Holocene climate change in Glen Affric, Northern Scotland: a multi-proxy approach

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ABSTRACT

A multi-proxy approach was used to generate a continuous, sensitive Holocene palaeoclimatic record for Glen Affric, north west Scotland. Fluctuations in lake-level were used as a proxy for shifts in precipitation. Rigorous site selection criteria and a new methodology were developed to interpret the lacustrine sediment record in terms of climatically driven changes in water depth by separating (a) allogenic from autogenic sediment inputs and (b) explicitly linking the marginal fen peat system responded to changes in lake-level. The sedimentary record from the lake site, Loch Coulavie, suggests that lake-level has fluctuated repeatedly throughout the Holocene. The comparative magnitude of changes in lake-level defined the relative intensity of shifts in precipitation.

Variations in mire surface wetness, as determined through humification analysis, from a series of four hydrologically isolated ombrotrophic blanket mire sites through the east-west trending glen, were used to generate a record of changes in effective precipitation. A reliable radiocarbon chronology obtained from both proxy records allowed the synthesis of these data sets and the definition of Holocene climate change in terms of relative shifts in temperature and precipitation. The data suggests that the early Holocene was more stable in terms of both temperature and precipitation, but that after *c.* 6200 BP (7200 cal. BP) both temperature and precipitation became highly variable. Several short-lived, abrupt high intensity shifts to increased precipitation occur at *c.* 6200 BP (7200 cal. BP), *c.* 5000 BP (5700 cal. BP), *c.* 3000 BP (3200 cal. BP) and *c.* 2400 BP (2350 cal. BP).

Holocene climatic variability within Glen Affric corresponds to records of changes in North Atlantic oceanic circulation patterns. The predominance of atmospheric systems, such as Atlantic westerlies, may also have controlled spatial climatic variability within the glen, with the periodic establishment of very steep west-east climatic gradients, steeper than at the present day.

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CHAPTER 1
HOLOCENE CLIMATIC CHANGE IN THE NORTH ATLANTIC REGION: A CRITICAL REVIEW

1.1 INTRODUCTION

Continuous high-resolution records from oceanic sediments have revealed a detailed depiction of climate change over the last 3.5 Ma, with the Earth's climate undergoing cyclical periods of warm (interglacial) and cold (glacial) conditions (Ruddiman *et al.* 1989). These cyclical periods of climate change are shown to correspond with variations in the Earth's orbital geometry around the sun, Milankovitch cycles. Section 1.2 reviews how high-resolution palaeoclimatic data have revealed climatic changes occurring throughout the last glacial with climatic shifts operating on sub-Milankovitch scales. This research has highlighted the connection between oceans, atmosphere and ice sheets, in particular within the North Atlantic Ocean, and indicates possible forcing mechanisms (Section 1.2).

The recent application of such high-resolution studies to North Atlantic, Holocene marine and ice core records (Section 1.3) has suggested that similar but more subdued ocean-atmosphere-ice feedback systems may have operated during the present interglacial. This would suggest that terrestrial sites along the Atlantic Ocean fringes will be sensitive to any changes within these coupled systems (Rind and Overpeck 1993). The impetus now is to look for highly resolved palaeoclimatic data within these climatically sensitive regions, such as north west Scotland. Section 1.4 critically reviews palaeoclimatic data from this region, with current weaknesses in these data sets such as insensitivity and ambiguity of climatic proxies driving the aims and methodological approach of this research (Sections 1.5, 1.6 and 1.7).

1.2 CLIMATE CHANGE DURING THE LAST GLACIAL: A PRECURSOR TO THE HOLOCENE

High-resolution ocean sediment and ice core records suggest a highly variable climate during the last glacial period (*c.* 110-11 ka BP (Lowe and Walker 1997)). Atlantic Ocean marine sediment records revealed evidence for short-lived (around 2000 years) intervals of warmer climate (interstades), with changes in the ocean temperature a response to changes in the position of the oceanic North Atlantic Polar Front (the boundary between warm, high salinity water flowing north, and cold low salinity water flowing south from the Arctic) (Ruddiman *et al.* 1977; Ruddiman and McIntyre 1981). Atlantic marine sediments also record a series of ice-rafted debris sediments (Heinrich events) which had occurred at 7-10ka intervals during the last glacial (Heinrich 1988). These sediments were deposited during large-scale ice-rafting events with ice streams originating from discharges from the Laurentide, Fennoscandinan and British ice sheets (Alley and MacAyeal 1994; Baumann *et al.* 1995; Fronval *et al.* 1995). Ice core records from Greenland generated further evidence of a highly variable climate during the last glacial with several frequent, substantial fluctuations in atmospheric temperature. These irregular well defined events (Dansgaard-Oeschger cycles), lasting between 1000–3000 years, consisted of an abrupt transition over decades to a relatively mild climate terminated more gradually in step-wise fashion (Dansgaard *et al.* 1993; Johnsen *et al.* 1992). Bond *et al.* (1993) and Bond and Lotti (1995) combined the terrestrial record with the marine record and proposed that a series of Dansgaard-Oeschger cycles formed a gradually declining temperature record over 10-15 000 years which culminated in a Heinrich event, followed by a dramatic, rapid (over a few decades) warming (Bond cycles).

Bond cycles and further data from the ice core record (Taylor *et al.* 1993) indicate that within the North Atlantic region over the last 20-80 000 years climatic change was being forced by a mechanism other than orbital forcing (Milankovitch cycles) with the atmosphere-ocean and ice sheet systems repeatedly undergoing rapid, massive reorganisations. What is not clear at present is how much these systems were responding to internal forcing mechanism such as ice-sheet dynamics (binge-purge cycles; McAyeal 1993) or to external forcing such as solar variability (Stuiver *et al.* 1997). There is increasing evidence of a global signal for Dansgaard-Oeschger cycles, Heinrich events and Bond cycles suggesting strong northern and southern hemisphere links within the ocean-atmospheric system or the role of external forcing mechanisms (Behl and Kennett 1996; Clapperton 1997; Hicock *et al.* 1999).

One such mechanism thought to play an important role in climatic change within the northern hemisphere during the last glacial is North Atlantic thermohaline circulation (Broecker and Denton 1990; Broecker 1994, 1997). Present-day climate within North West Europe, in particular regions along the fringes of the Atlantic ocean, such as Scotland (Section 2.6), are to a great extent controlled by the Atlantic Ocean thermohaline circulation. Any disturbance to this system, in particular through interruption of the formation of the southerly return current, the North Atlantic Deep Water Current, would prevent the flow northwards of warmer North Atlantic currents. Clearly there exist strong feedback mechanisms and thresholds within atmospheric and oceanic circulation, in particular within the North Atlantic system, which must be present in order ultimately to cause climate change (Broecker *et al.* 1992; Lehman and Keigwin 1992; Paillard and Labeyrie 1994; Cortijo *et al.* 1997;Vidal *et al.* 1997). This is demonstrated by the Younger Dryas event (Isarin *et al.* 1998) which marked a brief return (<1000 years) to very cold climatic conditions in particular within the northern hemisphere at around 11 000 BP (13 000 cal. BP). This period saw the return of glaciers and ice sheets to areas such as Scotland (Loch Lomond Readvance; Section 2.3) which had previously begun to retreat in response to warmer interglacial conditions as forced by Milankovtich cycles. The Younger Dryas event is thought to have been in response to a disturbance of the North Atlantic thermohaline circulation (Broecker *et al.* 1989). It has been suggested that the intensity of this event is unique, a response to the catastrophic release of melt water from huge glacial lakes (Broecker *et al.* 1989; Bjorck *et al.* 1996). However, it is also argued (see below) that although the intensity may be in response to a combination of unique circumstances this type of event may have occurred throughout the last glacial period and the Holocene.

1.3 CLIMATE CHANGE DURING THE HOLOCENE

For much of the Holocene (present interglacial) (or after around 10 000 BP (11 600 cal. BP)) the large ice masses of the last glacial are absent and thus any potential ocean-atmosphere-ice climatic system will operate differently. This section reviews the evidence for climatic change in the North Atlantic region during the Holocene.

Terrestrial palynological and peat stratigraphic data collected throughout the 19th and 20th centuries were used to produce climate change models such as proposed by Blytt and Sernander (Table 1.1). Lamb (1977) reviewed available palaeoclimatic data and cited evidence for climatic change during the Holocene within the zones of Blytt and Sernander (Table 1.1). This model of climatic change is limited by interpretations of ambiguous climate proxies with poor definition of both magnitude and frequency of climatic shifts. The data set is also restricted both temporally and spatially for much of the Holocene. Climatic changes and temporal variability proposed by Lamb become more detailed from around AD 900 when documentary evidence can be interpreted. Any of the climatic interpretations predate modern quantitative palaeoclimatology and are based on the author's subjective estimates and personal perspective (Bradley 2000). This model of Holocene climate change has remained within the literature (*cf.* Houghton *et al.* 1992; Matthews *et al.* 1997), in particular the Medieval Warm Period and the Little Ice Age upon which much palaeoclimatic and palaeoenvironmental interpretations have become fixed, in so-called 'suck-in and smear' (Baillie 1991; Hughes and Diaz 1996). Temperature records generated from Greenland ice-cores had suggested that climate change during the Holocene had been much more subdued (Lehman and Keigwin 1992), supporting the above model of limited climatic change.

However, recent Holocene palaeoclimatic research has started to question many of the above assumptions. As early as 1973 Denton and Karlen proposed that regional northern hemisphere glacier limit fluctuations occurred cyclically throughout the Holocene and the last glacial. This research suggested a highly variable climate and that controls operating during the glacial period also operated during the Holocene. The significance of this research was underestimated until high-resolution ocean-sediment and ice-core data began to suggest similar fluctuations.

Years (C^{14})	Climatic Period	Climatic Conditions
AD 1550-1850	Little Ice Age (LIA)	Severe climatic deterioration.
AD 800-1300	Medieval Warm Period (MWP)	Sharply renewed warming from about AD 800 led to an important warm epoch which culminates around AD 1100-1300 in Europe. Climates nearly as warm as in the postglacial warmest times.
AD 400-800		Reversion to colder and wetter climates
500 BC- AD 400		Gradual, fluctuating recovery of warmth and a tendency towards drier climate in Europe over the 1000 years after 60 BC particularly after 100 BC, leading to a period of warmth around 400 AD.
1 000-500 BC	Sub-Atlantic	Cooling of world climates, in Europe the most marked change is from 1200-700 BC and by 700-500 BC a prevailing fall in temperature of 2°C. An increasingly wet period with milder increasingly windy winters but cooler summers.
3 000-1 000-500 BC	Sub-Boreal	Climate regained former warmth and may have been warmer than previous period. Recurrent climate fluctuations in particular in rainfall with a 200 year periodicity underlying the more important fluctuations. Perhaps as a response to more variable circulations within the Atlantic.
6 000-3 000 BC	Atlantic	Warmest since postglacial period, within Europe summer temperatures reached a maximum by 6000 BC. Mildest winter since the post glacial with increased moisture. During the last 500 years an oscillation to cooler climates.
7 000- 6 000 BC	Boreal	Temperatures continued to rise. Colder seasons gradually became milder, though possibly with some dry and frosty winters and the summers became generally warmer than today.
8 300- 7 000 BC	Subarctic (Preboreal)	Climatic conditions undetermined but a renewed rise in temperature from around 8 300 BC.

Table 1.1 The Blytt-Sernander nomenclature for the sequence of European Holocene climatic episodes with the climatic interpretation based on Lamb (1977).

Further examination of differing climate proxies within the Greenland ice cores began to suggest that during the Holocene there were several abrupt transitions to cooler and warmer periods as recorded by changes in amounts of snow accumulation (Meese *et al.* 1994; Mayweski *et al.* 1996) and geochemical changes (O'Brien *et al.* 1995). These proxies reflect changes in the atmospheric circulation above Greenland, with rapid climate change events reflecting substantial changes within the atmospheric circulation. Although more subdued than glacial climate change, this climatic variability during the Holocene is thought to represent a more complex record of changes in source and strength of atmospheric circulation (Mayweski *et al.* 1997). Recent research from

North Atlantic marine sediments also suggest a highly variable Holocene climate (Bond *et al.* 1997; Bianchi and McCave 1999; Chapman and Shackleton 2000). These ocean records suggest a periodicity of climatic shifts of around 1500 years, also recorded within the ice core records (O'Brien *et al.* 1995), indicating a strong ocean-atmosphere link throughout the Holocene perhaps associated with the North Atlantic thermohaline circulation (Bond *et al.* 1997; Bianchi and McCave 1999). In particular a notable cooling event, with half the intensity of the Younger Dryas stade, is recorded in both oceanic and ice core records at around 7500 BP (8200 cal. BP) (Alley *et al.* 1997; Stager and Mayweski 1997; Klitgaard-Kristensen *et al.* 1998; Barber *et al.* 1999). Bond *et al.* (1997) propose that events such as the Younger Dryas and the event at 7500 BP (8200 cal. BP) are in response to cyclical changes in the oceanic-atmosphere circulation which also occurred throughout the last glacial indicating a continuation of climate forcing mechanisms. These records of climate change during the Holocene and potential climate forcing mechanisms are discussed further in Section 9.5.

Previous terrestrial Holocene palaeoclimatic records have been limited by (a) the sensitivity of the climate proxy (see further discussion below): climate proxies are sensitive to a range of control factors, not only climatic; (b) a restricted number of continuous high-resolution records similar to ice core and marine sequences and (c) strong regionally specific climate signals (see Chapter 9). Recent palaeoclimatic research using potentially climatically sensitive high-resolution terrestrial records such as tree-ring records (Briffa 1994; 1999), glacier limits (Nesje and Johannessen 1992), tree-line fluctuations and lake-levels (Haas *et al.* 1998), lake sedimentation variability (Campbell *et al.* 1998) and ombrotrophic mire stratigraphy (Barber *et al.* 1994a, b) indicate that Holocene climate change has been variable, with variability operating over decadal and millennial scales, and that this variability is also highly regional. As outlined above, marine and ice core records both stress the links between ocean and atmospheric circulation and within the North Atlantic the thermohaline circulation is seen by some as a key to this relationship. There has been some attempt to match terrestrial with oceanic and ice core records (Campbell *et al.* 1998; Klitgaard-Kristensen *et al.* 1998) with changes in these terrestrial records synchronous with marine cyclicity and climatic events such as the Little Ice Age and the event at 7500 BP (8200 cal. BP).

Climate variability on a decadal (12-14 years) scale recorded within the Greenland ice cores is thought to be a response to the North Atlantic Oscillation (NAO) (Hurrell 1995; Sutton and Allen 1997). The NAO is the large-scale alteration of atmospheric mass between the Icelandic low and Azores high. NAO extremes are thought to modulate and interact with atmosphere-ocean dynamics within the North Atlantic with anomalies in sea surface temperatures, Gulf Stream strength, zonality and strength of westerly flow atmospheric wave structure and frequency of the occurrence of sea ice and ice bergs (Cook *et al.* 1998). The connection between oceanic and atmospheric systems, in particular the interaction with changes in sea surface temperatures, indicate that the NAO is an important control on the intensity of Atlantic storm tracks within the North Atlantic region and into Europe (Sutton and Allen 1997; Rodwell *et al.* 1999; Ulbrich and Christoph 1999). The oscillatory nature of the NAO is considered a long-term feature of the North Atlantic climate system. Records of decadal variability recorded within a 300-year tree-ring record from North America and Europe (Cook *et al.* 1998), within Greenland ice core data up to 350 years (Appenzeller *et al.* 1998) and as a record of winter ice severity in the Western Baltic up to 500 years (Koslowski and Glaser 1999) are attributed to changes in the NAO. However, there is evidence to suggest that the NAO is an intermittent climate oscillation with active and passive phases (Appenzeller *et al.* 1998) and that the decadal variability of climatic change has become more pronounced since the 1950s, perhaps as a response to greenhouse gas forcing (Hurrell 1995; Ulbrich and Christoph 1999). The above discussion indicates that only highly resolved Holocene palaeoclimatic records (up to 500 years in duration) identify climatic forcing as a response to NAO extremes and, as suggested above, there is now a question about the strength of this oscillation and that more pronounced variability seen at the present day may be as a result of increased greenhouse gases (Ulbrich and Christoph 1999). However, what the NAO and associated climatic response to the extremes does highlight is the interaction and coupling between the ocean and atmosphere within the North Atlantic region and subsequent controls on the climate of the region.

Climate variability during the last glacial is thought have involved changes in atmospheric and oceanic circulation patterns, in particular within the North Atlantic, with increasing evidence to suggest that Holocene climate change may have had similar controls but operating at a different magnitude. Conditions within the North Atlantic including the presence of sea ice (Isarin *et al.* 1998) and an oscillating North Atlantic Arctic Front (cooler sea surface temperatures) (Austin and Kroon 1996) are cited as the primary controls on climatic conditions (temperature and position of westerly storm tracks) along the north western seaboard during previous climatic events such as the Younger Dryas. Changes in atmospheric circulation during the Holocene, in particular the location and strength of storm tracks (possibly as similar but perhaps more prolonged circulation patterns to the NAO extremes discussed above), have been proposed as the most likely control of centennial-millennial changes in precipitation as recorded by fluctuations ice accumulation over Greenland (Kapsner *et al.* 1995). Terrestrial sites along the North Atlantic fringes are known to have been sensitive to such ocean-atmospheric changes during the last glacial and so should also have been during the Holocene.

1.4 HOLOCENE CLIMATE CHANGE IN NORTH WEST SCOTLAND

The section critically reviews palaeoclimatic research within north west Scotland (Figure 1.1). Initial palaeoclimatic inferences in north west Scotland were generated from the palynological record with apparent synchroneity of vegetation dynamics considered indicative of climatic changes (*cf.* Godwin 1975). Species such as *Alnus*, were associated with specific environments, with the *Alnus* rise interpreted as a shift to wetter climatic conditions. However, further palynological and ecological investigations with the increased application of radiocarbon dating began to question the synchroneity of such events and the climatic sensitivity of vegetation dynamics (Bennett and Birks 1990; Chambers and Elliott 1989). Lags between climatic shifts determined through other proxies and vegetation change indicated that the response of vegetation was complicated by the interaction of other factors such as edaphic development, migration rates, plant maturation rates and inter-species competition (Huntley and Birks 1983; Pennington 1986; Birks 1989; Lowe 1993). Vegetation dynamics as a climate proxy are further limited by increased anthropogenic impact from around 5000 BP (5800 cal. BP), with burning, clearance and grazing by domestic animals all causing vegetation change (Lowe 1993).

More sophisticated approaches with the use of transfer functions (Guiot 1987, Guiot *et al.* 1993; Huntley 1992, 1993) and well defined indicator species (Zagwijn 1994) have tried to refine the palynological palaeoclimatic record further including generating quantifiable climatic values. However, the above limitations still apply and there is concern that any palaeoclimatic interpretation based solely on vegetation change generates a flawed and biased record (Tipping 1996; Edwards and Whittington 1997).

Within north west Scotland, Holocene palynological records indicate an initial climatic amelioration at the opening of the Holocene with local and regional variations reflecting the factors outlined above. Early Holocene fluctuations in woodland cover where such vegetation types were considered ecologically marginal are thought to be in response to climatic change (Fossitt 1990; Birks 1996). Such palaeoclimatic interpretations, although perhaps with these spatially limited vegetation types considered climatically sensitive, are still based on palynological events which overall are seen as complex responses to factors other than climate change (Lowe 1993). Such palaeoclimatic records generated from palynological data for north west Scotland as reviewed by Birks (1996) are very poorly defined particularly in terms of nature, timing and magnitude of climate shifts.

Considerable palaeoclimatic research has focused on the palynological event, the Mid Holocene Pine Decline (MHPD); this abrupt vegetation change consisting of a reduction in *Pinus sylvestris* (Scots Pine) between *c.* 4000-4500 BP (4500-5000 cal. BP) is recorded at several sites within the region (Pennington *et al.* 1972; Birks 1975; Bennett 1984). Such synchroneity across the region was thought to be in response to climate change, with increased wetness causing waterlogging, blanket peat expansion and pine reduction (Bennett 1984; Pennington *et al.* 1972).

