230-240 H3 240-254 H3/4	3/4
254-282	Tb3 Sh1 Th+ (erioph)	4
282-292	Tb4 (imb)	3/4
292-300	Tb2 Th1 (erioph) Sh1 Tl+	4

LAG4 50 metres from master core on bearing 090^0
Surface veg: *S. papillosum*

Depth (cm)	Troels-Smith	Humification
0-6	Tb4 (pap)	2
6-11	Tb2 Sh2 Th+ (rhynch) (pool?)	3
11-19	Sh3 Th1 Tb+ (pool muds)	4
19-28	Tb3 Th1	3/4
28-39	Tb3 Sh1 Th+ Tl+	4
39-70	Tb3 Sh1 (Th erioph 45-50)	3
70-100	Tb3 Th1 Tl+	3/4
100-122	Tb3 Sh1 Tl+ Th+ (erioph)	4
122-134	Tb2 Sh1 Th1 (erioph)	4
134-142	Tb3 Th1 (rhynch) (pool?)	3
142-163	Tb3 Th1 Sh+	4
163-218	Tb4 (imb) Th+	3
218-240	Tb4 Th+ (erioph) Sh+	4
240-250	Tb4	3
250-258	Tb2 Th2 (erioph) Sh+	4
258-300	Tb3 Sh1	5

Temple Hill Moss (TEM)

Sampled on 26/2/99

TEM1 10 metres from master core on bearing of 200^0
Surface veg: *S. magellanicum, E. angustifolium, E. tetralix*

Depth (cm)	Troels-Smith	Humification
0-22	Tb4 Th+ (mag/pap) 10-15cm Th4 (E. ang?)	3
22-100	Th2 Tb2 Sh++ (higher Sh2 Th2 grading to Tb2)	4
100-143	Tb2 Th1 Sh1	5-4
143-150	Tb4 (imb) Tl+ Th+	3
150-159	Th2 Sh1 Tb1 (fine ericaceous rootlets)	5
159-175	Tb4 (imb) Tl+ Th+	3
175-188	Tb3 (imb) Th1 Sh++	4
188-198	Th3 Sh1 (greeny pool rhynch)	4
198-200	Tb4 (imb) Th+	3
200-204	Tb3 Sh1	4
204-220	Tb3 Th1 Sh++ (erioph 213-216cm)	5
220-246	Tb2 Th2 Sh++	4/5
246-252	Th4 (erioph)	4/5
252-267	Th2 Tb1 Sh1 Tl+	5
267-300	Tb3 (imb) Th1	5
300-400	Tb2 Th2 Sh++ (erioph bands 319-321, 335-340, 372-378)	6/7
400-405	Th4 (erioph)	6
405-418	Sh2 Th1 Tb1	7
418-422	Th2 Tb2	6
422-480	Th2 Sh2 Tb+ Tl+ (birch)	7
480-490	Sh3 Tl (birch, dry crumbly black peat)	8
490-500	Tl3 Sh1 Th+ (birch)	8

TEM2 30 metres from master core on bearing of 020^0
Surface veg: *E. tetralix, E. vaginatum, S. magellanicum, S. fuscum*

Depth (cm)	Troels-Smith	Humification
0-30	Tb4 (mag)	1-2
30-48	Tb1 Sh1 Th2	4-3
48-57	Tb3 (imb) Th1	3-4
57-67	Tb2 Sh1 Th1	5
67-100	Tb3 (imb) Th1	4
100-140	Tb3 Th1 (erioph 130-140) Tb (imb) 122-125 H3	4
140-159	Tb4 th++ (locally Tb3 Th1 rhynch and erioph)	3
159-200	Th2 Sh1 Tb1	5/6
200-230	Tb2 Th2 (erioph)	5
230-254	Th3 (erioph) Sh1 Tb++ Tl+	5/6
254-300	Tb3 Th1 (erioph) Tl+	5
300-306	Tb4 (imb)	4
306-338	Tb3 Th1 Tl+	5
338-348	Tb4 (imb) Tl+	3
348-358	Th2 Sh2 Tb+	6
358-400	Th2 (erioph) Sh1 Tb1	6
400-450	Th2 Sh2 Tl+ (erioph roots in places)	7
450-500	Fen peat with birch (468-474 Tl2)	8

TEM3 30 metres from master core on bearing of 200^0
Surface veg: *S. magellanicum/S. papillosum* lawn