Numerous and spatially extensive pine macrofossils with ages clustered around 4000-4500 BP (4500-5000 cal. BP) (Birks 1975) indicated the presence of a once extensive pine woodland growing on drier peat surfaces, suggesting a period of drier climate prior to a shift to wetter climate. Fine temporal resolution palynological studies revealed that in regions (northern Scotland, Figure 1.1) with no previous record of such an event, pine underwent an expansion around 4400 BP (4900 cal. BP) under proposed drier and

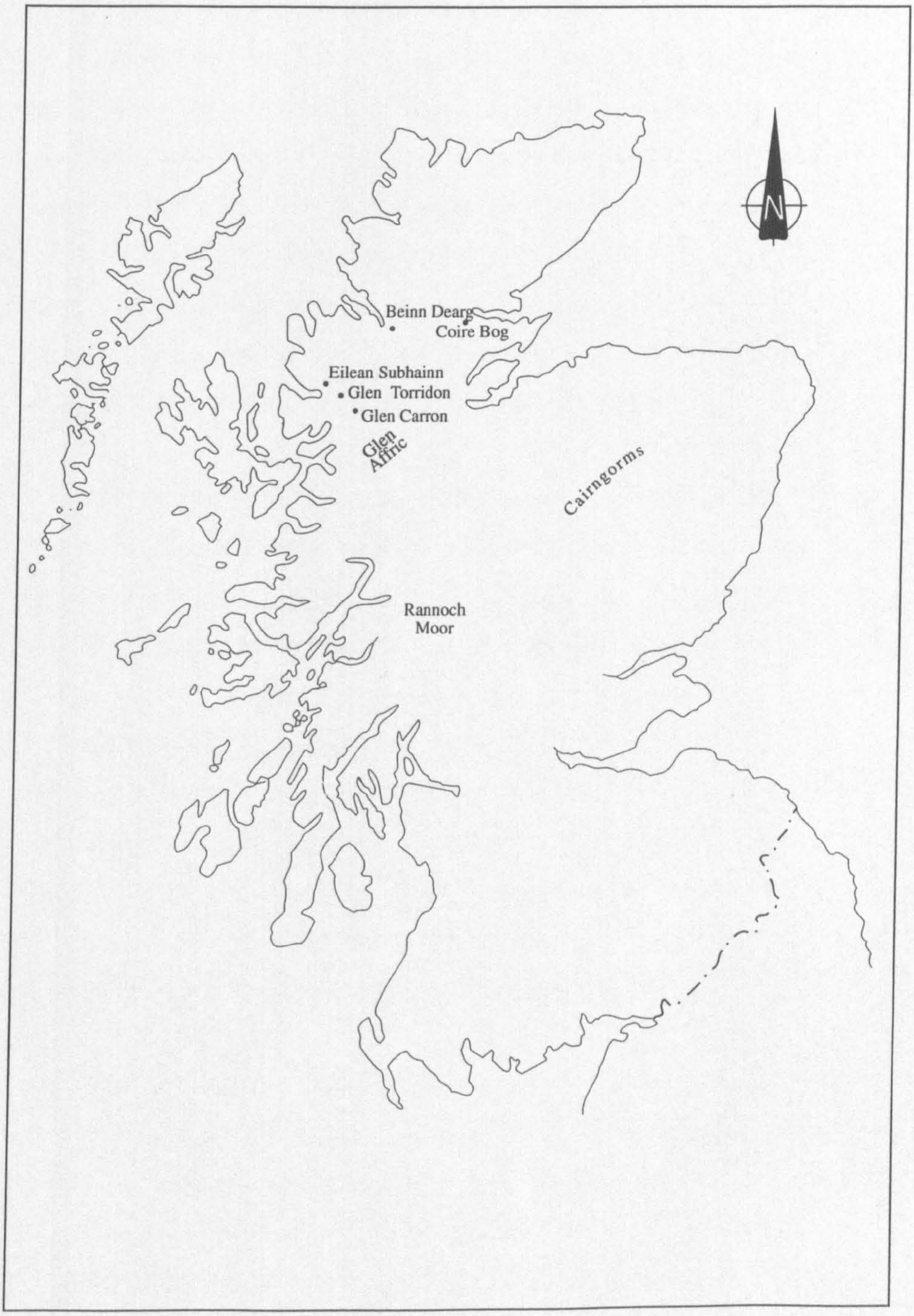


Figure 1.1 Location of previous research sites discussed in Section 1.4: Beinn Dearg (Binney 1997), Eilean Subhainn, Glen Torridon, Glen Carron (Anderson 1996, 1998), Coire Bog (Gear and Huntley 1991), Rannoch Moor (Bridge et al. 1990).

warmer climatic conditions with an abrupt reduction at around 4000 BP (4500 cal. BP) in response to proposed wetter climatic conditions (Gear and Huntley 1991). Further palynological research suggests that the MHPD in north west Scotland is no longer considered a synchronous event (Bridge *et al.* 1990; Bennett 1995; Anderson 1996; Anderson *et al.* 1998; Daniell 1998) with several regional and local temporal variations. Bennett (1995) reviewed the evidence for the MHPD event and highlighted difficulties in identifying a single cause for such an event, with factors such as pine macrofossil preservation (see below), human activity and the ecological response of pine to climatic change complicating any interpretation of the palynological record.

The Holocene 'precipitation' record generated by Dubois and Ferguson (1985) using variations in the deuterium:hydrogen isotopic ratio within fossilised pine wood represents a potentially unambiguous palaeoclimatic record for upland Scotland. Periods of increased precipitation, 'pluvials', were proposed, occurring at *c.* 3300 BP (3600 cal. BP), *c.* 4200-3940 BP (4800-4300 cal. BP), *c.* 6200-5800 BP (7050-6500 cal. BP) and *c.* 7500 BP (8200 cal. BP). However, (a) this record of periods of increased 'precipitation' was generated using pine wood from the Cairngorms (Figure 1.1) and with climate change potentially regional in extent, may not be applicable to the north west of Scotland; and (b) the ratio of deuterium:hydrogen within the pine wood reflects relative humidity (Dubois 1984), which has further controls such as wind speed and in particular temperature, and not simply precipitation (Tipping 1994).

In an attempt to define Holocene climatic change further, within northern upland Scotland, Bridge *et al.* (1990) (Figure 1.1) adopted a multi-proxy approach. However, proxies used were still heavily reliant on vegetation dynamics and therefore potentially unambiguous climate proxies (see above). Proxies such as Scots pine tree line variations, macrofossil preservation, pine expansions and dendrochronology indicated that density and distribution of pine had varied significantly during the Holocene. This record was compared to the record of precipitation changes generated by Dubois and Ferguson (1985); pine vegetation dynamics were considered a response to these variations in precipitation levels, but that there was a significant lag between a shift in climate and woodland change (Tables 1.2, 1.3). The authors also concluded that pine macrofossils were largely an artefact of preservation with preservation a result of increased mire wetness.

Shifts to Drier and/or Warmer Climate.

Anderson (1996, 1998)	**Anderson *et al.* (1998)**	**Binney (1997)**	**Bridge *et al.* (1990)**	**Gear and Huntley (1991)**
c.1600 BP (1400 cal. BP)	c. 1700 BP (1500 cal. BP)			
c. 3800-4100 BP (4150-4400 cal. BP)	c. 3800-4400 BP (5000-4150 cal. BP)	c. 3800 BP (4150 cal. BP)		
		c. 4250 BP (4800 cal. BP)		c. 4000-4400 BP (4500-5000 cal. BP)
		c. 4800 BP (5400 cal. BP)	c.4660-4970 BP (5350-5700 cal. BP)	
c. 5100-5350 BP (5800-5950 cal. BP)				
		c. 6400 BP (7300 cal. BP)		
c. 6800 BP (7600 cal. BP)			c. 6600-6800 BP (7400-7600 cal. BP)	
c. 8100 BP (9100 cal. BP)				

Table 1.2 Summary of proposed shifts to drier and/or warmer climate in north west Scotland (see Figure 1.1)

Shifts to Wetter and/or Cooler Climate

Anderson (1996, 1998)	**Anderson *et al.* (1998)**	**Binney (1997)**	**Bridge *et al.* (1990)**	**Gear and Huntley (1991)**
c. 950 BP (900 cal. BP)	c. 1000 BP (1000 cal. BP)			
c. 2900 BP (3200 cal. BP)	c. 2700 BP (3000 cal. BP)	c. 2400 BP (2350 cal. BP)	After c. 3250 BP (3500 cal. BP)	
c. 3600 BP (3900 cal. BP)	c. 3250-3650 BP (3500-3900 cal. BP)	c. 3500 BP (3750 cal. BP)	c. 3500-3800 BP (3750-4150 cal. BP)	
c. 4500 BP (5200 cal. BP)		c. 4100 BP (4550 cal. BP)		c. 4000 BP (4500 cal. BP)
c. 5700 BP (6250 cal. BP)	c. 5400-5800 BP (6000-6500 cal. BP)		c.5250-5700 BP (5900-6250 cal. BP)	
c. 7400 BP (8100 cal. BP)				

Table 1.3 Summary of proposed shifts to wetter and/or cooler climate in north west Scotland (see Figure 1.1).

However, temporal and spatial variations in pine macrofossil distributions were thought to reflect local influences on mire hydrology and preservation. The research is important in emphasising the need for a methodological approach involving multi-proxies but perhaps cannot be seen as an unambiguous palaeoclimatic record owing to problems with the climatic sensitivity of proxies used.

Within the upland regions of north west Scotland there is evidence for extensive Holocene geomorphic activity (Ballantyne and Dawson 1997). Climatic change, in particular increased precipitation, is considered to be an important forcing mechanism for a range of geomorphic processes. However, climate may be only one of several factors acting through a range of feedbacks and thresholds within the geomorphic system and as such, geomorphic events are considered an ambiguous climate proxy.

Recent palaeoclimatic research within north west Scotland has focused on obtaining a palaeoclimatic record from changes in ombrotrophic blanket mire stratigraphy. Proxies, including a record of mire surface wetness, are used to reconstruct past changes in ombrotrophic mire hydrology, subsequently interpreted as shifts in effective precipitation. Effective precipitation is the amount of precipitation minus evaporation with effective precipitation controlled by both by the amount of rainfall but also temperature. Therefore climate shifts are described as wetter and/or cooler, and drier and/or warmer. Anderson (1996, 1998) uses a multi-proxy approach with continuous palaeoenvironmental records obtained from three peat bogs located in the Wester Ross region (Figure 1.1). Binney (1997) uses similar proxies to obtain a record of changes in blanket mire hydrology recorded from several sites, amongst others, on Beinn Dearg (Figure 1.1). A regional synthesis of palaeoclimatic data generated from these proxies suggests that Holocene climate change within the north west region of Scotland was highly variable with a strong regional control (Anderson *et al.* 1998, Tables 1.2, 1.3). Anderson *et al.* (1998) propose that in particular there is evidence for a regional shift to wetter climatic conditions at *c.* 3650-3250 BP (3900-3500 cal. BP). This shift occurred abruptly over a decadal to century timescale and synchroneity with similar shifts in other regions suggests a continental or possibly even global scale change in climate.

The above discussion of evidence for Holocene climate change in north west Scotland reveals several shortcomings within this palaeoclimatic record:

- considering the climatic sensitivity of the region (Section 1.3) the palaeoclimatic data base is comparatively limited.
- those palaeoclimatic records generated concentrate on highly ambiguous data such as vegetation change, or climatically non-specific data such as effective precipitation.
- much research has concentrated on the explanation of specific events such as the MHPD rather than a continuous record. A continuous record would determine whether this event was, for instance, typical or exceptional.
- there is a concern that palaeoclimatic data have been generated from complacent sites and that only extreme events have been recorded with particular reference to the more recent palaeoclimatic research. Anderson (1996) and Binney (1997) record surface wetness variations from peat profiles interpreted as shifts in effective precipitation; however, the climatic sensitivity of this proxy is only correct if the mire is fully ombrotrophic, that is rain-fed (see Section 1.6.2). Although Anderson (1996) considers his sites to be hydrologically isolated they are effectively basins which collect water and therefore could lead to complacency within the mire surface record. Similarly Binney generated surface mire wetness data but from sites such as valley mires which will have a catchment area and therefore may also be complacent. Site selection and sensitivity of the existing proxy record are further discussed in Section 9.4. Therefore although regionally there appear to be synchronous climatic events (Tables 1.2, 1.3) there is a concern that these events are extreme and a full and sensitive record of all climatic shifts needs to be generated to provide an appropriate context.
- current data are also poorly defined in terms of frequency of events, through a lack of chronological controls.
- there is still very little understanding of magnitude and intensity of climatic events.

1.5 RESEARCH AIMS

The discussion above has criticised the sensitivity of currently available climatic proxies and aspects of the methodological approach to previous research into Holocene palaeoclimate within north west Scotland. These criticisms have driven the aims of this research. The principal aim of this research is therefore to generate an unambiguous continuous Holocene palaeoclimatic record. Unambiguous records are here defined as the identification of proxies that (a) are clearly associated with only one of the major climatic variables, temperature or precipitation, and (b) the cause of change within such proxies are demonstrably climatic with no link to non-climatic controls such as pedological or anthropogenic causes. The continuity of the record must come from defining climate proxies that involve the long-term and continuous accumulation of sediment. The fulfilment of this aim with the definition of Holocene climate shifts generates several objectives as follows:

(1) to define the role of climate variables, such as temperature or precipitation, in the palaeoclimatic record; the ambiguity that exists in current single proxy studies suggested that a multi-proxy record needed to be generated.
(2) to define the intensity of climatic shifts.
(3) to define the duration and regional synchroneity of climate events. The generation and comparison of multi-proxy data necessitates the development of independent, robust and internally coherent chronologies. Such chronologies would enable the definition of climatic shifts in terms of duration and, when compared to the regional record, determine regionally synchronous shifts. Further spatial refinement of Holocene climate shifts, for example the role of west-east climatic gradients, would be through the generation of sensitive, chronologically well defined climatic data from a series of sites within a region, comparable in all aspects other than climatic gradients.

1.6 METHODOLOGIES

The aim and objectives outlined above determined the selection of a study area and methodologies. West Glen Affric is in the mountainous Scottish highlands, everywhere above 250 m OD, which increases the responsiveness of the system to change (Section 1.3 and Chapter 2), and which reduces the likelihood of anthropogenic-induced change. An associated study (Davies 1999) also allowed the determination of the timing and extent of anthropogenic impacts and this is referred to in subsequent chapters. In addition, work on blanket peat formation presented in this thesis (Chapter 8) allowed the major pedological change in the region to be controlled for in reconstructions.

To generate an unambiguous climatic record, sensitive climate proxies with a continuous record within the study area were selected (Sections 1.6.1, 1.6.2). Careful site selection maximises the climatic control of the measured response (data set) within the proxy record (Sections 1.6.1, 1.6.2). A multi-proxy approach is used and climatic data are generated from proxies with different climatic sensitivities. Through a combination of these data sets, comparing one record with another, the climatic record is tested, defining the nature, intensity and magnitude of climatic shifts (Section 1.7). Critical to the definition of climatic events is the construction of an accurate chronological framework, through a radiocarbon chronology- the approach outlined in Section 5.6.

1.6.1 LAKE-LEVEL FLUCTUATIONS

Lakes fluctuate in size and depth in response to changes in catchment hydrology. Changes in catchment hydrology can be driven by climatic controls such as precipitation and evaporation and non-climatic controls such as vegetation change and soil structure. Therefore it is essential when investigating fluctuations in lake-level as a climate proxy that the lake site is sensitive to climatic controls on the catchment hydrology: Chapter 3 critically reviews site selection criteria.

Harrison *et al.* (1993, 1996) proposed that changes in lake-level at such climatically sensitive sites cannot be in response to changing radiation budgets, seasonal temperature regimes or cloudiness but that variation in the amount of precipitation is the most important control on lake-level. A hydrological model developed by Vassiljev *et al.* (1998) using a record of quantified changes in water depth agreed with the above in proposing that radiation and temperature variability in isolation are not enough to force changes in lake-level. However, when these variables were included within the hydrological model only a relatively minor change in precipitation was required to generate the observed fluctuations in lake-level. This hydrological model is site specific but highlights the sensitivity of lacustrine systems to relatively minor changes in the amounts of precipitation.

Lake-level fluctuations as a climate proxy in northern temperate regions have been widely investigated (Harrison *et al.* 1996), in particular in southern Sweden (Digerfeldt 1986, 1988, 1997) and the French Jura mountains (Magny 1992, 1993). These records suggest a highly variable record of Holocene climate change with strong regional shifts. The sensitivity of the record of fluctuating lake-level to precipitation has been further exploited by pollen analysts in an attempt to try and refine further the palynological record in terms of shifts in temperature and precipitation (Guiot *et al.* 1993).

Within Great Britain and, in particular within Scotland, lake-level fluctuations are an under researched climate proxy. This is highlighted by the record of Scottish lake-level data compiled by Harrison *et al.* (1996) and Yu and Harrison (1995) in the European Lake Status Data Base (ELSDB) (Table 1.4). More recent investigations generating lake-level data within Scotland (Milburn 1996; Smith 1996) are also shown in Table 1.4. Although lacustrine sedimentation should provide a continuous record, this existing record of lake-level fluctuations is limited, with only one or two changes in lake-level having been recorded. This is primarily due to many of these sites investigated principally as palynological sites. A hiatus within the pollen record was thus explained in terms of variable sedimentation rates associated with a change in lake-level (Boyd and Dickson 1987; O'Sullivan 1975a,b, 1976; Stewart *et al.* 1984). Palynological data are normally obtained from deeper water lacustrine sediment cores, which are not considered sensitive to all changes in lake-level (Section 3.3; Digerfeldt 1986). Where multiple cores including marginal cores are investigated (Edwards and Whittington 1993), several fluctuations in lake level can be recorded.

Mannion (1982) records at Linton Loch evidence other than a hiatus in the pollen record, with changes in sediment stratigraphy interpreted as changes in lake-level (Section 3.3.2). More recent studies (Milburn 1996; Smith 1996), although providing further lake-level fluctuation data, are not considered rigorous enough in terms of site selection and methodology to have generated a sensitive record of fluctuations in lake-level. Thus the record of lake-level fluctuations within Scotland is inadequate and inaccurate, with poor temporal and spatial resolution. Sites investigated to date are considered climatically insensitive with poor methodological and chronological controls.

BP C^{14}	Eastern Highlands O'Sullivan 1975a,1975b,1976	West Central Scotland Stewart *et al.* 1984	South East Scotland Mannion 1982	West Coast Scotland Boyd and Dickson 1987	Eastern Scotland Edwards and Whittington 1993	Eastern Lowlands Milburn 1996	North East Scotland Smith 1996
1 000				Open water Conditions- rise in lake levels.	Rise in lake level		
2 000					Hiatus in pollen record - lowered lake levels ?		
3 000		Hydroseral succession open water to shallow water to *Salix* carr.					Rise in lake levels
4 000					Water levels back to a steady state		Sedimentation hiatus - fall in lake levels
5 000		Increase in water levels	Fall in lake level	Fen peat accumulating		High lake levels with intersite variations in lake levels	
6 000			Subsequent rise in lake level	Open Water	Periods of shallowing water levels intermittent in nature - *Alnus* carr growth at some stages		
7 000	Falling water levels		Fall in Lake level			Period of abrupt falls and rises in lake levels	
8 000					High lake levels		
9 000						Low lake levels	

Table 1.4 Scottish lake-level fluctuations, as compiled by Harrison *et al.* (1996) and Yu and Harrison (1995) (see text) with additional data from Milburn (1996) and Smith (1996).

This record should not be considered as an accurate record of changes in lake-level for Scotland, and its use in palaeoclimatic interpretation is inappropriate (*cf.* Anderson *et al.* 1998). Continuous records of fluctuations in lake-level can be obtained from climatically sensitive sites (Section 3.2) with a rigorous methodology used to define the record of changes in lake-level (Section 3.3). Such climatically sensitive records of fluctuating lake-level can then be used to generate a record of shifts in precipitation.

1.6.2 OMBROTROPHIC BLANKET MIRE STRATIGRAPHY

Changes in ombrotrophic (rain-fed) raised mire stratigraphy have been used for over 100 years as an indicator of climatic change. The identification of recurrence surfaces or periods of peat rejuvenation were associated with a wetter climate and increased peat growth (reviewed in Barber 1982). The advent of radiocarbon dating indicated that these recurrence surfaces, the most notable being the 'Grenzhorizont', did not represent synchronous climatic events across north and western Europe. Local hydrological controls were considered responsible for this asynchroneity between sites. A move away from identifying fixed points within the peat stratigraphy saw the development of a series of analytical techniques to define a continuous record of changes in mire stratigraphy. Barber (1981) demonstrated a clear relationship between mire plant macrofossils, in particular *Sphagnum* remains, and climatically controlled changes in mire surface vegetation. Climatic data from such vegetation shifts have been further refined using multivariate statistical analysis (Barber *et al.* 1994a, b; Mauquoy and Barber 1999a).

Humification analysis provides a measure of the amount of peat decay (Sections 5.4, 8.1.2). A relationship between the amount of decay and mire surface wetness at the time of deposition was demonstrated by Aaby and Tauber (1975) and Aaby (1976). Such changes in mire surface wetness as recorded from an ombrotrophic raised mire were interpreted as climatically forced changes in the water table in response to the amount of effective precipitation, (precipitation minus evaporation) with both temperature and precipitation operating as climatic controls (*cf.* Barber *et al.* 1994b; Mauquoy and Barber 1999a, b). Several high-resolution climatic records have been generated from ombrotrophic raised mire sequences (*cf.* Barber *et al.* 1994a, b). However, the build up of a raised mire system requires the development of an isolated or perched water table to ensure peat accumulation and this restricts such mires to specific climatic and hydrological locations. No such raised mire systems are found in north west Scotland.

Chambers (1984) and Blackford (1990) developed the application of determination of peat humification to ombrotrophic blanket mire systems. Blanket mires are much more extensive than raised mires, across much of north west Scotland. Macrofossil analysis as a proxy of climatically controlled changes in surface wetness within blanket mires is limited, owing to the highly humified nature of such peat and the more restricted number of species associated with these mire systems. Testate amoebae analysis as an indicator of changes in mire hydrology is still under development with raised mires (Maquoy and Barber 1999b, Woodland *et al.* 1998) and remains untested within blanket mires. However, humification analysis from water shedding sites has been shown to generate a record of changes in peat decay sensitive to changes in mire surface wetness at the time of deposition (Chambers 1984; Blackford and Chambers 1991, 1995). Critical to obtaining such a sensitive record is site selection. Blanket mire sites must be water shedding with precipitation the sole hydrological input. Hydrologically isolated ombrotrophic blanket mires ensure that groundwater does not control the water table and in this study site selection criteria ensure that local controls on temperature and precipitation such as altitude are limited (Section 8.1.2). Thus although humification is a semi-quantifiable measure of surface wetness (Charman *et al.* 1999) this technique with the correct methodological approach is considered a sensitive measure of climatically controlled changes in mire surface wetness. Blanket mire humification has been used as a proxy climatic indicator at several sites within north west Scotland (Anderson 1996, 1998; Binney 1997; Anderson *et al.* 1998); these data are reviewed in Section 1.4. The review discussed the limitations of this proxy (non-specific climatic indicator) and the need to investigate climatically sensitive sites (see Sections 8.1.1, 8.1.2 and 8.1.3 for further discussion).

1.7 DEFINING THE CLIMATIC RECORD

Two independent and unambiguous climate proxies are used in this thesis to generate a record of Holocene climatic change within Glen Affric: lake-level fluctuations (Chapters 5, 6 and 7) and ombrotrophic mire stratigraphy (Chapter 8). These proxies have differing climatic sensitivities, with fluctuations in lake-level a response to changes in precipitation and the record of changes in mire surface wetness a response to changes in effective precipitation. Critical to obtaining maximum climatic sensitivity from both proxies is site selection as discussed in Chapters 3, 4 and 8. Good chronological control (Section 5.6) within both records ensures that recorded shifts in climate can be correlated between proxies. The combination of both data sets should define climatic change in terms of intensity of shifts in temperature or precipitation (Chapter 9).