Depth (cm)	Troels-Smith	Humification
0-28	Tb4 (mag/pap) Th+	2/3
28-60	Th2 Sh1 Tb1	4/5
60-86	Tb4 Th++ Tl+	4
86-108	Tb2 Sh1 Th1	4/5
108-147	Tb4 (imb) Th+ Sh+	4
147-151	Th4 (cusp – pool) Sh+	3
151-179	Th1 (erioph) Tb2 Sh1	5
179-190	Th3 (rhynch) Sh1 Tb+	5
190-202	Tb2 (cusp) Sh1 Th1	4
202-270	Tb4 Sh+ Th+	5
270-278	Tb4 (imb) Sh+ Tl+	4
278-288	Sh2 Tb1 Th1 Tl+	5/6
288-310	Tb4 (imb) Sh+ Tl+	4/5
310-412	Tb2 Sh1 Th1 (fresher peat in bands 384-388, 392-396)	6/7
412-455	Th3 Sh1 sedge peat (locally Tl+ birch)	8

Shirgarton Moss (SGM)

Sampled on 24/10/98

SGM1 1.5 metres from master core on bearing of 346^0

Depth (cm)	Troels-Smith	Humification
0-6	Tb3 Th1	1-2
6-11	Tb2 Th2 (yellow colour)	2
11-18	Sh2 Th2 Tb+	4
18-27	Tb2 Th2 Sh+	5
37-35	Tb3 Th1	4
35-57	Tb4 Th+	3
57-74	Tb3 (imb) Th1	4
74-120	Tb4 Th+	3
120-129	Tb2 Th1 Tl1	5
129-144	Tb4 Th+	4
144-146	Tb3 Th1	3
146-177	Tb4 Th+	4
177-192	Tb3 Th1 (erioph)	5
192-200	Tb3 Tl1 Th+	5
200-210	Tb3 Th1	5
210-234	Tb3 Tl1	4
234-250	Tb2 Th2	5
250-267	Tb1 Th3 (erioph)	5
267-281	Tb1 Th3 (erioph)	4
281-300	Th4 (erioph) Sh+	6

SGM2 10 metres from master core on bearing of 166^0

Depth (cm)	Troels-Smith	Humification
0-10	Tb4 Th+	2
10-31	Tb2 Th2	5
31-48	Tb3 Th1 (erioph)	4
48-67	Tb4 Th+ (erioph)	4
67-100	Tb4	3
100-123	Tb4 Th+	3
123-136	Tb4 Th+	4
136-138	Tb2 Th2 (erioph) Sh+	6
138-139	Sh3 Th1	6
139-150	Sh2 Th1 Tb1	5
150-238	Tb3 Th1 Tl+	4
238-252	Tb2 Th2 (erioph)	4
252-283	Th3 (erioph) Tb1	4
283-300	Th4 (erioph) Sh+	5

SGM3 30 metres from master core on bearing of 166^0

Depth (cm)	Troels-Smith	Humification
0-5	Tb4 Th+	2
5-13	Sh3 Th1 (erioph)	6
13-27	Sh1 Th3 (erioph) Tb+	5
27-51	Tb2 Th2 (erioph)	4
51-82	Tb3 Th1 (erioph)	4
82-113	Tb4 Th+	3
113-123	Tb4 Th+	4
123-146	Tb3 Th1 (erioph)	3
146-159	Tb3 Th1 (erioph) Sh+	4
159-180	Tb2 Th2 (erioph) Sh+	4
180-184	Th2 Sh2 Tb+	5
184-215	Tb3 Th1 Sh+	4
215-233	Th2 (erioph) Tb2 Sh+	4
233-257	Th2 Tb1 Tl1	6
257-275	Th2 (erioph) Tl1 Sh1 Tb+	5
275-300	Th3 (erioph) Sh1	7

SGM4 15 metres from master core on bearing of 346^0

Depth (cm)	Troels-Smith	Humification
0-10	Tb4 Th+	1
10-19	Tb4 Tl+	2
19-36	Th2 Tb1 Sh1 Tl+	5
36-60	Th2 Tb2 Sh+	4
60-107	Tb3 Th1 (erioph)	3
107-129	Th2 (erioph) Tb2 Sh+ (H grading to 5 lower)	4/5
129-132	Tb2 Th1 (erioph) Sh1	4
132-158	Tb3 Th1 (erioph)	3
158-167	Th3 (erioph) Sh1 Tb+	5
167-185	Th2 (erioph) Tb1 Sh1	5
185-200	Th3 (erioph) Sh1	6
200-215	Tb2 Th1 Sh1	5
215-227	Th2 Tb1 Sh1	6
227-253	Th3 (erioph) Sh1	6
253-300	Tb1 Th2 (erioph) Sh1	7