Ombrotrophic blanket mires sites were selected to test for the role of local climatic controls such as west-east climatic gradients within Glen Affric (Chapter 8). Synchroneity of shifts in effective precipitation were to be considered as a response to climatic shifts operating throughout the glen with such shifts further defined in terms of temperature and precipitation by correlation with the lake-level record (Chapter 9).

The climatic record of either shifts in precipitation or temperature with shifts defined in terms of intensity, spatial and temporal magnitude were to be compared to other records of climatic change within north west Scotland, to define potential regional climatic shifts (Chapter 9). Potential climatic forcing mechanisms such as changes in the North Atlantic Ocean circulation (Bond *et al.* 1997; Bianchi and McCave 1999), are then evaluated (Chapter 9).

PART 1 INTRODUCTION

CHAPTER 2
GLEN AFFRIC: PHYSICAL DESCRIPTION

2.1 INTRODUCTION

North West Scotland is a region of high relief dissected by deep broad river valleys; Glen Affric typifies such a landscape. Glen Affric lies 25 km southwest of Inverness, at 57° 15' N and between 5° 05 to 5° 17'W (Figure 2.1). Peaks of over 900m above sea level (asl) dominate the areas of high relief within the Glen Affric region, with the Five Sisters of Kintail (> 1000m asl) to the west (Figure 2.2). The watershed of the River Affric defines Glen Affric. The river rises in mountains to the west (Ben Attow) and southwest (Sgurr an Fhurail and Ciste Dubh) and flows eastwards along Glen Affric (Figure 2.2) into Loch Affric and Loch Beinn a' Mheadhoin. The Glen itself forms part of a series of deep valleys running south-west north-east through the mountains. Glen Affric connects to the west through a narrow steep valley, the upper reaches of Gleann Lichd, forming a low level route through the mountains, stretching from Loch Duich on the west coast through inland to Cannich in the east (Figure 2.2).

2.2 GEOLOGY

The underlying geology controls aspects of the hydrology, rates of erosion, sediment type, soil type and vegetation cover. Glen Affric lies in an area of northwest Scotland dominated by rocks of the Precambrian Moinian Supergroup. These comprise of a series of metamorphosed and highly folded sedimentary rocks, mica-shists or pelites (shales), granular quartzo-feldspathic rock or psammites (sandstones) and quartz-mica-schist or semi-pelites (sandy shales or siltstones) (Harris and Johnson 1991). Within the Moinian Supergroup, the Glenfinnan Group underlies the Glen Affric area (Figure 2.3) and is typically made up of interbanded psammites and pelites.

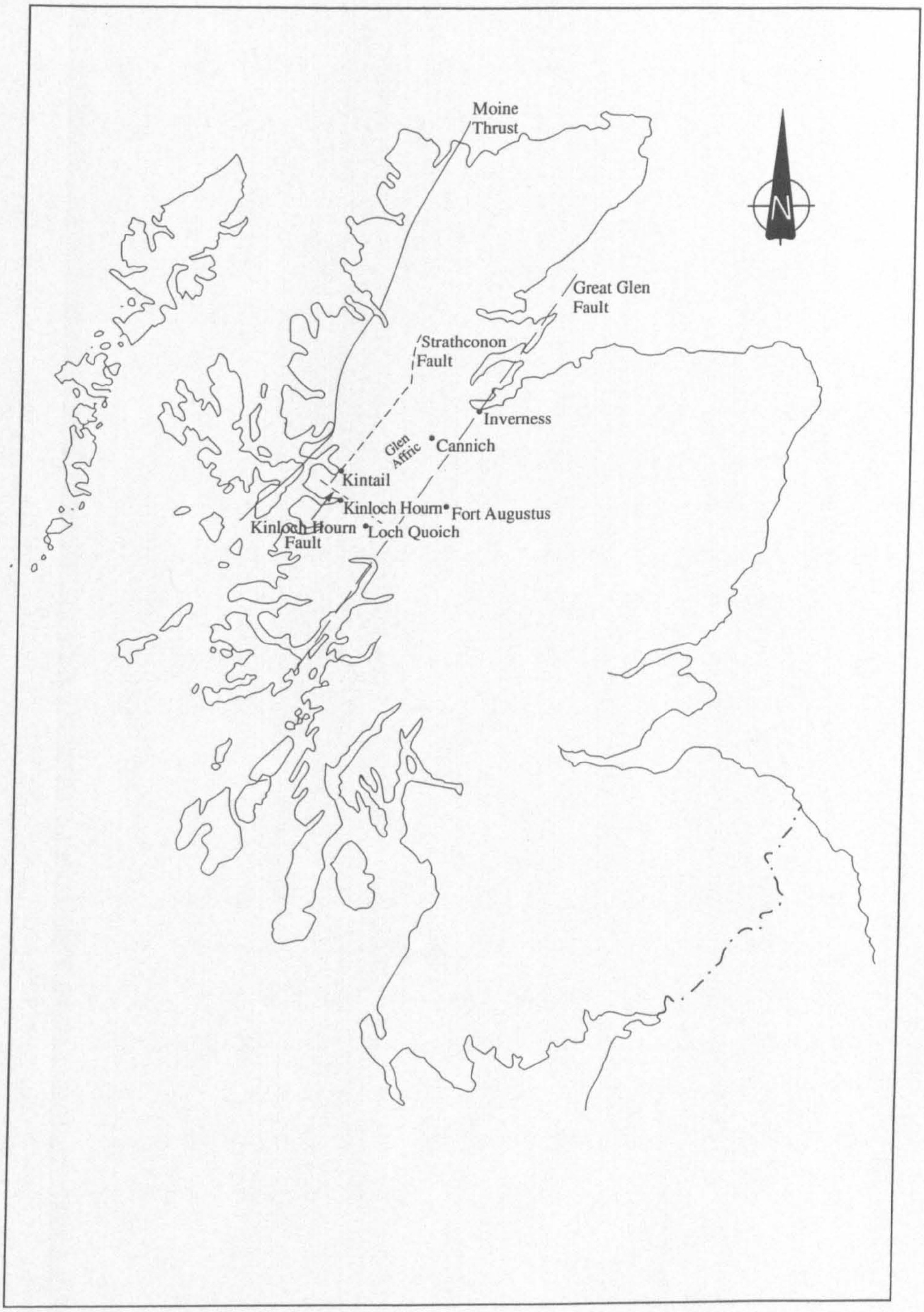


Figure 2.1 Locations of major geological faults within north west Scotland and the Glen Affric region (Section 2.2). Weather station sites sites discussed in Section 2.6 are also shown.

The mineralogy of these metamorphic quartz-mica rich rocks is of low base status (acidic). The metamorphic fabric and crystalline structure of the minerals makes these rocks highly impermeable (Read and Watson 1978).

Metamorphism

The Glenfinnan group was folded and deformed during at least two episodes of regional metamorphism (Harris and Johnson 1991).

(1) The first episode occurred around 1000 Ma with the interfolding of the Moinian rocks with the older Lewisan basement rocks beneath (granulitic gneiss) (Figure 2.3). Throughout this period and at later stages some local pegmatites and granites were intruded.

(2) The second and most intense period of folding and deformation occurred during the Caledonian Orogeny (450 Ma). During the orogeny the Moinian metasediments were stacked up along NW trending thrust faults. The youngest of these thrust zones is the Moine Thrust (430 Ma) which lies to the north west of Glen Affric (Figure 2.1).

Igneous Activity

Igneous activity associated with the Caledonian Orogeny is limited within the Glen Affric area with outcropping small dykes mainly of microdiorite with some dykes of appinite and felsite also noted. A sequence of more basic dykes represented by hornblende schists and ampibolites was also intruded locally during or prior to the main Caledonian deformation. Later igneous intrusions such as lampophyre dykes (lower Devonian 360 Ma) and camptonite (Late Carboniferous/Permian300 Ma) are recorded (British Geological Survey (Scotland) 1986).

Fault Activity

Faulting within the Glen Affric region is associated with the Caledonian Orogeny as the psammites and pelites are displaced along local faults which trend west-north-west/east-south-east (Davenport *et al.* 1989) (Figure 2.3).

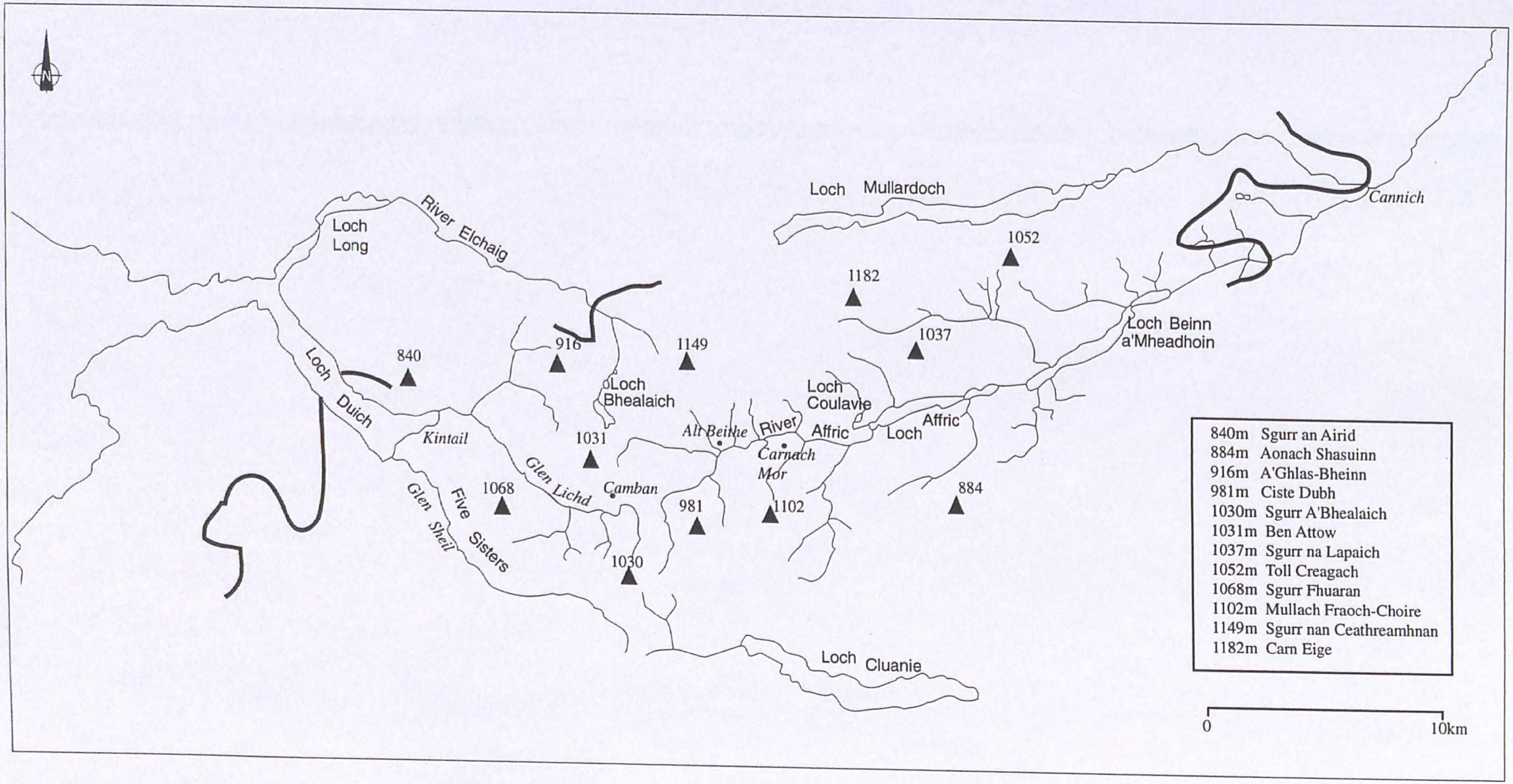


Figure 2.2 Physical features and site locations (see text) within Glen Affric. Known Loch Lomond Readvance ice limits are marked for Loch Duich and the eastern end of Glen Affric (redrawn from Bennet and Boulton 1993).

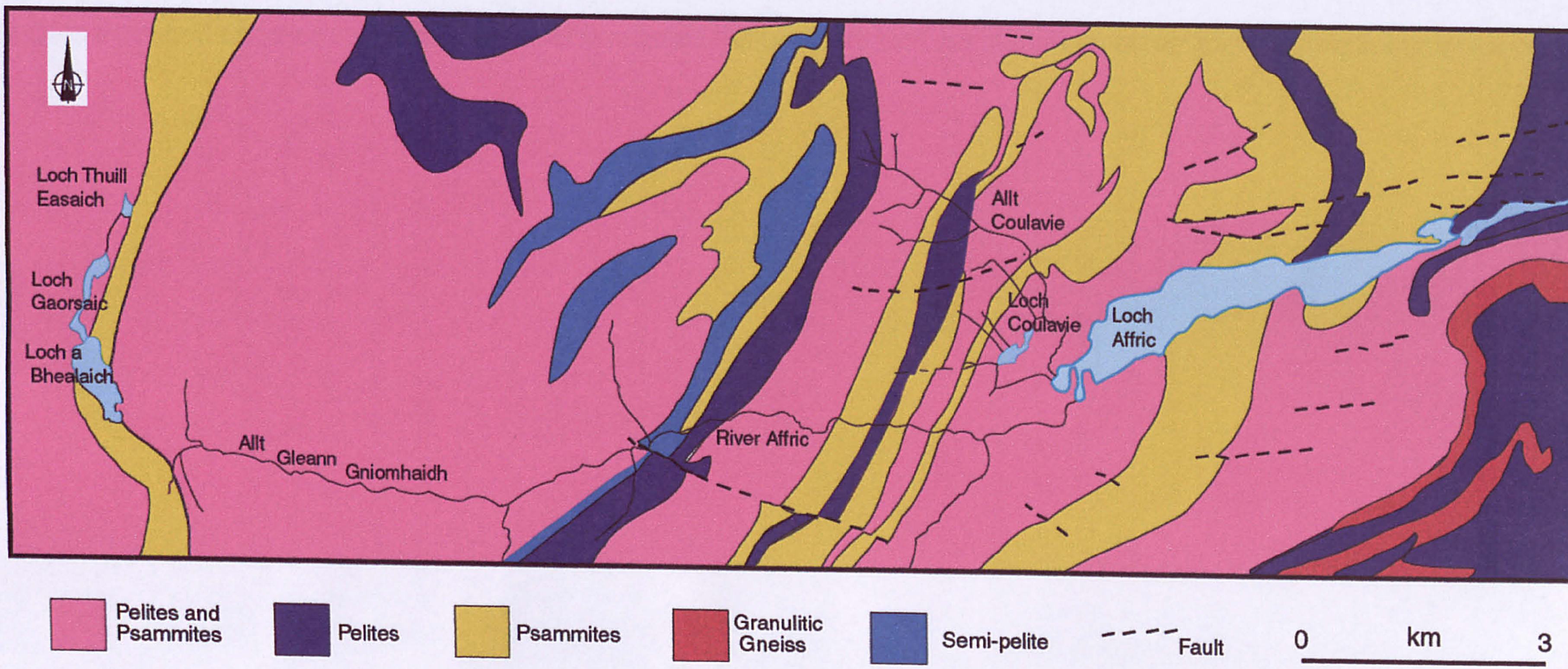


Figure 2.3 Solid geology of the Glen Affric region (Section 2.2). Redrawn from British Geological Survey (Scotland) (1986).

The Kinloch Hourn fault (which lies to the south west of Glen Affric) and the Strathconon Fault (to the north), are major regional faults lying within the metamorphic Moine basement and are associated with the Caledonian Orogeny (Figure 2.1). Seismic investigations indicate that these faults are sites of recent (last 30 years) seismic activity concentrated at shallow depths (<15km) and controlled by tectonic stress (Davenport *et al* 1989). Ringrose (1989) suggested that surficial drainage displacement observed along the Kinloch Hourn Fault is as a result of several successive fault displacements. The timing of these inferred fault movements is not established as the age of the drainage system is not known. Ringrose suggested, however, that these displacements are potentially postglacial (after 15 000 BP) in age as minor drainage systems would have been reset by the ice cover during the last glaciation. Ringrose (1989) also cites evidence for much more recent fault activity along the Kinloch Hourn fault with evidence for fault gouge emplacement into *in situ* peat after 2400 BP (2450 cal. BP) and seismically triggered soft sediment (lacustrine) deformation before 3500 BP (3900 cal. BP).

Davenport *et al.* (1989) and Ringrose (1989) propose that palaeoseismic events in Scotland are spatially and temporally associated with the enhanced potential for faulting at ice margins as a result of rapid isostatic rebound during deglaciation. Davenport *et al.* (1989) propose that high-magnitude (magnitude 6.5-7.0) seismic events occurred around 10 000 years ago and are associated with Loch Lomond Readvance ice limits. However, direct radiocarbon dated evidence for this early Holocene seismic activity is not well documented and further research into timing of proposed tectonic events such as rockfalls and landslides is required. It is difficult to infer seismic activity as triggered by Loch Lomond Readvance glacial rebound as a forcing mechanism for the recent seismic activity (2400-3500 BP (2450-3900 cal. BP) discussed above. Therefore, within Glen Affric, local faults within the metamorphic Moine sediments could have been active during the Holocene with seismic activity triggered either by enhanced potential faulting during isostatic rebound associated with the Loch Lomond Readvance (Section 2.3) or through the effects of underlying regional tectonic stress.

2.3 GLACIAL HISTORY

The following section is a brief evaluation of glacio-geomorphic activity and landscape evolution within the Glen Affric region. The climatic sensitivity of north west Scotland to changes in Atlantic Ocean (sea surface temperatures and the position of the Atlantic Polar Front) and atmospheric circulation patterns (Section 1.2) has ensured that throughout the Quaternary, under favourable climatic conditions, ice would have accumulated within this region. In the last 800 ka years there have been at least 10 cycles of glacial and interglacial periods and during the whole of the Quaternary there are considered to have been around fifty of these events (Shackleton *et al.* 1990).

Ice masses on the west coast of Scotland are considered to have been warm-based and fast moving and therefore highly erosive (Ballantyne *et al.* 1998). Within the Glen Affric region of Scotland the underlying structure of the Moinian rocks in part controlled the passage of ice across the region. Lines of weakness within the folded Moinian meta-sediments were exploited by the successive periods of ice cover. Ice would also have moved down the ancient valleys and watersheds developed in the west of Scotland by the early Tertiary. The landscape within the Glen Affric region is the result of several periods of glaciation and associated types of glacial processes (Haynes 1995).

In the last glacial period, the Late Devensian, ice began to retreat from Scotland at around 18-14 000 BP when the climate was still cold but under conditions of probable precipitation starvation (Sutherland 1993). The period of ice retreat was regionally complex with ice readvances such as the Wester Ross Readvance, the timing of which is not secure but Ballantyne (1993) and Ballantyne and Sutherland (1987) infer a date of 13.5-13 000 BP. This regionality reflects the interaction of ice masses, increased precipitation and the position of the North Atlantic Polar Front as it shifted northwards during deglaciation (Ballantyne and Sutherland 1987; Clapperton 1997).

Ice had disappeared from the lowlands and from lower ground in the Highlands by around 13 000 BP (Hall 1997). Rapid climatic amelioration is associated with the opening of the Lateglacial Interstadial (at or around 13 000 BP) with mean July temperatures considered to have been similar to those of the present day, 14-16°C

(Atkinson *et al.* 1987). Based on the palaeoclimatic reconstructions generated by Atkinson *et al.* (1987), Ballantyne and Sutherland (1987) propose for the mountains in Wester Ross, within high-altitude corries, that the climate would have been suitable for the year-round presence of ice. It is suggested that for other upland regions of Scotland, ice could have been present throughout the Lateglacial Interstadial (Clapperton 1997; Ballantyne *et al.* 1987).

The period of climatic amelioration during the Lateglacial Interstadial was brief with the cooling trend that culminated in the Loch Lomond Stadial starting shortly after 11 000 BP (Hall 1997).

2.3.1 THE LOCH LOMOND STADIAL

The Loch Lomond Stadial (Younger Dryas) lasted around 800-1000 ^{14}C years and marked the return of ice to regions of Scotland. In the north west Highlands of Scotland ice started to accumulate in the centres associated with previous ice sheets.

Within Glen Affric there is some difficulty in defining Loch Lomond Readvance ice limits. Pennington (1977) suggested from sediment stratigraphy in Loch Affric that the lake had been within the limits. Ice limits suggested by Bennett and Boulton (1993) and Sutherland (1993) indicate that the western ice limit lay along the west coast (Loch Duich) and the eastern limit lay in the River Affric valley 5 km west of Cannich (Figure 2.2). In inferring these limits, however, Bennett and Boulton (1993) indicate that problems were encountered owing to the proliferation of possible limits. The glacial limits chosen were inferred from the moraine fragments that showed the strongest lateral continuity. Tate (1995) extended the eastern limit within Glen Affric by examining the pollen record from several sites in and around the proposed limit (Figure 2.2). Difficulties encountered with the location of critical pollen sites has meant that this limit could not be further defined (Tate 1995). Within upper Glen Affric the ice was thought to have been limited to the main valleys and also active within the upland corries (Bennett and Boulton 1993). The higher peaks are thought to have remained ice free but would have been subjected to periglacial activity and deep frost weathering (Tate 1995).

Rapid climatic amelioration brought the Loch Lomond Stadial to an abrupt end with the ice masses decaying as temperatures returned over a period of decades to those similar to the present day (Atkinson *et al.* 1987; Dansgaard *et al.* 1989; Taylor *et al.* 1993).

Field observations within Glen Affric note that glacial tills associated with ice decay have been deposited as small terminal moraines (GR NH061200) and very localised patches of hummocky moraine within the main valley floors and along the valley sides. Tills are generally poorly sorted and consist of locally derived mica-and quartz-rich metamorphic bedrock blocks within a finer clay sand matrix (Figure 4.4). Fluvioglacial deposits are not noted within upper Glen Affric (Tate 1995) and tend to be located at the western coastal margins and in the east (Sutherland 1993). Features such as rouches moutonnées (Figures 4.3, 8.3) and moulded ice scoured bedrock (Figure 4.4) would suggest that glacial erosion rather than deposition was predominant.

Debate within the literature on the nature of Loch Lomond Readvance ice decay and deglaciation is based on the interpretation of end moraines and till deposits. Detailed mapping of moraines in central Skye, west of Glen Affric, was interpreted by Benn *et al.* (1992) as evidence for two distinct stages of deglaciation: (1) an initial period of ice-margin fluctuation; and (2) a period of uninterrupted retreat accompanied by local *in situ* ice stagnation. Benn *et al.* (1992) stressed that glacial features classed as 'hummocky moraines' cannot be assumed to share a common origin, with field mapping and interpretation needed to fully understand local climatic controls and ice dynamics. Bennett and Boulton (1993), from aerial photographic interpretation, mapped areas of hummocky and linear moraines within the Glen Affric region and have interpreted these structures as representing a series of Loch Lomond Readvance ice-retreat events. The authors propose that for this mainland region, ice decayed in an active and progressive manner with no *in situ* stagnation event noted at a scale similar to that proposed by Benn *et al.* (1992). The 'hummocky and linear moraines' interpreted by Bennett and Boulton (1993) for Glen Affric are noted from field observations to consist of a series of ice-smoothed bedrock knolls (Figures 4.3, 8.3), suggesting that for within Glen Affric the evidence for a series of ice retreat events is not secure.

2.4 CURRENT GEOMORPHIC ACTIVITY.

The geomorphic features and associated geomorphic activity within Glen Affric are essentially a legacy of glacial processes, such as those outlined above. Section 2.3.1 suggests that within the limits of the most recent period of ice cover within Glen Affric, the Loch Lomond Readvance, ice would have removed any glacial features associated with Late Devensian glacial activity. Therefore, glacio-geomorphic features within much of Glen Affric are thought to date from the retreat of the Loch Lomond Readvance ice, and reworking of sediment within these features is thought to have occurred during the Holocene. This section is a brief description of some of the main geomorphic features noted within Glen Affric because geomorphic activity within a lake catchment can represent a major sediment input (Section 3.2.2). A further review of the potential timing and forcing mechanisms of these events is provided with discussion of geomorphic activity and lake sedimentation in Chapter 9.

- Talus slopes derived from rock-fall debris have accumulated beneath steep corrie back walls. The metamorphosed sediments within Glen Affric are subject to frost shattering along foliations within the rock structure and thus large amounts of rock-fall debris have accumulated beneath exposed rock surfaces. These features are considered to be relict although reworking of the debris is occurring at some sites (see below).
- Large debris cones accumulated at the foot of rock gullies are the result of reworking of rock-fall derived debris and frost shattered debris from the upper slopes, for example in Coire Cròm (NH1520) above Loch Affric. There is also evidence for present-day reworking of rock debris within the larger debris cones (Figure 4.3) (*cf.* Hinchcliffe 1999).
- Numerous smaller scale debris flows and rotational slumps are noted along the valley sides and are associated with smaller gullies and the reworking of till on these slopes. Many of these small-scale features are now vegetated but some may have been active as recently as the last few 100 years (*cf.* Innes 1983, 1997) with reworked debris noted close to the surface (Ward 1997). These smaller scale rapid mass movements are thought to be triggered by single extreme events such as intense rainstorms (Berrisford and Matthews 1997; Curry 2000; Hinchliffe 1999).

- Along the valley floor there are several large alluvial fans notably above Camban (NH 054184), Alt Beithe (NH 079202) and Loch Coulavie (NH 135216) (Figures 2.2, 4.1, 4.3). Sediment sources for these fans are thought to be frost-shattered debris from higher slopes above and the reworking of glacial till plastered along the valley sides. Many of these fans have coalesced and are at present well vegetated, indicating that only during periods of high sediment loads and water discharge is there any addition of large amounts of sediment to these fans. Small streams flow across the surface of these fans, reworking the surface deposits. Finer grained sands are deposited out across the fan surface during periods of seasonally higher stream discharge.
- Rock falls of various magnitude are noted throughout Glen Affric. Larger scale rock slope failure features are noted throughout much of Scotland, located near Loch Lomond Readvance ice limits (Ballantyne 1986; 1997). Schists with pronounced foliation planes, such as those within Glen Affric, will readily form shear stresses within the rock. The most likely trigger for these rock slope failures is high-magnitude seismic activity, potentially the result of rapid post-glacial uplift (Section 2.2) associated with the retreat of the Loch Lomond Readvance ice (Ballantyne 1997; Davenport *et al.* 1989).
- The River Affric and its tributaries drain much of the Affric valley eastward into Loch Affric. Sedimentological evidence from a palaeochannel in the valley floor at Carnach Mor (Figure 2.2) suggests that from the earliest Holocene (c. 9 500 BP (10 600 cal. BP)) the river has been meandering across the valley floor within predominantly peaty floodplain sediments (Davies 1999). Gravel terrace fills, essentially sequences of gravel, with thin overbank sediments and peat appear to be only of late Holocene age. The terraces, although of limited height indicate that the river has further incised into its floodplain during periods of increased discharge or sediment supply (Johnston 2000).

2.5 SOILS

The geology of the Glen Affric region is dominated by acidic rocks (Section 2.2) which produce soils with a low base status and low availability of plant nutrients. The glacial till deposits within the valley are also predominantly derived from these acidic rocks.

The high humidity of the region (Section 2.5.3) leads to the incomplete breakdown of vegetation, the accumulation of raw acid humus and development of peaty soil types (Davidson and Carter 1997). The combination of impermeable metamorphic acid rocks and the oceanic climate results in a landscape dominated by acid peaty soils and extensive blanket bog (Figures 4.3, 8.3, 8.10, 8.16, 8.22). The dominant soil-forming processes within this western region of Scotland are weathering, leaching, podzolization and gleying (Davidson and Carter 1997). The spatial complexity of soil type in Glen Affric reflects the interaction over time of soil parent material, topographic position, climate and vegetation. Evidence for Holocene pedogenesis is discussed in Section 8.6 where new data from this study are evaluated.

The following discussion focuses on present day distribution of the main soil types within Glen Affric (Bibby *et al.* 1982).

- In areas with little drift cover, essentially bedrock such as on parts of the valley floor, extensive peat deposits have accumulated within the hollows and over bedrock knolls (Figures 4.3, 8.3, 8.10).
- Tills are predominantly composed of poorly sorted mica-schists and psammatic type rocks (Figure 4.4). These mounds of till are associated with the development of peaty podzols and peaty gleys. Peaty podzols and peaty gleys are also found on the more freely draining steeper rocky slopes.
- The flood plain is associated with alluvial deposits, sands and gravels with overbank deposits of silts. These sediments have developed peaty alluvial soils with some peats accumulating in the hollows; for example, infilling palaeochannels.
- Mineral alluvial soil types are associated with the alluvial fans that have accumulated on the valley slopes through the glen. These fans have accumulated through the reworking of glacial till deposited on the upper slopes. Minerogenic peats have accumulated on the fan surface and during periods of high stream discharge minerogenic sediments are flushed across the fans maintaining the high minerogenic content of the soil.

Within these soil types discussed there will be small-scale variations dependent on the nature of the parent material and microclimatic conditions associated with altitude, slope and aspect of slope.

2.6 PRESENT DAY CLIMATE

2.6.1 INTRODUCTION

Major climatic controls for the North West of Scotland are (a) proximity to the Atlantic Ocean and (b) topography. The influence of Atlantic Ocean thermohaline circulation with reference to palaeoclimatic changes in Scotland has been discussed in Section 1.2. The present-day influence of the northward flowing North Atlantic current, originating in the Gulf of Mexico, moderates winter temperatures, generating milder winters along the west coast of Scotland than would otherwise be expected from a region with a similar northerly latitude (Roy 1997). The greatest influence of the Atlantic Ocean is in the generation of cyclonic weather systems. The first landfall for these moisture-laden Atlantic air masses is the north west of Scotland. The interaction of these air masses and the topography control much of the climatic variation in the north west of Scotland.

Birse *et al.* (1970) interpreted climatic data generated from weather stations across Scotland and presented the data as thematic maps incorporating the main climatic variables, precipitation, temperature and wind speed. The distribution of weather stations, as used by Birse *et al.* (1970), is poor within the Glen Affric region, in particular for climate records at altitude. McClatchey (1996a) and Prudhome and Reed (1998) note that within the whole of Scotland there are few climatic data sets above 300 m above sea level (asl) with only 30 rainfall gauges above 500 m asl. Around Glen Affric the main weather stations are at Inverness, Loch Duich, Fort Augustus and Loch Quoich (Figure 2.1). Climate data from these stations are not always complete for all climatic variables, in particular for wind speed, and data are therefore interpolated between stations. Interpolation of data will incorporate local site conditions and so much of the small-scale climatic variability, the result of changes in topography and altitude, will be missing from data sets. The climate data sets presented by Birse *et al.* (1970) are averaged over the year and so much of the seasonal variation, in particular with windspeed and precipitation data, are also undetectable. Birse *et al.* (1970) give some indication of seasonal temperature variation through the calculation of accumulated temperature using different basal temperatures (Section 2.5.2). In the following discussion, general climatic descriptions for north western Scotland and the effects of local topography and altitude alongside data from Birse *et al.* (1970) are used to describe the climate within the Glen Affric region.

2.6.2 TEMPERATURE

Major controls on regional temperature variations within north west Scotland are altitude and proximity to the Atlantic Ocean. Typically the west of the country has summers around 0.5-1°C cooler than the east with August the warmest month (Roy 1997). Mean July temperatures at sea level in the west (Kintail) are 13.3°C and in the east (Inverness) 13.9°C (Birks 1996). In summer the east is warmer owing to the influence of continental air masses, with less cloud, whereas the west with predominantly westerly air masses remains relatively cloudy and subsequently cooler. Along the west coast of Scotland the proximity of the ocean and the North Atlantic current gives rise to mild winters and low numbers of frost days, with February the coldest month (Roy 1997). This ameliorating influence of the North Atlantic declines inland. The mean January temperature at sea level in the west (Kintail) is 4.4°C and 3.3°C in the east (Inverness) (Birks 1996).

Overlying these west-east temperature gradients is the influence of altitude. In general, temperatures decrease with altitude, typical lapse rates being 10.0-10.5 °C/km during daylight and 4.0-4.5°C/km during darkness (Harrison 1997). These values agree with lapse rates calculated for the Cairngorm Mountains with a annual mean rate of 7-10 °C/km (McClatchey 1996a). Lower temperatures at altitude are also reflected in winter airfrost data with the number of frost-free days decreasing by 17 days per 100 m of altitude (Harrison 1997).

Birse *et al.* (1970, page 1)) classify temperature in terms of accumulated temperature that is;

'the integrated excess or deficiency of temperature with reference to a fixed basal temperature expressed as the number of degree days over the period of one year'

As a measure of the potential growing season and summer warmth the basal temperature is 5.6°C, which represents the approximate level at which plant growth commences. A lapse rate of 0.6°C per 100 metres is used to control for temperature decreases with increasing altitude. This lapse rate may have underestimated the decrease in temperature with increasing altitude as suggested by the lapse rates discussed above.

The accumulated temperature data, as generated by Birse *et al.* (1970), suggest both oceanic and altitudinal controls on temperature, as discussed above, within Glen Affric. The western coastal fringes and the area around Cannich to the east are classed as cool and have 1100-1375°C days. Further east the Inverness coastal area is classed as warm with >1375°C days. Upland areas at around 200m altitude are classed as cool becoming cold up to 800 m in altitude with 550-825°C days. Areas over 800 m altitude are classed as very cold with 275-550°C days and peaks over 1000 m altitude are classed as extremely cold with 0-275°C days.

To indicate winter severity, the accumulated temperature with a basal datum of 0°C is calculated for a period of one year and again is expressed in degree days (Birse *et al.* 1970). A lapse rate of 0.65°C per 100 metres is used. The western coastal fringes are classified as having moderate winters, reflecting the moderating influence of the ocean. Cannich to the east has moderate winters (50-110 °C days), with Inverness having fairly mild winters (20-50°C days). Again, accumulated temperature values reflect an altitudinal control on temperature. Areas along the valley floor at 200 m and, on the valley sides up to 800m in altitude, are classed as having rather severe winters with 110-230 °C days. Above 800 metres in altitude the winters are classed as very severe with 230-470 °C days and areas above 1000 metres in altitude classed as having extremely severe winters with >470°C days.

This type of climate classification (Birse *et al.* 1970) does not take into account some of the local and seasonal controls on temperature:

- Valleys surrounded by high mountains have lower winter temperatures due to katabatic cooling. Cold air from the snow-covered mountains flows down hill into the valley floor lowering the air temperature and often creating frost hollows (Harrison 1997).
- Snow accumulation will depress the average air temperatures in particular during spring where in upland areas snow lie is extensive. This could result in a seasonally greater temperature decrease with altitude- a higher lapse rate as discussed above.

- Steep sided valleys and in particular north-facing slopes will receive less sunlight during the winter owing to the low angle of the sun and so these slopes will be in the shade for lengthy periods (Harrison 1997). At altitude cooler air temperatures will aid in the preservation of snow patches in the north-facing corries.
- Deep valleys with large lochs which do not freeze in the winter may be warmed locally by a few degrees. The unfrozen lake water being circulated through the water column is at about 4°C. The ameliorating effects of this locally warmer air are confined to areas close to the loch (Roy 1997).

2.6.3 PRECIPITATION

Rainfall in the west of Scotland is associated with west-south-westerly airflows bringing in frontal depressions from the Atlantic Ocean. A west-east rainfall gradient exists across Scotland as the influence of the westerly cyclonic systems on the amount of precipitation declines further east. Along the western coastal fringes annual average rainfall figures are around 1500 mm per annum (pa). For coastal mountainous regions the annual rainfall figure rises to around 3000 mm pa. Towards the east coast of Scotland in the "rain shadow" the annual rainfall figure falls to around 750 mm pa (Birks 1996).

High amounts of rainfall occur during the autumn and early winter when westerly cyclonic systems are most frequent and most vigorous. Lowest rainfalls occur in April, May and June when the climate is frequently dominated by more continental anticyclone systems from the south and east (Roy 1997).

Orographic enhancement of rainfall over the mountains on the west coast of Scotland exacerbates the west-east precipitation gradient. The amount of rainfall induced by these west-south-westerly air flows over the mountainous areas of western Scotland is four to eight times greater than in eastern Scotland (Weston and Roy 1994).

Overall rainfall increases with altitude. In the west the increase is around 2.5 mm/m but further east in the "rain shadow" of the dominant westerlies the increase is *c.*1.0 mm/m (Weston and Roy 1994). In Wester Ross studies suggest this increase with altitude is as high as 4.5 mm/m (McClatchey 1996b). In this region as in others along the west coast the mountains rise very steeply from sea level and so force the moist air upwards very rapidly, generating higher rainfall increases with altitude.

The above discussion suggests that the annual rainfall figures quoted above may have underestimated the total amount of rainfall for regions at high altitude. Rainfall patterns over short time periods- for example- seasonal extremes in precipitation, are not discernible from the mean annual rainfall data. Topography, direction and distance to moisture sources will control rainfall patterns. For western Scotland the primary moisture source is the Atlantic Ocean (Prudhomme and Reed 1998). This leads to an inherent spatial complexity of rainfall patterns within a region of high and varied relief such as Glen Affric.

A reduction in the number of westerlies and an increase in more continental anticyclonic weather systems, where the airflow tends to be from an easterly direction, results in higher rainfall in the east than in the west. If this easterly continental airflow persists then this can give rise to drought conditions in the west and north west (Roy 1997).

The amount of snowfall and the length of time snow exists are the product of complex interactions between topography, distance from the coast and nature and direction of snow-bearing systems (Harrison 1997). Snowfalls in Scotland occur from November through to April with the snowiest month being January (Roy 1997). With altitude the duration of snow cover increases by 15-20 days per 100 metres and on average the snow cover on upland areas is 200 days in the year (Harrison 1997). Snow is less likely to fall and lie on the west coast as the typical south-west or westerly winds are not cold enough to give snow over lower ground. Snowfall and snow lie are therefore more likely with increasing altitude and in areas further east. Snow lie at altitude in combination with persistent cooler temperatures will generate further cooling due to katabatic winds (Section 2.6.2). Snow at altitude is redistributed by the wind stripping snow from exposed ridges and slopes, leading to redeposition and accumulation of snow in the northern and eastern corries and persisting as summer snow patches (Section 2.6.2).

Rainfall data generated by Birse *et al.* (1970) suggest that nowhere on the west coast of Scotland does evaporation exceed precipitation. Only further east do values record a Potential Water Deficit (PWD) where evaporation exceeds precipitation. This situation of high humidity in the west is maintained by the high rainfalls associated with the west-south-westerly air flow. Within Glen Affric only in the east, near Cannich, is there a PWD value of 0-25 mm and further east towards the Inverness coastal area the PWD rises to >75 mm (Birse *et al.* 1970). To define rainfall and evaporation patterns further Birse *et al.* (1970) map the amount of summer rainfall (April to September) in millimetres that exceeds summer evaporation. The mapped 500 mm isopleth in Glen Affric, running north-south, falls across the head of Loch Beinn a' Mheadhoin arbitrarily splitting the glen into a much wetter western section and a much drier eastern section. Within the glen there is therefore a marked west-east precipitation gradient with a seasonally drier eastern section.

There are no snowfall data for Glen Affric. The proximity to the west coast and dominance of westerly airmasses would indicate that snowfall and snow lie at low level is not that common. With increasing altitude on the higher mountains within the Glen, snowfall and snow lie would be throughout most of the winter and early spring months. Within the Glen, from personal observation, snow patches on the higher mountains (above 800 m asl) persist into June and early July.

2.6.4 WIND

The western fringes of Scotland are fully exposed to the Atlantic Ocean and tend to be the windiest region in Scotland with the highest frequency of strong winds and gales. The windiest months are November and March and are associated with periods of intense cyclonic activity as the passing of deep depressions will bring the stormiest weather. The dominance of these cyclonic weather systems results in a prevailing wind direction of broadly south-westerly to westerly (Harrison 1997).

In general there is an increase in windspeed with altitude with an increase of 6-9 m/s/km typical for upland regions in the British Isles (McClatchey 1996a). Wind speeds and patterns are topographically controlled and in steep-sided glens can be complex with reverse circulations, downslope winds and winds being funnelled down valleys (Roy 1997).

Birse *et al.* (1970) have interpolated windspeed data with altitude from a limited number of weather stations to produce windspeed data for Scotland (Figure 2.1, Section 2.6.1). Within Glen Affric the windspeed data reflect the increase in windspeed with altitude. Windspeeds along the valley floor (~200 m asl) are given as 2.6-4.4 m/s, with altitude this rises to 4.4-6.2 m/s at 400 m asl and up to 6.2-8.0 m/s at around 800 m asl. At greater than 800 m asl, including peaks over 1000 m, the windspeed is given as >8 m/s. These are mean values over a one-year period and so hide extreme seasonal events, in particular during storms associated with the windiest months of November and March. Wind patterns will also vary diurnally with the passage of weather systems altering wind direction and speed. Interpolation of the windspeed data-set spatially homogenises the data with local topographic controls on windspeed and direction, such as local shelter due to either the topography or tree cover and down-valley wind funnelling, concealed.

2.6.5 SUMMARY

Glen Affric shares the same general climate trends as suggested by the regional data for the main climatic variables (Roy 1997). Climate within the Glen has a strong oceanic influence with the prevailing westerly airflow and associated cyclonic Atlantic depressions dominating the climate patterns. Within the glen the main climatic gradients are:

- a west-east summer potential evaporation gradient controlled by variations in the amount of summer precipitation and evaporation;
- variations in temperature, wind speed and rainfall with altitude;
- localised temperature amelioration associated with proximity to Atlantic Ocean.

The surface topography within the glen results in a spatially complex climate with marked differences over short distances. Factors such as altitude, slope, aspect, topographic exposure, local drainage conditions and soil thermal properties can generate microclimatic variations over single metres in area (Harrison 1997; Roy 1997).

2.7 VEGETATION COVER

The present vegetation cover through Glen Affric reflects climatic gradients, soil type and human activity. Birse *et al.* (1970) defined bioclimatic sub-regions of Scotland based on climatic data (accumulated temperature, accumulated frost, exposure and potential water deficit). Within Glen Affric the main divisions reflect the oceanicity of climate and, in particular, divisions based on the 500 mm isopleth of excess summer precipitation over summer evaporation (Section 2.6.2). The ameliorating influence of the ocean on air temperature is restricted to the western coastal fringes. The west coast is divided from the rest of the region and classed as temperate incorporating a longer growing season and lower humidty. In association with humidity, altitude was used by Birse *et al.* 1970 to classify the bioclimatic sub-regions, with areas of increasing altitude classed as boreal reflecting cooler conditions and shorter growing days 675-1375 °C days. Mountain tops were classed as arctic, with cold to very cold conditions and a short growing season <675 °C days. These are very broad classifications and as discussed in Section 2.6 there is spatial complexity in climatic conditions and therefore in the growing conditions suitable for differing vegetation types. The critical climatic gradients appear to be those associated with altitude, in particular exposure and precipitation with potential evaporation during the summer a key climatic variable.

Soil type is also a control on vegetation type. As outlined in Section 2.5 the dominant soil type within the region is peat with local variations reflecting changes in parent material, altitude and minerogenic sediment supply. The vegetation noted within Glen Affric today is closely associated with these local variations in soil type but the role of human impact, in particular historic and prehistoric human impact, is also responsible for the present vegetation cover (Chapter 4).