Mallachie Moss (MAL)

Sampled on 27/2/99

MAL1 5 metres from master core on bearing of 290^0
Surface veg: *S. capillifolium, S. s. Acutifolia* (unknown), *E. tetralix, C. vulgaris*

Depth (cm)	Troels-Smith	Humification
0-25	Tb4 (pap?)	1-2
25-30	Th4 (erioph) Sh+	4
30-100	Tb2 Th2 Sh++ (more Th at top grading to Tb)	4/5
100-112	Tb3 Th1 Sh++	4/5
112-115	Th1 Sh1 Tb2	6
115-125	Tb4 (cusp) Th+	3
125-150	Tb2 Th2 (136-144 Th4 erioph)	6
150-167	Tb4 (cusp) Th+	3-4
167-200	Th2 Tb1 Sh1	6
200-250	Th1 Sh1 Tb2 (219-224 Th3 erioph Sh1)	5
250-255	Tb4 (cusp)	3/4
255-273	Sh2 Tb2 Th+	6
273-278	Sh2 Tb2 Th+	5
278-281	Tb4 (cusp)	3
281-286	Tb4	4/5
286-290	Tb4 Tl+	3/4
290-300	Th1 Sh3	5/6

MAL2 30 metres from master core on bearing of 290^0
Surface veg: *S. rubellum, Cladonia, E. vaginatum, C. vulgaris, Racomitrium, H. cupressiforme*

Depth (cm)	Troels-Smith	Humification
0-20	Tb4	1-2
20-48	Tb2 Th2 (erioph rootlets)	4/5
48-64	Tb2 Th2	5
64-85	Tb3 Th1	3/4
85-100	Tb2 Th2	4
100-140	Tb2 Th2	5
140-155	Tb3 (cusp) Th1 (pool?)	4
155-157	Tb2 Sh2 Th+ (pool muds?)	4
157-167	Tb2 Th2	4
167-175	Tb4 Th+ (cusp pool?)	4
175-200	Tb2 Th2 (erioph)	5
200-220	Sh2 Tb2 Th+	5
220-232	Tb4 (cusp) locally H3	4
232-237	Th2 (erioph) Tb2 (cusp)	4
237-300	Sh1 Tb1 Th2	5/6

MAL3 10 metres from master core on bearing of 110^0
Surface veg: *Hypnum, S. rubellum, E. tetralix, E. vaginatum*

Depth (cm)	Troels-Smith	Humification
0-35	Tb4	1-2
35-46	Th3 (erioph) Sh1	4
46-55	Th2 Tb2 Sh+	5
55-74	Tb1 Th3 (erioph) Sh+	4
74-126	Tb2 Th1 Sh1 (locally Th2 erioph)	5
126-132	Tb4 (cusp)	3/4
132-150	Tb3 Sh1	5
150-155	Tb4 (cusp?) Tl+	3/4
155-162	Tb2 Sh1 Th1	5
162-188	Tb4 (cusp?) Sh+ Th+ (some rhynch)	4/5 (grading)
188-200	Tb2 Th1 Sh1	5
200-230	Sh3 Tb1 (locally Th1 erioph)	6/7
230-252	Th2 (erioph) Tb2 Sh+	5
252-274	Tb3 Th1	4
274-288	Tb4 (cusp)	3
288-300	Tb3 Sh1 Th+ (some cusp)	4

MAL4 30 metres from master core on bearing of 110^0
Surface veg: *S. cuspidatum, E. vaginatum, E. tetralix*

Depth (cm)	Troels-Smith	Humification
0-27	Tb4	1-2
27-80	Th2 (erioph) Sh1 Tb1 (erioph more abundant at top)	5
80-90	Tb2 Th2	4/5
90-100	Tb2 Th1 Sh1	4/5
100-128	Tb3 Sh1 Th+	4
128-145	Tb3 Sh1 Th+	4/5
145-157	Tb4 Sh+ Th+	4
157-181	Tb4 (cusp)	3/4
181-200	Sh2 Th1 (erioph) Tb1	5
200-224	Tb3 Th1 Sh+	4/5
224-250	Tb4 (cusp) locally Sh1	3
250-270	Tb2 Sh2 (locally Th1 erioph)	4
270-300	Tb4 (cusp pool 288-290cm H3)	3/4

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