Glen Affric can be broadly divided into east and west Glen Affric based on the vegetation cover (Macaulay Research Landuse and Research Institute, Landcover of Scotland 1996). The east of the glen, from the head of Loch Affric to Loch Beinn a'Mheadhoin, along the loch shores is dominated by Scots Pine (*Pinus sylvestris)* and birch *(Betula)* woodland with a tall heather (*Calluna vulgaris* and *Erica* sp.) understorey. The west of the glen is in comparison virtually treeless, with trees restricted to areas out of reach of grazing animals. Along mountain stream gullies and inaccessible banks of the River Affric, species such as rowan (*Sorbus*), willow (*Salix*), and birch (*Betula*) are noted with a few sparse Scots Pine (*Pinus sylvestris)* near Loch Affric. In areas of extensive low altitude, blanket peat vegetation is classed as undifferentiated heather moor and is dominated by species tolerant of acid water-logged conditions such as flying bent (*Molinia caerulea*), deer grass (*Trichophorum cespitosum*), bog mosses (*Sphagnum* sp.), cotton grass (*Eriophorum vaginatum*), bog heather (*Erica tetralix*) and heather (*Calluna vulgaris*). Where soils are less water-logged (peaty podsols) vegetation cover is classified as coarse grassland with species such as white bent (*Nardus stricta*) and flying bent (*Molinia caerula*). At higher altitudes the soils are classified as sub-alpine with some peats, while the vegetation is classified as montane vegetation with sedges such as *Carex nigra*. In the valley floor the more minerogenic alluvial sediments and the alluvial fans are associated with herbaceous grassland cover.

PART 2 LAKE-LEVEL FLUCTUATIONS

CHAPTER 3
SITE SELECTION AND METHODOLOGY

3.1 INTRODUCTION

Previous discussion (Chapter 1) has already considered lake-level fluctuations as a climate proxy. In this chapter the methodologies behind site selection and the identification of lake-level fluctuations within the sedimentary record are critically reviewed.

Site selection criteria are established to ensure that the lake is sensitive to changes in climate and that the lacustrine sediment record is less likely to be disturbed by factors other than changes in lake-level. Site selection criteria identified from the literature are outlined and critically assessed and factors seemingly overlooked by previous researchers highlighted (Section 3.2). In light of this review the application of these criteria to the study area and the subsequent site selected are discussed in Chapter 4 (Section 4.1).

A robust methodology is needed to identify correctly changes in lake-level within the sedimentary record before the record of changing lake-levels can be subsequently interpreted as reflecting shifts in climate (Section 1.6.1). The methodological approach used in the identification of lake-level fluctuations within the sedimentary record as outlined in the literature is critically reviewed (Section 3.3).

3.2 CRITERIA FOR SITE SELECTION

The criteria for site selection can be divided into two broad categories: (a) morphological considerations and (b) geomorphological considerations. Within these categories are factors such as basin morphology and catchment-based sediment and hydrological inputs, which together with climate and lake-level fluctuations will determine the lacustrine sedimentary record. Non-climatic controls on lake-level such as the presence of an outlet and groundwater are also discussed, as site selection should ensure that fluctuations in lake-level are primarily controlled by climate (Section 1.6.1).

3.2.1 MORPHOLOGICAL CONSIDERATIONS

Morphological considerations are concerned with factors such as basin shape and size which will control the amount of wind-generated currents, wave action and non-climatic factors which will disturb the sedimentary record:

(a) Lake surface area.
(b) Depth of water body.
(c) Exposure to prevailing winds.
(d) Sediment substrate composition.
(e) Basin slope.

Wind-generated currents and wave action have the potential to erode, transport and deposit sediments within the basin. The amount of sediment mixing due to wind is related to the depth and surface area of the lake and the effective wind fetch, factors (a)-(c) (Larsen and MacDonald 1993; Verschuren 1999). Wind-generated currents transport sediment from initial areas of deposition and focus sediment in the region of the lake exposed to prevailing winds and with greatest fetch distance (Dearing 1997). Wind-driven currents causing sediment erosion and redistribution are therefore more likely to disturb the sediment record in a large, shallow lake exposed to the prevailing winds.

In smaller lakes (<100 ha surface area) currents are not normally generated by wind (Larsen and MacDonald 1993). Wave action in these small lakes is more likely to disturb sediments (Larsen and MacDonald 1993; Dearing 1997; Verschuren 1999). If the lake is sheltered from prevailing winds then sediment disturbance through wave action is predominantly restricted to the exposed margin. Sediment accumulation is then relatively undisturbed on the basin margin in the lee of the prevailing wind (Dearing 1997).

The amount of sediment mixing through wave action will vary with the strength of the wind, which will vary on a seasonal basis. Extreme events such as storms will also disturb sediments to a greater extent compared to average wind speeds (Dearing 1997). Climate shifts operating over longer timescales will also alter the wind speed, wind direction and possibly the frequency of extreme events. This temporal variability in wind speed and direction suggests that no lake is likely to be entirely free from potential sediment mixing and that within the sedimentary record wave-induced sediment mixing will occur (Larsen and Macdonald 1993). Sediment infilling of the lake over time will change the water depth, altering the susceptibility to sediment mixing by wave action (Dearing 1997).

The type of substrate will influence the amount of sediment erosion and resuspension through wave action. Particle size and cohesiveness of sediments will require different amounts of energy for erosion and transportation (Larsen and MacDonald 1993; Dearing 1997). For example, cohesive clay-rich, more compacted sediment will be less easily eroded by wave action than recent, low-density organic-rich sediment (Dearing 1997). The amount of littoral vegetation is also important, as this will afford the edge of the lake some protection from wave action and removal of sediment (Walker 1970; Dearing 1997).

Sediment disturbance on basin margins is also through slumping of sediments under gravity and is greatest on steep basin slopes. However if the sediment is fine-grained then slumping will occur on less steep slopes, between 4° and 14° with inorganic sediments in particular having a lower cohesive strength than organic sediments and will fail on slopes of around 5° (Dearing 1997; Bennett 1986). In conjunction with controls such as internal sediment cohesive strength and basin slope, trigger mechanisms such as tectonic activity will also cause slope failure and marginal sediment slumping.

Site selection based on present-day wind speeds, directions and water depth minimises control of these variables on the sedimentary record. Wave-induced sediment mixing is at a minimum within the marginal clay-rich sediments accumulating in a small lake (around 100 ha surface area) of moderate water depth (<10 metres) with some littoral vegetation and sheltered from the prevailing winds.

A basin with moderately sloping margins in a region of low tectonic activity would avoid sediment disturbance through marginal slumping. However, as discussed above this type of sediment disturbance can occur on even moderate slopes with fine-grained minerogenic sediments typical of lacustrine sedimentation (Sly 1978). Locating such a basin may be difficult. Thus sediment analysis techniques such as x-ray analysis and internally consistent radiocarbon dates can be developed to test for this type of sediment disturbance.

3.2.2 GEOMORPHOLOGICAL CONSIDERATIONS

Geomorphological criteria such as the presence of an outlet and groundwater are discussed in this section as non-climatic factors controlling lake-level. Other geomorphological criteria discussed such as catchment-based sediment and hydrological inputs operate as both climatic and non-climatic controls on lake-level and sediment type.

(a) Surface Outlet

The presence of a surface outlet is considered by some authors to be a key factor in the climatic sensitivity of a lake (Street-Perrott and Harrison 1985; Dearing and Foster 1986; Fritz 1996; Dearing 1997). These authors suggest that there is a substantial dampening of the climatic signal from lakes with a surface outlet. During times of increased precipitation any rise in lake-level is compensated for by an increase in outlet stream discharge. The impact of increased hydrological inputs into the lake is further complicated by down-cutting of the outlet stream during periods of increased discharge, altering over time the magnitude of lake-level fluctuations to periods of increased precipitation. Dearing and Foster (1986) suggest that if the outlet flows over a rock bar then a rise in lake-level could result in increased discharge but not in down-cutting of the outlet; thus the outlet stream is at a fixed level over time. Therefore the highest lake-level is the level of the outlet stream. Should the lake-level fall below the level of the outlet stream the basin becomes effectively a 'closed' basin (no outlet stream) and under these conditions may be more sensitive to climatic shifts forcing changes in the hydrological balance of the lake (Vasseljev *et al.* 1998). Lake-level fluctuations studied from several lakes in southern Sweden (Digerfeldt 1986, 1988; Harrison and Digerfeldt 1993; Yu and Harrison 1995a) and in the French Jura mountains (Magny 1992) suggest that the presence of an outlet may give muted lake-level fluctuations but that they do occur.

(b) Groundwater inputs and outputs

Precipitation infiltrating through the soil may evaporate, flow through the soil as through-flow or percolate under gravity to form groundwater (Price 1996). Within the groundwater catchment topographic highs such as mountain ranges will form recharge areas adding water to the groundwater table. Topographic lows such as river valleys will form discharge areas where groundwater is removed. Groundwater flow moves slowly through a permeable substrate, an aquifer, from areas of recharge to areas of discharge. Larger regional catchments (over 100 km in area) often incorporate a series of smaller catchments (100 m to 10 km in area) with each local catchment connected to the regional groundwater flow. The extent of regional flow within the groundwater system will depend on the catchment area topography and permeability of the aquifer (see discussion below).

Lakes lying within a groundwater catchment can receive groundwater discharge from an aquifer (discharge lakes), act as source to recharge an aquifer (recharge lakes) or both receive and release water to groundwater flow (through-flow lakes) (Brown 1995). Within the literature (Digerfeldt 1986, 1988, 1997; Magny 1992) lakes with a groundwater input are considered to behave as through-flow lakes where groundwater control on lake-level is a function of:

(1) the position of the lake within a groundwater catchment. Lake-level fluctuations within a closed lake lying on a permeable substrate close to the discharge zone will be controlled by changes in the rate of discharge (see below). Lakes lying within the groundwater recharge region will be dominated by changes to the amount of groundwater recharge (see below) (Almendinger 1990).
(2) the permeability of the aquifer on which the lake lies. Groundwater will move through most substrates with the rate of movement and the amount of groundwater

within the substrate being a function of (a) the number and size of the pore spaces, porosity, reflecting the amount of groundwater the substrate can retain and (b) how well these pores are connected (permeability), the connectivity allowing for the movement of the groundwater (Leeder 1982; Price 1996).

Many researchers may have underestimated the complexity of groundwater control on lake-level fluctuations. Magny (1992) investigating sites in the French Jura mountains suggested that sites that lay on highly permeable karstic limestone within the region may have been less responsive to shifts in precipitation due to the overriding influence of groundwater. Harrison and Digerfeldt (1993) in their review do not list groundwater control on lake-levels as a key criterion for site selection. Street-Perrott and Harrison (1985) suggest that the direct response to climate change of a lake may be obscured or dampened by long-term variations in groundwater. Street-Perrott and Harrison (1985) and Dearing and Foster (1986) indicate that groundwater control on lake level should be restricted through site selection.

The following discussion outlines the complexity of groundwater controls on lake-level and reviews the approach of some researchers to the role of groundwater and the lake-level record.

In an attempt to define fluctuations in lake-level and groundwater control Almendinger (1990) has modelled the interaction of lake-level and groundwater based on the position of the lake within the groundwater catchment. The model assumed uniform groundwater flow within an homogenous permeable substrate, a single linear discharge point and a closed basin (no surface outlet). From the model Almendinger suggests that basins lying furthest away from the discharge point are the most sensitive to changes in the regional groundwater table. Almendinger (1993) and Digerfeldt *et al.* (1992) applied this model and implied that lake-levels within such basins were controlled by regional groundwater recharge and were more climatically sensitive than lakes with both surface water and groundwater controls.

However, although Digerfeldt *et al.* (1992) and Digerfeldt (1988; 1997) imply that changes in regional groundwater are climatically driven, controlling lake-level, what they may have underestimated is the complexity of controls influencing groundwater discharge and recharge within the catchment. Models such as those developed by Almendinger (1990) are often simplified and cannot reflect the complexity of:

(1) variations in substrate permeability within the groundwater catchment. The models are generated assuming that the aquifer is homogenous whereas aquifer permeability varies widely as the type of substrate changes within the catchment. This is often the case with complex glaciated landscapes. Basal seepage within a basin lying on a highly permeable substrate may decrease over time as fine-grained sediment infilling the lake will reduce the porosity of the substrate, reducing the amount of groundwater discharge through the lake (Dearing and Foster 1986). Digerfeldt *et al.* (1992) suggests that this situation may exist in theory but is rarely found in the field. Tectonic activity can reactivate cracks and fissures in the bedrock beneath the basin, and this may dramatically increase basal seepage (groundwater output) from the lake which in turn could be misinterpreted as a climatically induced lake-level lowering (Dearing and Foster 1986).

(2) the relationship between the lake and groundwater discharge, as this often consists of more than one linear zone of discharge. Springs, seepage areas and lakes all act as zones of discharge within a catchment (Price 1996; Todd 1959). Rivers serve as groundwater discharge zones and non-climatic controls on river base level will affect the rate of groundwater discharge. Localised non-climatic controls such as tectonic activity (Gregory and Schumm 1987) and increased stream discharge due to anthropogenic activity within the catchment will cause fluvial incision and lower the base level (Howard and Macklin 1999). Conversely, episodes of river damming through, for example, peat accumulation or geomorphic activity will reduce the rate of groundwater discharge as the river base level is raised (Almendinger 1990).

(3) the function of lakes within the groundwater catchment. An increase in the amount of precipitation and groundwater recharge may induce a regional rise in the water table and a rise in lake-level but may also increase the rate of aquifer discharge from both discharge and through flow lakes (Brown 1995). Lakes will change from acting as discharge to recharge lakes (see above) reflecting the seasonal variations in groundwater level (Brown 1995).

(4) the level of regional versus local groundwater flow within the catchment. Lakes lying within a groundwater catchment are not hydrologically isolated. Changes in lake-level within one lake will have an effect on the level of other lakes. Local lake-

level changes will be superimposed on shifts in regional groundwater changes (Carter and Novitzki 1988; Almendinger 1990; Brown 1995).

The amount of regional groundwater recharge to a lake may be a response to both climatic and non-climatic factors within the groundwater catchment area. Non-climatic factors include:

(1) changes in the catchment vegetation cover. Vegetation type will modify the amount of runoff (surface and subsurface) and evapotranspiration within the catchment and thus the amount of water entering the groundwater system (Street-Perrot and Harrison 1985; Dearing and Foster 1986; Almendinger 1993; Brown 1995). Changes in vegetation cover will affect catchment hydrology but it is a complex relationship. It is often difficult to calculate how these changes will influence the levels of groundwater recharge. Vegetation cover did not remain static throughout the Holocene. Palynological analysis within the catchment will provide data on the timing, nature and the extent of vegetation dynamics (Section 3.2.2 (e)).

(2) edaphic factors. Changes in soil structure will alter the amount of water entering the groundwater body (Section 3.2.2(c)). Blanket peat development and podzolisation within the catchment could lead to potential localised and complex groundwater and surface water interactions (Dearing and Foster 1986) (Sections 3.2.2(c) and (e)).

The relationship between catchment hydrology and lake-level where basins lie on permeable substrates is enormously complex (Dearing and Foster 1986; Almendinger 1990; Digerfeldt *et al.* 1992; Brown 1995). Almendinger (1990, 1993) and Digerfeldt *et al.* (1992) warn against assuming that a lake can be considered to be hydrologically isolated, and that groundwater will always have a role in fluctuations of lake-level. To select a site where regional groundwater recharge controls lake-level and to assume groundwater recharge is climatically driven as suggested by Digerfeldt *et al.* (1992) is problematic. There are difficulties in modelling the groundwater inputs and outputs within the lake basin. The non-climatic factors discussed above may be linked to climate change but their impact on groundwater recharge means that groundwater fluctuations cannot be seen as an unambiguous precipitation proxy.

A basin lying on an impermeable substrate would reduce the role of groundwater control on lake-level. A site such as an impermeable bedrock basin, where groundwater inputs and outputs are at a minimum, should ensure a lake-level record sensitive to variations in precipitation within the lake catchment. No rock type is entirely impermeable, with faulting and fracturing allowing for some groundwater flow. However, rock types such as metamorphosed sediments are considered highly impermeable. The intense heat and pressure involved in metamorphism (a) alters mineral structure through partial melting of individual grains, reducing porosity and (b) generates predominantly plastic deformation with rocks folding rather than fracturing, further reducing permeability (Read and Watson 1978; Craig 1991). Site selection on this type of substrate will minimise groundwater flow within the basin. With groundwater outputs minimised, water outputs will depend on the presence of an outlet and the amount of evaporation over the lake surface and not the amount of groundwater discharge.

(c) Surface water inputs.

If groundwater inputs are at minimum then water input to a lake will be through surface flow and fluvial inputs (Dearing and Foster 1986). This discussion will outline surface flow inputs to a lake and potential non-climatic controlling factors.

Precipitation falling within the catchment will reach the lake through a number of flow paths (Ward and Robinson 1999):

(1) direct precipitation onto the lake surface
(2) overland flow: water that flows over the ground surface. Overland flow will only occur when the water table is close to the ground surface and the soil infiltration capacity is reduced.
(3) through-flow: water which infiltrates the soil surface and moves laterally through the upper soil horizons. Through-flow will respond to high intensity rainfall events by increasing flow rates. Through-flow is also the mechanism by which water stored within the soil profile contributes to the baseflow of streams.
(4) stream flow: water within streams will have followed the above three flow paths.

Which mechanism will predominate in a particular catchment is dependent on climate, soil condition and lithology, with vegetation cover and topography important secondary controls at the hillslope scale (Burt 1992). Overland flow is rarely observed within temperate, upland catchments as vegetation cover and steep slopes facilitate high soil infiltration values (Ward and Robinson 1999). Within such catchments through-flow is considered the most important mechanism by which precipitation will reach streams and lakes. Within a steep catchment the critical factor influencing the amount of through-flow is the soil infiltration capacity. The soil infiltration capacity can be altered over time by:

(1) changes in the soil structure. Through-flow is greatest in the surface layers of the soil. Thin soils overlying impermeable bed rock and soils that are markedly stratified will have high amounts of through-flow (Knighton 1998). For example, soils in which an iron pan develops a short distance from the surface will facilitate greater through-flow of precipitation (Ward and Robinson 1999). Iron pans will develop within soil profiles in response to podzolisation with the rapid leaching of base minerals due to changes in environmental factors such as increased precipitation (Pennington *et al.* 1972; Davidson and Carter 1997). Anthropogenic clearance of the vegetation cover is also considered as a forcing mechanism in the acceleration of podzolisation and the development of iron pans (Moore 1993).

(2) changes in vegetation cover. Through-flow beneath vegetation is high as the roots and plant litter within the soil allow for greater infiltration rates. Vegetation changes, in particular deforestation, will alter the soil structure as suggested above. Deforestation may intensify stream flow by adversely affecting soil structure and volume, and rainfall interception reducing the infiltration rate and increasing the amount overland flow and reducing the amount of soil water storage (Calder 1986, 1990). The interaction between soil structure and vegetation is complex with the impact of vegetation change on the hydrology often temporary, in particular where vegetation change is as a result of anthropogenic clearance (Dearing and Foster 1986). New vegetation cover would reach an equilibrium allowing for precipitation to infiltrate the ground surface and through-flow to establish.

(3) the build up of blanket peat. Peat build up within a catchment is a complex interaction between vegetation, soil development, climate and anthropogenic impact (Moore 1993). Once peat cover is established the structure of the peat changes, to consist of a surface aerobic acrotelm layer with high hydraulic conductivity and a much thicker anaerobic saturated catotelm layer beneath (Ingram 1978; Clymo 1991). Peat rapidly dissipates excess water through the acrotelm ensuring that the surface does not become saturated with water, as this would restrict growth and

accumulation of the peat itself (Ingram 1987; Wheeler 1999). Thus far from acting as a sponge and impeding through-flow the living peat surface will provide a layer through which rainwater is actively transported (Baird *et al.* 1997; Burt *et al.* 1997, 1998). Therefore for this reason it is argued here that the initial vegetation change from forest to peat cover (Section 3.2.2 (e)) may have caused a change in catchment hydrology but once established the peat would have continued to act as an effective medium for through flow.

(4) geomorphic activity within the catchment. Slope processes and mass movements such as landslides will affect the soil thickness and soil composition, affecting in turn the soil structure and thus infiltration capacity. The impact of geomorphic events may only be short-lived before soil development processes and through-flow once again reach equilibrium.

Through-flow is the main contributor to stream flow and as outlined by Dearing and Foster (1993), stream flow may be the dominant hydrological input for some lake catchments. Thus many of the above non-climatic factors will affect the amount of stream flow. Geomorphic activity within the catchment may also have a direct impact on stream flow with debris flows and landslips damming or altering the stream course and diverting water away from the lake.

Digerfeldt (1986, 1988), Harrison and Digerfeldt (1993) and Magny (1992) discuss the implications of hydrological outputs from the lake system such as the presence of an outlet stream but do not consider hydrological inputs. The volume of water reaching the lake by the mechanisms discussed above will, of course, have climatic controls. Controlling factors such as soil structure and vegetation change will have climatic controls but will also operate as non-climatic controls on the amount of surface water reaching the lake. It is suggested here that these authors may have underestimated the control of non-climatic factors, in particular vegetation change, on the amount of surface flow entering a lake. Site selection should ensure that within the lake catchment all hydrological inputs are identified and that the role of non-climatic factors such as vegetation cover and geomorphic processes are fully defined.

(d) Sediment inputs.

Variations in the type and composition of lacustrine sediment may be a result of (a) endogenic changes in lake productivity such as diatom and algal population fluctuations (Fritz 1989); (b) changes in allogenic sediment inputs occurring within the catchment; and (c) lake-level changes. Interactions between changes in lake productivity and allogenic sediment input may produce highly variable sediment compositions which may be difficult to interpret. If the sedimentary record within the lake is to be interpreted as a record of changes in lake-level, it is critical to be aware of and attempt to control for all sediment inputs into the lake system.

Direct allogenic sediment inputs from within the catchment may be climatically forced but there are non-climatic factors which will control sediment input into the lake. These factors include:

(1) sediment source: the amount and type of sediment available for transfer to the lake is dependent on the geology and the history of sediment deposition within the catchment. Catchments with a history of glacial deposition, for example, will have a large store of sediment available for reworking but the amounts of this sediment type available for reworking may change over time (Brazier *et al.* 1988).

(2) sediment availability: the erosion potential within the catchment is primarily dependent on the amount of bare ground and relief. In large catchments relief and slope gradient are important controls on the amount of erosion with high relief associated with more active slope processes generating sediment (Phillips 1990; Milliman and Syvitski 1992). At a regional and upland catchment scale, factors such as vegetation cover (Section 3.2.2 (e)) and land use are of greater significance in controlling the amount of sediment erosion (Knighton 1998). Recent changes in vegetation cover associated with anthropogenic activity such as increased grazing intensity are known to have increased lacustrine sediment yields (Foster and Walling 1994).

(3) sediment transport: sediment eroded from the catchment is principally transported to the lake *via* stream channels. Once within the stream, sediment may not enter the lake directly but sediment can be stored temporarily or permanently as alluvium within the stream channel, alluvial fans and terraces (Phillips 1991). Any change to

the fluvial system through a range of geomorphological and environmental factors, such as channel switching or discharge increases will alter the amount of sediment stored or transported (Richards 1993; Walling 1983).

(4) tectonic activity. There is evidence for increased seismic triggering of geomorphic activity such as landslips and rockfalls during the early Holocene in response to isostatic rebound (Ballantyne 1997, Section 2.2). These events can potentially generate large amounts of sediment which could input directly into the lake or enter *via* stream input.

Dearing and Foster (1993) indicate that the significance of river inflows on sedimentation patterns is related to the catchment area:lake area ratio. A larger catchment:lake area ratio (>10) can support a channel network and so sediment input is dominated by fluvial processes. With a catchment:lake area of <10 the sediment yield within the catchment is dominated by slope processes. Within this type of catchment the sediment yield within the lake is high as there is no fluvial sediment storage network.

To control for allogenic sediment inputs, site selection would include the following criteria:

(1) a site with a low catchment:lake area ratio. However such a site with a large lake surface area would be susceptible to other sediment control factors such as sediment disturbance through wind-generated currents (Section 3.2.2 (a)).

(2) a basin lying in a catchment dominated by glacial erosion as opposed to deposition, reducing the amount of sediment available for erosion through slope processes and subsequent deposition within the lake.

(3) a basin in a region of low tectonic activity. This will reduce the potential for seismically triggered geomorphic events.

(4) a basin in a catchment with low relief. This will reduce the amount of sediment generated through geomorphic activity such as slope processes.

Within a temperate upland region such as northern Scotland it would be difficult to fulfil all the site selection criteria outlined above to control for sediment input. Site selection should therefore ensure that within the catchment all potential sources of sediment input and the processes generating sediment input can be identified and the nature and type of the sediment input defined.

(e) Vegetation Cover.

The interaction between vegetation cover and lake catchment processes is complex, and is a key non-climatic control on catchment hydrology and sediment inputs (Sections 3.2.2 (b), (c) and (d)).

Locating a catchment within northern Scotland where during the Holocene vegetation change was minimal, would be impossible, however desirable. It is critical, therefore, that some account of Holocene vegetation change is made if changes in lake-level resulting from climate shifts are to be differentiated from those resulting from vegetation control on the basin hydrology.

High-resolution palynological analyses within the lake catchment can define the vegetation dynamics and indicate periods of anthropogenic impact on the vegetation cover. Pollen analysis from a single site will only have limited applications at the catchment scale, perhaps providing only a generalised picture of vegetation change. However, this type of analysis offers the best approach to ascertaining the control of vegetation cover in catchment hydrology. Palaeoecological data have been applied to the interpretation of lake-level fluctuations by previous researchers (Digerfeldt 1988; Digerfeldt *et al.* 1992; Magny 1992). However, although the data are primarily used to establish a chronology and to ascertain human activity and associated soil erosion, regional changes in vegetation are not considered as forcing mechanisms for changes in lake-level even within regions where groundwater is the controlling factor on lake-level (Digerfeldt *et al.* 1992). Although Magny (1992) does suggest that deforestation during the Middle Ages may have reinforced the impact of a lake-level rise, he associates this with the 'Little Ice Age' in the Jura Mountains.

A high-resolution record of vegetation dynamics when compared to the climatic record generated will further define the nature, timing and extent of vegetation change and provide additional data to describe fully the complex relationship between climate change, vegetation cover and lake-level fluctuations.

3.2.3 PRINCIPAL SITE SELECTION CRITERIA

The discussion above outlines how site selection criteria simplify the lacustrine hydrological and sedimentological system, and limit the number of non-climatic variables. The lake-sediment record can subsequently be interpreted more reasonably as a series of climatically controlled fluctuations in lake-level.

Digerfeldt (1986, 1988) and Harrison and Digerfeldt (1993) have outlined site selection criteria which they suggest would ensure an undisturbed sedimentary record and climate sensitivity of the lake. These criteria include:

(1) small lakes <100ha
(2) basin margins that are gently sloping
(3) shelter from prevailing winds
(4) absence of an outlet stream
(5) the basin lies in a region of low tectonic activity.

However the work of Almendinger (1990), Brown (1995), Dearing and Foster (1986), Dearing (1997), Larsen and MacDonald (1993) and Verschuren (1999) highlighted other criteria that need to be considered, especially if the sediment record within the lake is to remain undisturbed by factors other than lake-level fluctuations (Section 3.2.1).

In light of the literature review further site selection criteria were applied during this research to define:

(1) The role of groundwater control on lake-level. The discussion in Section 3.2.2 (b) has highlighted the complex relationship between groundwater and lake-level. Basins lying on impermeable substrates minimise groundwater control on lake-level.
(2) The role of catchment sediment inputs on lacustrine sediment type. If the lacustrine sedimentary record is to be interpreted as a series of lake-level fluctuations it is critical to identify all potential sediment sources within the catchment.
(3) The role of non-climatic variables operating within the catchment influencing a range of sedimentary and hydrological controls such as changing vegetation cover

(Section 3.2.2.(e)). Palynological analysis as discussed in Section 3.2.2 (e) provides data on vegetation changes within the catchment and, in the absence of archaeological data, information on human activity.

3.3 INTERPRETATION OF THE LACUSTRINE SEDIMENTARY RECORD

Periods of higher lake-level may be indicated by geomorphological evidence such as old relict shorelines above the present-day lake-level (Brown 1995; Digerfeldt 1986). However, much of the evidence is stratigraphically recorded, with fluctuations in lake-level affecting a variety of limnological, sedimentological and biological processes. Site selection criteria, as discussed in Section 3.2, control for non-climatic controls on lake hydrology and sedimentation and thus ensure that the lacustrine record can be interpreted as a series of changes in lake-level. An equally robust methodological approach is required to allow the correct interpretation of the sedimentary record.

A pioneering methodology for identifying lake-level fluctuations within the lacustrine sedimentary record has been developed by Digerfeldt (1986, 1988). Key aspects of Digerfeldt's methodology are:

(1) the combination of different lines of sedimentological evidence. Discrete lines of evidence may have alternative explanations that are not as a result of fluctuations in lake-level, but it is the combination of lines of evidence that strengthens the interpretation. Other researchers (Magny 1992; Harrison and Digerfeldt 1993) have used this approach to define lake-level fluctuations.

(2) the record of lake-level fluctuations is contained within the near-shore sediments, but these sediments are the most prone to reworking and removal of the record (Section 3.2.1). Digerfeldt (1986) suggests that with (a) a coring strategy of a transect of cores taken from the basin edge out across to the basin centre and (b) chronological control, based on radiocarbon age determinations and pollen stratigraphy within the sedimentary record from each core, any sediment changes (lake-level fluctuations) can be synchronously traced out across the basin.

Previous approaches to the interpretation of the sedimentary record are here identified and critically reviewed.

3.3.1 DISTRIBUTION OF LAKE VEGETATION

The distribution of lacustrine aquatic plants is in part determined by water transparency, the degree of exposure and type of substrate; however, the most important factor is water depth (Hannon and Gaillard 1997; Spence 1964; Wheeler 1999). Most aquatic plant species can be classified into emergent, floating-leaved and submerged with potential water depth ranges identified for each taxon (Spence 1964; Harrison and Digerfeldt 1993).

A lake-level lowering should result in the displacement of shallow near-shore aquatic species out towards the centre of the basin and a rise in level will result in their shoreward displacement (Digerfeldt 1986; Gaillard and Digerfeldt 1991). Palaeoecological techniques such as macrofossil and pollen analysis are used to identify changes in aquatic plants at any one depth over time. Digerfeldt (1986) proposes that this type of analysis can quantify changes in water depth with species with a well-defined water depth range the most useful in identifying periods of lake-level fluctuations. However, in interpreting this type of palaeoecological data factors other than variations in water depth need to be considered:

(1) changes in lake productivity: aquatic species may change over time in response to variations in the nutrient status of the lake and changes in the nature of the substrate not governed by fluctuations in the lake-level (Spence 1964; Walker 1970; Birks 1973).

(2) changes in shoreline exposure: aquatic vegetation will be well established if the shoreline is sheltered; changes in the strength and prevailing wind direction will affect the degree of shoreline exposure (Dearing 1997).

(3) taphonomic factors: variations in the aquatic type and/or macrofossil numbers can be an artefact of changes in the conditions of organic matter preservation.

(4) sediment reworking: aquatic macrofossils and pollen are considered to be essentially local in origin; that is, deposited close to the parent plant (Davis *et al.* 1971; Harrison and Digerfeldt 1993). The presence of aquatic vegetation suggests that reworking of the sediment is at a minimum with the plants themselves effectively reducing the amount of sediment turbulence (Walker 1970). However, there is strong evidence for reworking of pollen within lake sediments through resuspension

by wave action and wind-generated currents (Davis 1973; Davis and Brubaker 1973; Peck 1973). Birks (1973, 1980) suggests that aquatic macrofossil distribution is a complex interaction between limnological processes (wind-generated currents, wave action, thermal mixing) and specific ecological adaptations for seed and fruit dispersal by aquatic species.

(5) natural sediment infilling: any record of lake-level change generated from aquatic plant remains must be interpreted against a background of natural sediment infilling and terrestrialization. Aquatic species associated with hydroseral succession will colonise the lake substrate as the water depth becomes shallower due to sediment infilling and not in response to a climatically forced lake-level change (Walker 1970; Hannon and Gaillard 1997) (Section 4.6).

There is often difficulty in locating a site which fulfils site selection criteria (Section 3.2) and contains a sedimentary record rich in aquatic remains suitable for the type of analysis suggested above. Digerfeldt (1988) noted that for many of his sites in southern Sweden, including Lake Bysjon, there was insufficient aquatic macrofossil material to generate lake-level fluctuation data.

There are only a few aquatic taxa that provide absolute quantitative data on water depth. Many taxa thrive in a range of water depths (Spence 1964; Harrison and Digerfeldt 1993). Thus within the macrofossil analysis the interpretation of quantitative water depth changes is rarely achieved and lake-levels are often described qualitatively as becoming higher or lower (Gaillard and Digerfeldt 1991).

Magny (1992) uses the association of structurally different carbonate concretions and aquatic macrophtyes to determine changes in lake-level. The carbonate concretions show a range of differing structures with each structure dependent on a specific aquatic plant species, environment of deposition and a certain water depth. This technique is shown to be successful in determining lake-level fluctuations but is limited to specific site types as it relies on the biochemical relationship between an environment suitable for the deposition of lacustrine chalk and aquatic vegetation.

Despite the limitations discussed above Dearing (1997) suggests that the advance and retreat of littoral vegetation as inferred from macrofossil analysis represents one of the strongest lines of sedimentary evidence for lake-level change.

3.3.2 SEDIMENT COMPOSITION

A change in sediment composition is a frequently interpreted variable in the determination of lake-level fluctuations. Site selection criteria should ensure that non-climatic controls on sediment composition are limited (Section 3.2). The methodology applied to the interpretation of changes in sediment composition must take into account potential non-climatic controls on sediment composition. The methodological approach should fully define the environment of sediment deposition using both sediment composition and the stratigraphic relationships of the sedimentary sequence.

Within lacustrine depositional environments such as carbonate lakes the presence of marl (a high calcium carbonate sediment type) is interpreted as an indicator of increased water depth. Magny (1992) and Mannion (1982) demonstrate how at such carbonate lake sites the stratigraphic relationship between sequences of marls, gyttjas and peats can also be interpreted as changes in water depth. Although these sediment types can potentially be interpreted as changes in lake-level, the sediment stratigraphic interpretations are limited to lacustrine systems dominated by carbonate precipitation.

Digerfeldt (1986) suggests that lacustrine littoral sediments are normally characterised by a larger quantity of reworked 'coarse' minerogenic matter and 'coarse' organic matter originating from macrophytic vegetation along the shore. A past lowering of lake-level would be associated with an outward displacement of the shore and an increase in the amount of 'coarse' matter within the sediment resulting from the associated increase in littoral erosion. Digerfeldt (1986) outlines a methodology to define this 'coarse' material. The material retained after washing a sediment sample through 0.5 mm and 0.2 mm sieves is dried in an oven at 105°C and weighed. The amount of 'coarse' minerogenic and 'coarse' organic material is then determined through loss-on-ignition of this material at 550°C for four hours. The weight in grams of each 'coarse' component is quoted.

Digerfeldt *et al.* (1992) determine the coarseness of minerogenic material only with sediment grain size classed as 0.4-2 mm and > 2 mm. Periods of increased amounts of coarse grain size are used in the interpretation of lake-level lowering. Although outlined as a key part of the methodological approach (Digerfeldt 1986), the amount of coarse material within the lake sediment as an analysis technique is inconsistently applied from site to site (*cf.* Digerfeldt 1988; Harrison and Digerfeldt 1993). Several criticisms can be made of the approach. There are controls on the amount of coarse organic material present within the sediment which are unrelated to changes in lake-level and littoral reworking:

(1) reworking of the littoral edge of a lake can be a function of the amount of shoreline exposure to wave action and wind-induced water currents (Dearing 1997). An increase in storminess could produce greater amounts of coarse organic material within the sediment record without changing the lake-level.

(2) resistance to physical breakdown: variability in the resistance of certain plant species to the processes of physical breakdown (Grosse-Brauckmann 1986).

(3) the amount of organic material produced at the shore margin reflects the productivity of the lake shore with productivity controlled by factors such as nutrient status, acidity and substrate (Spence 1964; Walker 1970).

(4) an increase in the amount of coarse organic material within the sediment may reflect changes in the marginal vegetation associated with fen peat build up. This build up of fen peat may be as a result of natural infilling and hydroseral succession within the lake but once established may continue to be a source of coarse organic material into the lake.

Likewise changes in the minerogenic content of lacustrine sediment will have a series of sediment controls not driven by changes in lake level (Dearing and Foster 1986; Dearing 1997):

(1) the character of the shore deposits will determine the amount and type of sediment available for reworking. Basins lying in bedrock basins will have little or no minerogenic littoral sediments available for reworking. Thus within such lake systems the marginal sedimentary record is complacent to littoral reworking events.

(2) High-energy geomorphic and fluvial events will input coarse minerogenic sediments from the catchment into the lake. A range of non-climatic and climatic factors

(Section 3.2.2) will control the amount and type of sediment input to the lake *via* geomorphic and fluvial inputs.

Compared to marine or fluvial systems the amount of energy available for sediment transport and erosion within lakes, in particular small lakes is limited, with the most active current winnowing of finer sediments taking place in shallower shoreline areas (Jones and Bowser 1978; Sly 1978). The clastic components of most lake sediments are dominated by silt and or fine sand size materials (Jones and Browser 1978). Detailed grain size analysis of minerogenic sediments, using techniques such as laser diffraction, within the lacustrine record offers further definition of sediment source, mode and energy of deposition, through the examination of:

(1) the degree of particle size grading before, during and after the deposition of minerogenic sediment pulses across the basin can provide data on changing energy of deposition (Dearing 1997).
(2) the degree of sediment sorting can provide data on the depositional mechanism: sediments that are deposited rapidly are generally poorly sorted and sediments that have been reworked by wind or water are well sorted (Tucker 1991). Generally sorting improves with more prolonged sediment transport. However, the degree of sediment sorting also reflects (a) the sediment source; that is what type of material is being eroded and reworked; (b) the grain size itself, coarser (gravels) and fine sediments (silts and clays) are generally more poorly sorted than sand-sized sediments which are more easily transported and therefore sorted by wind or water.
(3) Spatial trends within synchronous units, that is across the basin, within the grain size data set can provide an indication of sediment transport direction.

Campbell (1998) proposed that changes in the coarseness of lacustrine sediments are a response to fluctuations of inlet and outlet stream discharge. Periods of high precipitation and high stream discharge are indicated by an increase in the median sediment grain size as all the finer grained sediments will have been transported out of the lake system. Criticisms of this methodological approach are (a) the assumption of a direct climatic control on the fluvial system and the amount of discharge (Section 3.2.2) and (b) as with all lacustrine sediment interpretations, the complex interactions between seasonal limnological processes, sediment grain size transport and redeposition and fluvial currents are difficult to simplify into a direct climatic interpretation.

Dearing (1997) suggests that clearly defined changes in sediment composition associated with different hydrological conditions offers one of the strongest lines of sedimentary evidence for identification of lake-level fluctuations. The application of Digerfeldt's methodology remains problematic and is highly dependent on the sediment type available for reworking and is susceptible to the misinterpretation of allogenic sediment inputs. Changes in sediment composition allied to approaches that further define sediment type and source represent valuable data for the interpretation of the sedimentary record (Section 3.4).

3.3.3 THE SEDIMENT LIMIT

The application of variations in the sediment limit within the lacustrine sedimentary record as a method to identify lake-level fluctuations is proposed by Digerfeldt (1986, 1988) and has been applied by other workers such as Harrison and Digerfeldt (1993). This aspect of Digerfeldt's methodology has been criticised by Dearing (1997) and is similarly critically reviewed below.

Within a lake the sediment limit is defined as the highest limit for the deposition of organic sediment (Digerfeldt 1986, 1988). Above the sediment limit 'fine' organic material cannot permanently accumulate as the sediment is constantly redistributed through wave action and wind-generated currents. Digerfeldt (1986, 1988) proposes that the sediment limit can be identified within present-day lakes and this water depth used to quantify past changes in lake-level. Within the sedimentary record Digerfeldt (1988) defines the sediment accumulated at the sediment limit as having an organic content of 50% determined on a dry weight basis. The organic matter content is determined using loss-on-ignition at 550°C. A rapid increase in organic content to greater than 50% dry weight indicates that the sediment limit has risen with increasing water depth. A rapid decrease in organic content indicates a fall in the sediment limit and a fall in lake-level.

Digerfeldt (1988) illustrates how changes in the sediment limit recorded within the sedimentary record of a transect of cores traced synchronously across a basin can be interpreted as series of fluctuations in lake-level. The magnitudes of the fluctuations in lake-level as proposed by Digerfeldt (1988) are defined using the present-day sediment limit within the lake, and the depth of the sediment limit recorded within the near-shore stratigraphy. Being able to quantify changes in lake-level is one of the main strengths of this approach.

There are, however, several criticisms against the interpretation of changes in the sediment limit as a record of lake-level fluctuations:

(1) the identification of the sediment limit of organic sedimentation depends on the sediment record remaining undisturbed by factors other than fluctuations in lake-level. Dearing (1997) suggests that sediment controls such as wind strength and frequency, wind exposure, water depth and underwater slopes, sediment type and littoral vegetation (Section 3.2.1) vary from site to site and change over time and make an interpretation of the sediment limit very difficult. In particular extreme events such as storms will cause disturbance of the sediment at a greater water depth than that associated with average conditions.

(2) the most sensitive cores for the detection of changes in the sediment limit are marginal cores but these are most prone to sediment reworking. Digerfeldt (1988) attempts to identify these hiatuses by constructing a pollen-stratigraphic record generated from both deeper water and marginal sediments. However, the identification of reworking and a hiatus is possibly not as readily defined from a pollen stratigraphic record as Digerfeldt implies.

(3) the sediment limit is defined from the organic content of the sediment. The organic content of the sediment can vary through changes in biomass productivity which may have non-climatic controls such as nutrient content and acidity. Digerfeldt developed this methodology using lake sites with highly minerogenic sediment input: thus within these lakes the sediment limit may well be a clearly defined limit between predominantly organic and minerogenic sedimentation. However, lakes with high levels of organic productivity will have a less well defined sediment limit.

(4) the definition of the sediment limit is an interdependent variable. The sediment limit is defined from the loss-on-ignition value, and clearly any changes in allogenic mineral content unrelated to fluctuations in lake-level will control the percentage loss-on-ignition value (Section 3.3.2).

(5) within Digerfeldt's methodology the determination of the magnitude of lake-level fluctuations depends on the assumption that controls on sediment limit at present

day are similar to those throughout the Holocene. Dearing (1997), as outlined in point (i), indicates that the controls on sediment limit may have changed over time.

This methodology seems to have limited applicability to all temperate lake sites. Interpretation of changes in sediment limit as reflecting fluctuations in lake-level is, however, deeply flawed owing to sedimentological controls on sediment composition.

3.4 METHODOLOGICAL APPROACH TO THE INTERPRETATION OF THE LACUSTRINE RECORD

Within the approaches reviewed above one of the major limitations was the inability to differentiate within the lacustrine stratigraphy sediments associated with fluctuating lake-levels and sediments associated with catchment-based sediment input events. Differentiating between these sediment types is a critical part of the methodological approach used in this research. This methodological approach was developed based on a clear understanding of:

(1) how a lake system responds sedimentologically to changes in the hydrological balance.
(2) basinal catchment sediment inputs and potential climatic and non-climatic controls (Section 3.2.2).

This methodology is based on the identification of sediment types associated with particular depositional processes that are a response to fluctuating lake-level or catchment sedimentary input events. Clearly this requires a comprehensive and precise record of lacustrine sediment stratigraphy based on:

(1) complete sediment description involving sediment analysis techniques such as loss-on-ignition and x-ray analysis (Section 5.2). Data generated on sediment composition, stratigraphic relationships and sedimentary structures can be interpreted as indicating potential mode and environment of deposition (Section 3.3.2).
(2) detailed grain size analysis using laser diffraction (Section 5.3). Variations in grain size data can further define the energy and mode of deposition and could differentiate between allogenic and endogenic sediment inputs (Section 3.3.2).

(3) a record of sediment stratigraphy from marginal and deeper water sediments. Fluctuations in lake level may result in different sediment types deposited across a basin thus the mode of deposition for these sediments must be incorporated into any interpretation of the stratigraphic record.
(4) The establishment of a precise chronological framework (Section 5.6). Secure chronological control of sedimentary events enables correlation of events recorded in deeper water and marginal sediments (Section 3.3.2).

Alongside the interpretation of changes in sediment composition a record of changes in aquatic vegetation (Section 3.3.1) over time across the basin forms part of the methodological approach. Within many lacustrine systems the establishment of marginal aquatic plants is recorded as part of a fen or marsh system, with the presence of the associated sedge peat within the lacustrine sedimentary record considered a response to a lake-level lowering (Korhola 1992; Digerfeldt 1997). However, as discussed in Section 3.2.2 the establishment of a fen or marsh may be in response to natural sediment infilling and hydroseral succession. Walker (1970) and more recently Klinger (1996) have challenged the hydroseral development model with the indication that many fen vegetation systems will persist for thousands of years without progressing to a 'climax' vegetation, ombrotrophic mire or woodland. Fens are considered sensitive to changes in their normal patterns of water storage and movement (hydrology) with only slight variations causing the biota to respond with large changes in species richness and ecosystem productivity (Mitsch and Gosselink 1986). Defining and identifying the response of an established and persistent fen to fluctuations in lake-level was considered a methodological approach worth developing and is discussed in the following section.

3.4.1 FEN PEAT STRATIGRAPHY

Fen and marsh vegetation types associated with lacustrine margins are described as permanent wetlands (Spence 1964; Wheeler 1999). In such wetlands deeper water marsh vegetation is inundated for most of the year and marginal fen type vegetation is inundated for only part of the year. Thus within a fen and to a limited extent within a marsh the water table remains close to the surface for much of the year, providing an environment where detrital plant material will accumulate as peat (Walker 1970; Wheeler 1999).

The behaviour of the water table and water movement within such wetlands and peat systems is complex (Siegel 1988). Within the fen peat system there is an average water table around which fluctuations will occur varying in magnitude, frequency and timing (Wheeler 1999). In fens dominated by groundwater, changes in discharge and recharge rates both regionally and locally (Section 3.2.2 (b)) will alter the water level within the peat (Gosselink and Turner 1978; Carter and Novitzki 1988; Siegel 1988). In a lake system with reduced groundwater input (a basin lying on impermeable bedrock) the water table within the fen is effectively controlled by the lake-level. Many fen peat systems are known to have a periodically unsaturated surface layer but the hydrological importance of this layer is not well known (Wheeler 1999). Ingram (1992) suggests that only fens dominated by *Sphagnum* could have a capacity for hydroregulation comparable to that of ombrotrophic bogs, possessing internal structures such as an acrotelm with a high hydraulic conductivity overlying a catotelm, a deeper layer with a very much lower hydraulic conductivity (Section 8.1.2). The hydraulic conductivity of a peat can be predicted from the bulk density or fibre content, with the conductivity of the peat decreasing with decreasing fibre content associated with increased amounts of decomposition (Mitsch and Gosselink 1986). The type of plant material that makes up the peat also affects the hydraulic conductivity. Peats composed of predominantly grasses and sedges are more permeable than peat containing predominantly mosses (Ingram 1983). Therefore within a fen peat the unsaturated surface layer, similar to the 'acrotelm', composed of poorly decayed fibrous sedges and grasses could provide the main water flow path within the system (Wheeler 1999) with a potentially lower rate of water movement at depth within the 'catotelm'. However, McNamara *et al.* (1992) indicate that within a fen there is little or no change in the hydraulic conductivity value with increasing depth.

The above discussion highlights that within the fen peat system, structural differences, an acrotelm and catotelm and peat composition could facilitate water movement in response to changing water level (lake-level). Fen peat accumulating on the margin of a lake is generally wedge-shaped. The surface of the water table will therefore slope in relation to the peat surface across the fen. Peat accumulating close to the lake edge will have a water table that is potentially permanently near the surface, only falling below the surface during low lake-levels. Further away from the lake margin the water table is at a greater depth below the fen surface, only rising close to the surface during high lake-levels. The water level in peat accumulating at a distance from the lake edge is least influenced by changes in lake-level and may be part controlled by the terrestrial water table.

Walker (1970) and Wheeler (1999) suggest that if the lake-level (water table) was close to the accumulating surface of the peat then fluctuations in lake-level may alter the amount of surface peat decay and the degree of humification. The hydrological controls on the rates of surface peat decomposition are complex and it cannot be assumed that the increase in frequency or duration of a higher water level will alter decomposition rates (Mitsch and Gosselink 1986). However, Walker (1970) suggests that with rising lake-levels there is little time for extensive humification and peat will accumulate rapidly. When lake-level falls the surface of the peat will dry out and extensive humification will take place with the rapid oxidative breakdown of the plant material (Gosselink and Turner 1978).

Determination of the amount of peat decay (humification) as an indicator of surface wetness at the time of deposition has been applied to the peat record in raised mires (Aaby and Tauber 1975; Aaby 1976; Barber *et al.* 1994; Mauquoy and Barber 1999) and ombrotrophic blanket mire systems (Blackford 1990; Blackford and Chambers 1991, 1995; Chambers *et al.* 1997) (Section 1.6.2). It is proposed that fen peat systems share similar critical attributes with these mire systems, suggesting that a humification record generated from a lacustrine fen peat would also provide a record of shifts in mire surface wetness, here controlled by lake-level. This is based on:

(1) the implied acrotelm/catotelm structure of fen peats, discussed above, which suggests that there are sites of differential decay rates within the fen peat structure, with plant material within the acrotelm decaying at a greater rate than within the catotelm.
(2) the fibrous fen peat composition (high hydraulic conductivity) alongside an acrotelm/catotelm structure facilitates the movement of the water table controlling the amount of time plant material spends in the aerobic acrotelm, and therefore the degree of humification.

(3) fluctuations in water level occur across fen mire systems, indicating that prolonged periods of increased or decreased water levels can control the amount of surface peat decay (degree of humification) (*cf.* Blackford 1993).

Clymo (1991) implies that the amount of organic matter decay (humification) in peat systems is an interaction between plant productivity, hydrology and the processes of decay. Within the fen mire system fluctuations in lake-level act as an hydrological control on peat decay. Further controls on the degree of humification include:

(1) vegetation type: within the fen system water depth is thought to be the primary control on vegetation type. However, the response of vegetation type to fluctuations in water level is complex (Section 3.3.1). Differential rates of decay of vegetation types will exert a control on the degree of peat humification which may or may not be in response to a change in water level.

(2) terrestrialization and hydroseral development: any changes in the degree of peat humification and inferred fluctuations in lake-level need to be interpreted against a background of shallowing water depth and lateral fen growth through sediment infilling.

(3) decay at depth: within ombrotrophic mire peat records a long-term increase in decay (humification) with depth (*cf.* Anderson 1996, 1998; Blackford and Chambers 1995) is attributed to the continued decay, at a reduced rate, of peat within the catotelm (Clymo 1984, 1992). The proposed presence of a catotelm within a lacustrine fen peat also suggests the likelihood of continued decay at depth within these mire systems.

(4) rate of lake-level fluctuations: the rapidity of a rise or fall in lake level will control the vegetation type and the rate of peat accumulation and subsequently as outlined in point (1) the degree of peat humification.

(5) exposure of peat surfaces during very low lake-levels could result in removal by erosion of humified peat from the record (Walker 1970), under-representing prolonged periods of highly humified peat (low lake-levels) within the humification record.

It is proposed that humification changes within a lacustrine peat stratigraphy could provide a record of fluctuations in lake-level. The potential range of sensitivities to changes in lake-level across the fen surface (discussed above) could also provide a mechanism for determining the magnitude of fluctuations in lake-level. This is the first time, to this authors, knowledge, that this methodology has been used to define fluctuations in lake-level. The determination of the degree of humification for fen peats follows the methodology outlined by Blackford and Chambers (1993). This methodology and the data analysis are discussed in Section 5.4.

PART 2 LAKE-LEVEL FLUCTUATIONS

CHAPTER 4
SITE SELECTION AND SITE DESCRIPTION

4.1 SITE SELECTION

Glen Affric is situated within the climatically sensitive region of north west Scotland (Section 1.3). Therefore the application of site selection criteria outlined in Chapter 3 should ensure that a lake within the glen selected to investigate lake-level fluctuations as a climate proxy is:

(1) sensitive to climatically induced changes in the hydrological balance of the lake catchment and

(2) contains a lacustrine sedimentary record which can be interpreted as a record of fluctuations in lake-level.

The site selection criteria ensure that the lake selected has relatively simple hydrological and sedimentological systems and a limited number of non-climatic controls. This ensures that the response of this system to climate change can be interpreted from the lacustrine sedimentary record.

An initial desk-top survey using topographic maps at scales 1:50 000, 1:25 000 and 1:10 000 and aerial photographs of Glen Affric was used to identify potential sites. These sites were selected using the criteria described in Chapter 3:

(1) Closed basins: basins with no water output through stream flow or groundwater. The geology of the region (Section 2.2) consists of impermeable meta-sediments which ensure that groundwater flow is at a minimum (Section 3.2.2 (b)). However, no basin should be considered hydrologically isolated from groundwater flow. Basins with no outlet stream were confined to the high-altitude corries. Within these basins the sedimentation rate was considered to be too low to provide a high-resolution sedimentary record (*cf.* Rapson 1985). Potential lake sites along the

valley floor all had outlet streams. Field survey work was required to assess the nature of the outlet stream.

(2) Water and sediment inputs: topographic maps showed the stream inputs to potential sites and air photographs were used to outline other potential sediment inputs from geomorphic activity within the catchment such as alluvial fans (Section 3.2.2 (c, d)). Field survey work was required to define these inputs. Sites with several large input streams, or where a lake was connected to other lakes upstream were rejected. These complex hydrological systems were considered as climatically insensitive with a greater potential for non-climatic controls on lake-level and ambiguity in sedimentary interpretations.

(3) Basin size: lake area could easily be calculated from the topographic maps. Basins selected needed to be small in size <100 ha (Section 3.2.1).

(4) Basin sheltered from the prevailing winds: the prevailing wind direction within Glen Affric is westerly to south westerly, but is complicated by topography (Section 2.5.4). Lakes sheltered from prevailing winds will reduce the amount of wave action and wind-induced disturbance of the sedimentary record (Section 3.2.1). Lakes situated along the valley sides close to the valley floor are often sheltered by the undulating topography of low ridges and rocky knolls (Section 3.2.1).

(5) Basins with vegetated margins: aquatic and near-shore vegetation was depicted on the 1:10 000 map series and could be further identified from air photographs. Present-day aquatic vegetation could indicate that these types of vegetation were also present in the past. The interaction of aquatic vegetation and changes in water depth can be interpreted as fluctuations in lake-level (Sections 3.3.1, 3.3.4). The nature and extent of the aquatic vegetation within potential sites would be defined during field survey.

Two potential sites were identified from the above desktop survey. These sites were then further investigated in field survey, in particular the nature of the outflow, the sediment and water inputs and the extent and diversity of aquatic vegetation.

<u>Loch an Fheadain</u> (GR NH 117207) (Figures 2.2, 8.9, 8.10) is a very small (15 metres in diameter) well vegetated basin. This basin was considered unsuitable as:

(1) the outflow stream was difficult to define and in particular did not appear to flow over a rock bar suggesting that in the past incision could have taken place with the increase in stream discharge dampening the climatic signal (Section 3.2.2).

(2) the basin lay <10 metres above the River Affric and the outlet stream flowed into the River Affric. It was felt that the hydrology of the basin could be controlled in part by fluvial processes operating within the river. For example the base level of the outlet stream may have been lowered by incision by the River Affric resulting in a fall in lake-level.

(3) from the surrounding topography it was difficult to determine whether this basin was lying on impermeable bedrock or more permeable glacial till or fluvioglacial sands and gravels.

Loch Coulavie was the second site. Loch Coulavie lies 0.5 km north west from the head of Loch Affric at 250 m asl, GR NH130210, Latitude 57° 15' N and Longitude 5° 6' W (Figure 2.2). The following site description of Loch Coulavie is made with reference to the site selection criteria as outlined in Chapter 3.

4.2 BASIN MORPHOMETRY

Loch Coulavie is a small basin 10 ha in area with an irregular basin morphometry. It is made up of a narrow, 70 m wide by 300 m long, north-east section and a wider, 180 m wide by 330 m long, south-west section. These two sections are separated by a narrow stretch of water 15 m wide (Figures 4.1, 4.5).

Approximate water depths using a graduated survey pole were taken from a small rowing boat in both lake sections (north-east and south-west) and across the narrow stretch of water. Care was taken not to push the survey pole into the sediment substrate. The larger south-west section deepened quickly at the margins to >2 m depth in the centre. The substrate was rocky with aquatic vegetation restricted to a narrow strip (<3m wide) around the lake margin (Figure 4.1). Within the narrow stretch of water, water depths shallowed rapidly to around 90 cm. Bedrock is exposed on both shores and beneath the water at this section of the loch (Figures 4.1, 4.5). Within the north-east section the water depth at the deepest point was around 1.5 m.

Figure 4.1 Photograph of Loch Coulavie taken looking west across the lake. Within the lake the deeper south-west section is separated from the well vegetated shallower north-east section by a narrow area of shallow water (Section 4.2). The topography around the basin is mainly undulating bedrock covered by blanket peat (see Figure 4.5). In the foreground the well vegeated relict fans and surface stream channel are visible.

The substrate consisted of organic-rich lake mud, suggesting that this section of the lake acts as a sediment trap and sediment deposition is the dominant process. This section of sediment accumulation within the basin would indicate that there is potentially a sediment record that is continuous, highly resolved and sensitive to changes in water depth. The lake margins in this section were well vegetated with aquatic plants noted up to 50 metres from the shore (Figures 4.1, 4.3).

The basin morphometry suggests that the loch can be divided into two sections by a narrow bedrock bar which is at present covered by water. This rock bar could potentially operate as a threshold within the lake. A fall in lake-level would isolate the smaller north-east section effectively ensuring no output. The increased sensitivity of a basin once the lake-level falls below the level of the outlet has been discussed in Section 3.2.2.

4.3 CATCHMENT HYDROLOGY

The surface catchment hydrology for Loch Coulavie is shown in Figure 4.2. The present-day catchment of Loch Coulavie includes several small streams, the largest of which is the Allt Dubh. The smaller streams rise at around 400-550 m asl and flow down the steep valley slopes (20-31°) and into the south-west section of Loch Coulavie. The Allt Dubh rises at around 700 m asl, flows over gentler higher altitude slopes (13°) then down the much steeper slopes of the valley sides and into the north-east section of the loch. All the streams entering the loch are sourced from the steep valley slopes above the lake. The present day catchment area is 1.75 km^2.

In the present day catchment the Allt Coulavie stream does not flow into Loch Coulavie but instead flows directly into Loch Affric (Figures 4.2, 4.3). The catchment of the Allt Coulavie includes Coire Coulavie and is 3.7 km^2 in area rising to an altitude of >900 m (Figure 4.2). The watershed between the Allt Dubh and the Allt Coulavie is well defined at altitude by a narrow ridge; however, at lower altitudes close to Loch Coulavie the watershed is less well defined (Figures 4.2, 4.3). At this point on the watershed the 1:10 000 topographic map shows a small stream flowing west towards the Allt Coulavie and the Allt Dubh flowing directly into Loch Coulavie (Figure 4.2).

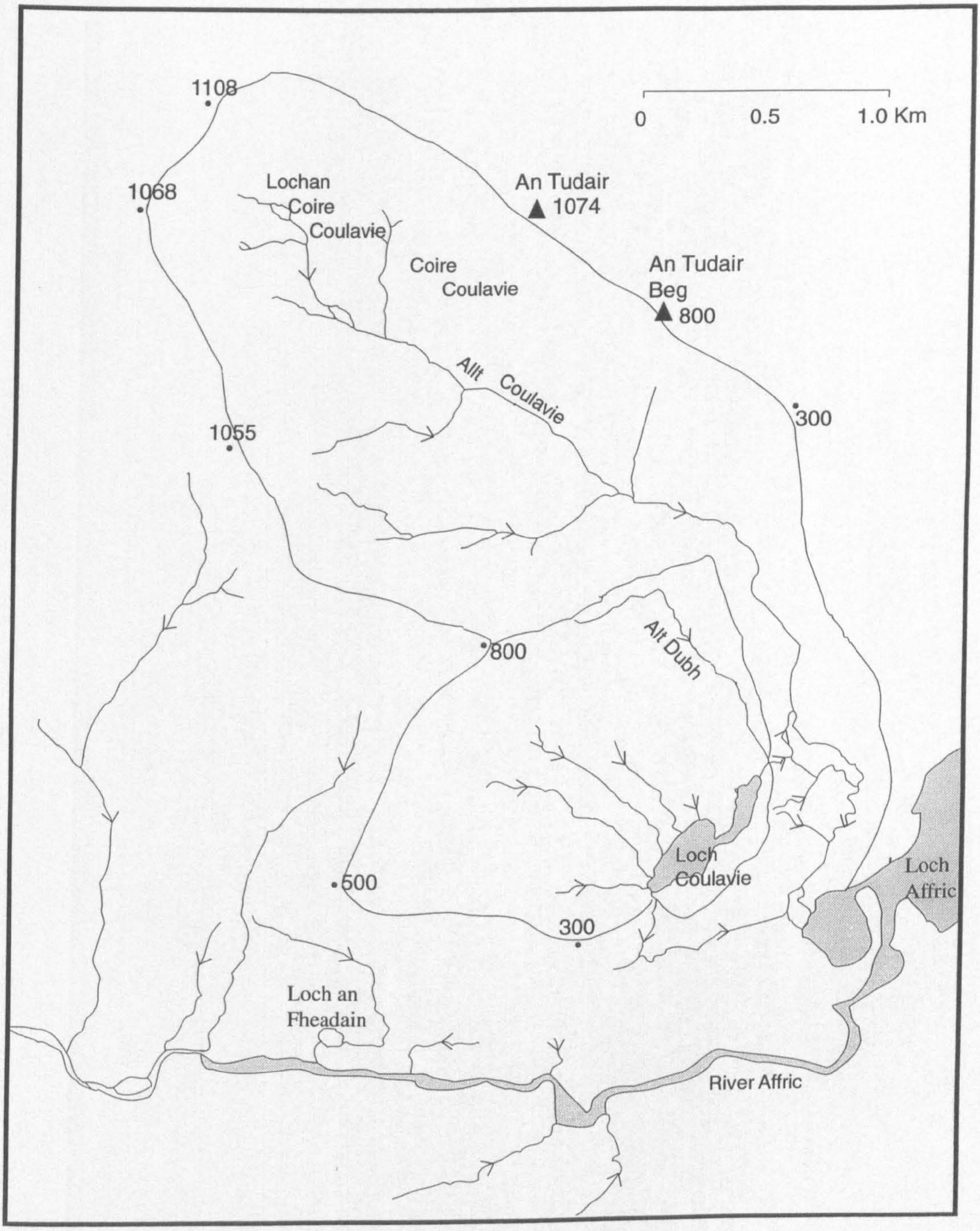


Figure 4.2 Catchment area for Loch Coulavie and the river Allt Coulavie (black line) (Section 4.3). Spot heights are in metres.

Figure 4.3 Photograph of the north-east section of Loch Coulavie showing the extent of marsh and fen vegetation within this section of the lake (see Figure 5.1). The Allt Coulavie is visible (to the left) flowing past Loch Coulavie. The eroded peat that marks the watershed between the catchments of the Allt Coulavie and Loch Coulavie is also visible (middle foreground, left) (Section 4.3). Active debris cones are visible on the slopes above Loch Affric (background) (Section 4.4).

This would suggest that at present there is a well defined watershed between the two streams at this point. In the field however the evidence for this small stream flowing into the Allt Coulavie is less well defined as the area is complicated by a series of eroding peat hags (Figure 4.3). This poorly defined watershed between the Allt Dubh and the Allt Coulavie at lower altitude suggests that there is the possibility that water and sediment from the Alt Coulavie catchment may have entered Loch Coulavie.

Loch Coulavie has a single outlet stream. This outflow is a small stream which at present flows out over a rock bar (Figure 4.4). A boulder-rich till (Figure 4.4) caps the rock bar, which has since been incised (up to 1 metre) by the stream. However, the presence of the rock bar suggests that during periods of increased discharge the stream cannot erode through the stream bed (Dearing and Foster (1986), Section 3.2.2 (a)). The rock bar ensures that the height of the outflow has not altered over time and that lake-level can fluctuate below this outlet level (Section 3.2.2 (a)). The presence of an outlet does represent a limit on the highest lake-level with potential magnitudes of lake-level rises underestimated as the highest lake level is constrained by the level of the outflow.

The Loch Coulavie basin lies on highly impermeable bedrock (Moinian meta-sediments Section 2.2, Figure 2.3) and so groundwater inputs and output are considered to be minimal (Section 3.2.2 (b)).

4.4 SEDIMENT SOURCES AND AVAILABILITY.

The following section outlines potential sediment sources and sediment availability within the Loch Coulavie basin (Section 3.2.2 (d)). Sediment type reflects the geology of the catchment; therefore within the Loch Coulavie catchment the psammitic and mica schist rock types (Section 2.2) would generate a sediment consisting of predominantly quartz and mica.

The glacial history of Glen Affric (Section 2.3) would suggest that during the last period of ice cover (the Loch Lomond Readvance) Loch Coulavie was well within the ice limits (Figure 2.2) but that higher altitude peaks may have remained ice free.

Figure 4.4 Loch Coulavie outlet stream flowing over a bedrock bar. The outlet stream has incised through a local patch of glacial till capping the bedrock (Section 4.4).

Therefore the Loch Coulavie basin and catchment would have been affected by predominantly erosive processes during the Loch Lomond Readvance period (Section 3.2.2 (d)). Field evidence would agree with this assumption with Loch Coulavie formed in an ice-scoured bedrock basin lying in a topography of predominantly peat-covered bedrock knolls and rouche moutonnées (Figures 4.1, 4.3). Till is restricted to small localised patches around the loch and consists of angular, poorly sorted, locally derived psammitic and mica schist rock types in a mica-rich sand matrix (Figure 4.4). Within the Loch Coulavie catchment there is no evidence for fluvioglacial deposits. Within the high-altitude Corie Coulavie the corrie walls have extensive scree and talus debris slopes possibly derived from exposure and periglacial processes during the Loch Lomond Readvance period (Section 2.3).

The alluvial fans along the northern shore of Loch Coulavie beneath the steep valley side slopes would suggest that much of the sediment available for sediment transport within the Loch Coulavie catchment has already been reworked and deposited as these series of fans (Figure 4.5). These fans are now considered as relict features with slopes extensively well vegetated and no evidence for the active deposition of very coarse poorly sorted alluvial fan sediments (Figure 4.1). Field observations would suggest that these fan sediments may have been extensively reworked across the fan surface itself with sediments noted to consist of coarse to medium sands.

At present, streams entering Loch Coulavie flow over bedrock on the upper slopes and alluvial fans on the lower slopes (Figures 4.2, 4.5). Sediment supply and sediment inputs to the lake will have changed over time. Of particular concern is the role of sediment supply from the alluvial fans into the lake system over time, further discussed in Sections 7.7 and 9.3.

Figure 4.5 shows mapped landcover and geomorphic features within the Loch Coulavie basin catchment. The main geomorphic features are the alluvial fans with few other features. It is thought that geomorphic activity as a direct input of sediment to the Loch Coulavie basin (Section 3.2.2(d)) was limited.

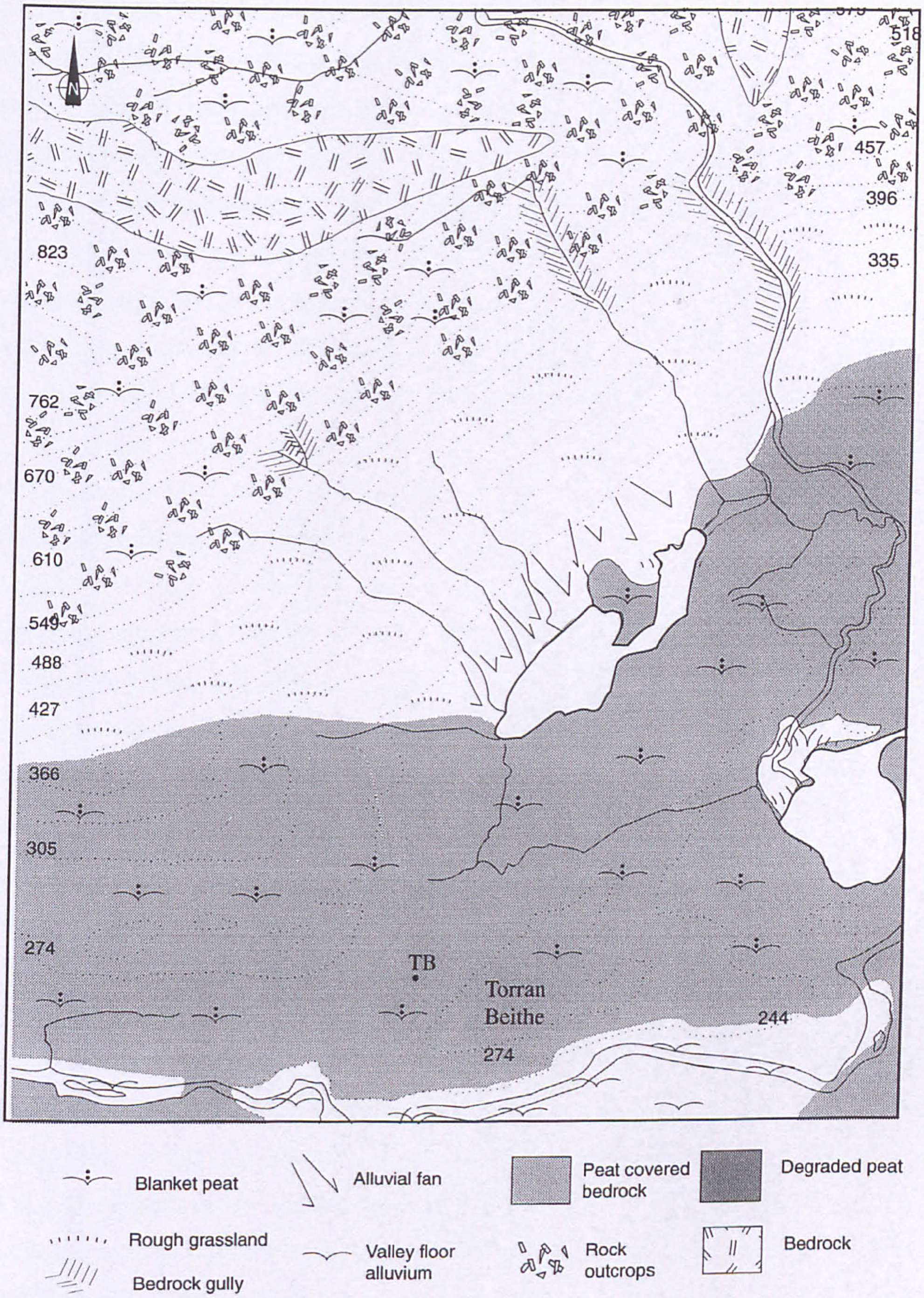


Figure 4.5 Geomorphological features and landcover within the Loch Coulavie catchment (see Figure 4.2). TB marks the pollen analysis site Torran Beithe (Section 4.7). Contour intervals are in metres.

Soils within the catchment are predominantly peats and peaty podzols (Section 2.4). Section 3.2.2 (c) discussed the limited impact the spread of blanket peat would have on the hydrology of the catchment. However, such extensive blanket peat cover may have a control on slope stability with peat effectively stabilising slopes and reducing sediment supply to the basin. The role of peat and sediment availability is complex as once blanket peat is established the erosion of this substrate can then act as an organic sediment source within the catchment. The potential timing, rate of spread and forcing mechanisms for peat accumulation within the Loch Coulavie catchment are discussed in Section 8.6 with its role in controlling sediment supply discussed in Section 9.3.

4.5 PRESENT-DAY CLIMATE.

Present-day climate within the Loch Coulavie catchment is effectively similar to that described for Glen Affric (Section 2.6), with the dominance of Atlantic cyclonic air masses and associated frontal systems controlling the temperature, amount of precipitation and wind direction. Local temperature variability is due to controls such as altitude, seasonal katabatic cooling and aspect of slope (Section 2.6.2). Loch Coulavie lies in the relatively wetter west of Glen Affric as described by the arbitrary 500 mm PWD isopleth which is interpreted as lying 6-7 km to the east of the lake (Section 2.6.3). Seasonal variations in precipitation are reflected in seasonal fluctuations in lake-level with (from personal observations) the summer lake-level 30-40 cm lower than the winter level. The amount of snowfall and the duration of snowlie within Glen Affric was difficult to define. Within the Loch Coulavie area snow (from personal observation) accumulates within the high-altitude (>800 m asl) north-east facing Corie Coulavie and persists until June or early July. The prevailing wind direction is west to south-westerly and funnelled down Glen Affric would blow across the full length of Loch Coulavie. The north-east section of Loch Coulavie is sheltered from these prevailing winds owing to the undulating topography (Figures 4.1, 4.5).

4.6 PRESENT-DAY VEGETATION COVER

Vegetation cover is described using field survey data.

- Bedrock knolls and the accumulated peat cover support blanket mire species tolerant of acidic water logged conditions (Section 2.6) with species such as *Molinia caerulea, Eriophorum vaginatum, Calluna vulgaris* and *Sphagnum* species.
- The more nutrient-flushed alluvial fans along the north shore of Loch Coulavie support a herbaceous grassland.
- Above the alluvial fans the steep slopes (22-30° up to 500 m asl) are relatively well drained and the peaty podzols support coarse grassland with species such as *Nardus stricta* and in the waterlogged hollows, blanket mire species. Tree species, for example *Sorbus* and *Betula*, in the catchment are restricted to gullies on the upper slopes.
- The aquatic and near-shore vegetation around Loch Coulavie again reflects the poor nutrient status and acidity of the parent rock. Aquatic vegetation (identification of species from Stace 1995 and Haslam *et al.* 1975) varies with water depth: thus in deeper waters (>1m) *Potamogeton* sp. is noted and with decreasing water depths towards the shore (~1m) *Equisetum* sp. and *Myriophyllum alterniflorum.* Submerged mats of *Menyanthes trifoliata* gave an often undulating surface to the lake bed. At shallow water depths ~40 cm *Carex rostrata* forms the dominant vegetation type along with *Equisetum* sp. With increasingly shallow water (10-20 cm) more terrestrial species are noted such as *Myrica gale*, *Carex echinata*, *Juncus squarrosus* and *Sphagnum* sp. (Figure 5.1).
- On-shore vegetation consists of acid mire plants such as *Eriophorum angustifolium, Myrica gale, Molinia caerulea, Erica tetralix, Calluna vulgaris, Potentilla erecta, Drosera* sp. and *Sphagnum* sp. Further inshore the surface becomes slightly drier and more *Calluna vulgaris* and *Molinia caerulea* are noted.

Based on freshwater loch plant associations the aquatic vegetation types around Loch Coulavie can be classed as swamp and fen type vegetation (Spence 1964; Walker 1970). Swamp vegetation is classed as fully aquatic with species rooted within the substrate and standing in permanent water (for example *Carex rostrata*). Fen vegetation is associated with partial aquatic species where the water table during the summer lies below the soil surface and in the winter the surface is flooded. Typical fen species include grasses such as *Molinia caerulea* and sedges along with a variety of acid-loving plants rooted in an organic substrate (Spence 1964; Walker 1970). It is difficult to separate fully swamp (aquatic) from fen (terrestrial) as swamp and fen type vegetation will seasonally transgress depending on the water level (Spence 1964). Within the nearshore vegetation of Loch Coulavie typically mire species are noted, which would indicate that within the aquatic species, interactions are in response to paludification processes and the spread of mire species (Klinger 1996).

Once established both fen and swamp vegetation will accumulate as organic peat with macrofossil plant remains preserved under the permanently waterlogged conditions. The type of nearshore vegetation over time is primarily dependent on water depth (Spence 1964) (Section 3.3.1). The water depth will alter over time through processes such as sediment infilling, terrestrialisation and through climatically forced changes in lake-level. The presence of extensive aquatic vegetation such as fen and swamp within a lake has been shown to have persisted for thousands of years without becoming fully terrestrialized, that is through forest invasion (Klinger 1996; Walker 1970). Within the accumulated peat-rich marginal sediments of Loch Coulavie there is a potential record of changes in aquatic vegetation in response to lake-level fluctuations (Sections 3.3.1, 3.3.4).

4.7 HOLOCENE VEGETATION HISTORY

A high-resolution palynological record from basin peat at Torran Beithe (NH 130210, 260 m asl) (Figures 2.2, 4.5) provides a detailed Holocene vegetation history (Davies 1999) (Section 3.2.2 (e)). The basin lies on the undulating peat covered northerly slopes of Torran Beithe, a rocky ridge rising to 302 m asl (Figure 4.5). The present-day basin has a diameter of 56 metres and the catchment is well defined by enclosing rock ridges. Davies (1999) suggests that the pollen source area for the basin is approximately 100-300 metres. Loch Coulavie lies 350-450 metres from the sampling site at Torran Beithe but the vegetation record from this pollen site can be used to describe the vegetation history of the Loch Coulavie catchment (Figure 4.5).

Sediment type	Pollen zone	Chronology	Vegetation type
Sphagnum peat	Cyperaceae-Poaceae-*Calluna-Pinus*	190-0 BP (150-0 cal. BP)	Evidence for continued pastoral grassland cultivation. Permanent shift from *Calluna* to Poaceae and Cyperaceae at 180 BP (150 cal. BP).
Increasingly *Sphagnum* rich peat	*Calluna*-Cyperaceae-*Sphagnum*	1290-190 BP (1200-150 cal. BP)	Changes in blanket peat vegetation associated with increased amounts of burning indicated by charcoal. Presence of taxa associated with pastoral or disturbed grassland from 550 BP (540 cal BP). Evidence for cereal cultivation c. 410 BP (490 cal. BP).
Herbaceous peat	*Calluna*-Cyperaceae	3600-1290 BP (3890-1200 cal. BP).	Spread of blanket peat associated with gradual woodland reduction. Still substantial local *Pinus* populations but gradually declining. Increase in grassland taxa from 2760 BP (2850 cal. BP).
Peat with *Pinus* and Cyperaceae remains	*Pinus-Betula-Calluna*	6360-3600 BP(7230-3890 cal. BP)	Predominantly *Pinus-Betula* woodland with limited distributions of *Ulmus* and *Quercus, Corylus*. A peak in *Alnus* at 5630 BP (6410 cal. BP). Abrupt decline in *Pinus* at 3765 BP (4110 cal. BP) associated with temporary increase in *Quercus, Alnus, Corylus* and *Betula*.
Amorphous organic mud to peat (6590 BP(7400 cal. BP)	*Pinus-Betula*-Cyperaceae	6720-6360 BP (7540-7230 cal. BP)	*Pinus* established as the major canopy tree around the basin (6550 BP (7390 cal. BP) associated with decreasing amounts of *Betula*.
Highly organic mud. Minerogenic inwashing (6930-6790 BP:7690-7580 cal. BP)	*Betula –Pinus sylvestris-Calluna vulgaris-Corylus avellana* type.	7890-6720 BP (8590-7540 cal. BP).	Transition to woodland communities with increasing amounts of *Pinus*, and local presence of *Quercus* and *Ulmus*. *Pinus* established around basin by 7205 BP (7950 cal. BP) with increasing amounts of *Calluna*.
Amorphous gyttja and minerogenic lake mud.	*Betula-Corylus avellana* type.	8800-7890 BP (9740-8590 cal. BP).	Open grassy *Betula-Juniperus* scrub and *Empetrum* heath. Early Holocene woodland development ie *Betula* and *Corylus* shades out *Juniperus* scrub.

Table 4.1 Summary of Holocene vegetation history as generated by the palynological record from Torran Beithe (Davies 1999); see text for further discussion of the major changes in vegetation.

Table 4.1 Summarises the main pollen zones, stratigraphy and vegetation history recorded at Torran Beithe. The timing and nature of major changes in vegetation cover are discussed below.

4.7.1 HOLOCENE VEGETATION HISTORY

The basal pollen zone at Torran Beithe *Betula-Corylus avellana* type (*c.* 8880-7890 BP (9740-8590 cal. BP)) records an initial vegetation cover of *Empetrum* and *Juniperus* and implies a compressed early post-glacial sequence due to slow sediment accumulation rates (Davies 1999). Early Holocene (*c.* 7890 BP (8590 cal. BP)) woodland cover is predominantly *Betula* and *Corylus* with the presence of *Pinus sylvestris*, with local establishment of *Pinus sylvestris* by *c.* 7205 BP (7950 cal. BP) and full establishment and dominance of *Pinus sylvestris* woodland by *c.* 6550 BP (7390 cal. BP).

Between *c.* 6390-3765 BP (7230-4110 cal. BP) vegetation cover is predominantly *Pinus-Betula* woodland and *Calluna*. Within this period there are long-term and short-term fluctuations in *Pinus*; during the long-term fluctuations in relative abundance of *Pinus* and *Calluna*, *Pinus* was able to maintain dominance, with other tree taxa such as *Quercus* and *Alnus* remaining relatively minor.

At *c.* 3765 BP (4110 cal. BP) there is an abrupt reduction in *Pinus sylvestris* with the reduction in tree cover occurring over *c.* 60 ^{14}C years (70 cal. years). However with the decline in *Pinus* there was a temporary increase in *Quercus, Alnus* and *Corylus*. So although the *Pinus* decline represents an abrupt vegetation change to a more open deciduous woodland, potentially the amount of woodland cover was not altered. After *c.* 3600 BP (3890 cal. BP) there was a gradual decline in woodland cover and an increase in the spread of blanket peat. Blanket peat continued to spread with the further reduction of tree taxa from *c.* 2030 BP (1970 cal. BP) with *Pinus* remaining as a local tree taxon. The reduction of tree taxa is not thought to have been a steady process but may have occurred in pulses, reflecting small-scale or subtle local environmental gradients and thresholds for tree regeneration (Davies 1999).

From *c.* 2760 BP (2850 cal. BP) there is an increase in Poaceae and grassland taxa. From around 1250 BP (1170 cal BP) changes in the blanket peat species are thought to be in response to increased amounts of burning as suggested by increasing amounts of charcoal and associated vegetation changes, principally *Calluna* to Cyperaceae to *Calluna*. Pollen analysis would suggest that by *c.* 190 BP (150 cal. BP, AD *c.*1800) the area was virtually treeless with vegetation cover similar to the present day.

4.7.2 ANTHROPOGENIC ACTIVITY

Archaeological evidence for anthropogenic activity within Glen Affric is limited (Davies 1999). The palynological record is therefore used to infer periods of human activity. A rise in ruderals from *c.* 2760 BP (2850 cal. BP) could possibly be the result of anthropogenic activity and grazing but with low pollen frequencies and sporadic occurrence of taxa this is difficult to interpret fully and differentiate from the impacts of grazing from native mammals. From around 1250 BP (1170 cal. BP) and in particular from around 550 BP (540 cal. BP) evidence from the charcoal record would suggest that vegetation changes were the result of increased anthropogenic activity. Pollen taxa associated with disturbed ground increase at this time with a cereal grain recorded at *c.* 410 BP (490 cal. BP) implying local cereal cultivation and pastoral activity possibly on the more minerogenic alluvial soils around Loch Coulavie (Figure 4.5).

4.8 SITE SELECTION SUMMARY

The following discussion highlights the potential advantages and limitations of using Loch Coulavie as a site for the interpretation of lake-level fluctuations as a proxy climatic record.

- Loch Coulavie represents a lake sensitive to climatically forced changes in the hydrological balance within the catchment. Catchment-based surface inputs and not groundwater dominate the hydrological inputs into the lake (Section 3.2.2 (b)). Variations in precipitation are likely to control lake-level. At present the lake-level is known to alter on a seasonal basis, thus any further long-term enhancement of the seasonal variations in rainfall over evaporation would give rise to potentially sustained changes in lake-level.

- The basin morphometry of Loch Coulavie suggests that changes in sediment composition within the north-east section will provide a sensitive record of lake-level fluctuations. The narrow rock bar provides a mechanism for this section of the lake to become closed during low lake-levels (Section 3.2.2 (a)). This closed basin would be more sensitive to fluctuations in lake-level which would potentially flip this section of Loch Coulavie from a small closed basin during low lake-levels to a much larger open basin during high lake-levels.

- The sediment record from the north-east section of the lake is likely to provide the highest resolution as it is in this section that sedimentation is the dominant process. The sediments within this sheltered section are also less likely to have been disturbed by wave action (Section 3.2.1). The aquatic vegetation noted within this section would have responded to changes in lake-level and thus provide another line of evidence in the interpretation of the sedimentary record.

- Within the Loch Coulavie catchment it is thought that geomorphic activity such as slope processes may have been limited owing to the low amounts of glacial till deposited within the catchment (Section 4.4 and 3.2.2(d)).

- In Sections 3.2.2 (e) and (f) the importance of palaeoecological data for the site catchment was discussed. The high-resolution palynological record from the peat-filled basin at Torran Beithe (Section 4.7) provides a record of the major changes in Holocene vegetation cover. These data can be inferred for the Loch Coulavie catchment. These major changes in vegetation cover within the catchment and the subsequent impacts on the catchment hydrology can be compared to the lake-level record generated. Thus for this site potential climatic forcing through changes in precipitation can be separated from changes in the vegetation cover affecting evapotranspiration and surface flow.

- The palynological record (Section 4.7.2) also indicates periods of human activity within the Loch Coulavie region. The data suggest that for this area with extensively bedrock and blanket peat cover, human impact was restricted (Davies 1999), with the most reliable evidence for activity from around 1250 BP (1170 cal. BP) and more specifically from around 550 BP (540 cal. BP).

Limitations of using Loch Coulavie include;

- The presence of an outflow stream, which is discussed in Sections 3.2.2 (a) and 4.3. The level of the outflow stream represents the maximum level the lake can rise to and thus the level of any lake-level rise identified from the sediment record will be underestimated.

- Allogenic sedimentary inputs will alter the sediment composition but will not represent a response to a fluctuation in lake-level. Allogenic sediment inputs, in particular fluvial deposits, are in response to non-climatic and climatic controls within the catchment. The catchment area:lake area ratio for Loch Coulavie and its present-day catchment is 15:1 and as suggested by Dearing and Foster (1993), the sediment inputs to the lake are dominated by fluvial processes. As discussed in Sections 3.2.2 (c) and (d) fluvial inputs of water and sediment have a range of non-climatic controls. At Loch Coulavie the alluvial fans built up along the north shore and associated streams represent a sediment input with climatic and non-climatic controls which will have altered over time. Interpretation of the sedimentary record with Loch Coulavie must take into account these potential sediment inputs (Sections 3.3.2, 3.4). An approach incorporating specific sediment analysis techniques is required to differentiate between variations in the sediment composition due to allogenic sediment inputs and those resulting from fluctuations in lake-level (Section 3.4).

- Changes in catchment area. As outlined in Section 4.3, within the Loch Coulavie catchment sediment and water inputs from the Allt Coulavie catchment could have entered Loch Coulavie. Within the sedimentary record it would be difficult to identify this type of event.

Loch Coulavie represents a site that is climatically sensitive, with a sediment record that if fully investigated using a robust methodological approach can be interpreted as a series of changes in lake-level. Pollen analysis and further definition of the role of blanket peat within the catchment (Section 3.2.2 (c)) provide clarification of the climate record generated from proposed changes in lake-level, removing the ambiguity of potential non-climatic forcing mechanisms such as vegetation change.

PART 2 LAKE-LEVEL FLUCTUATIONS

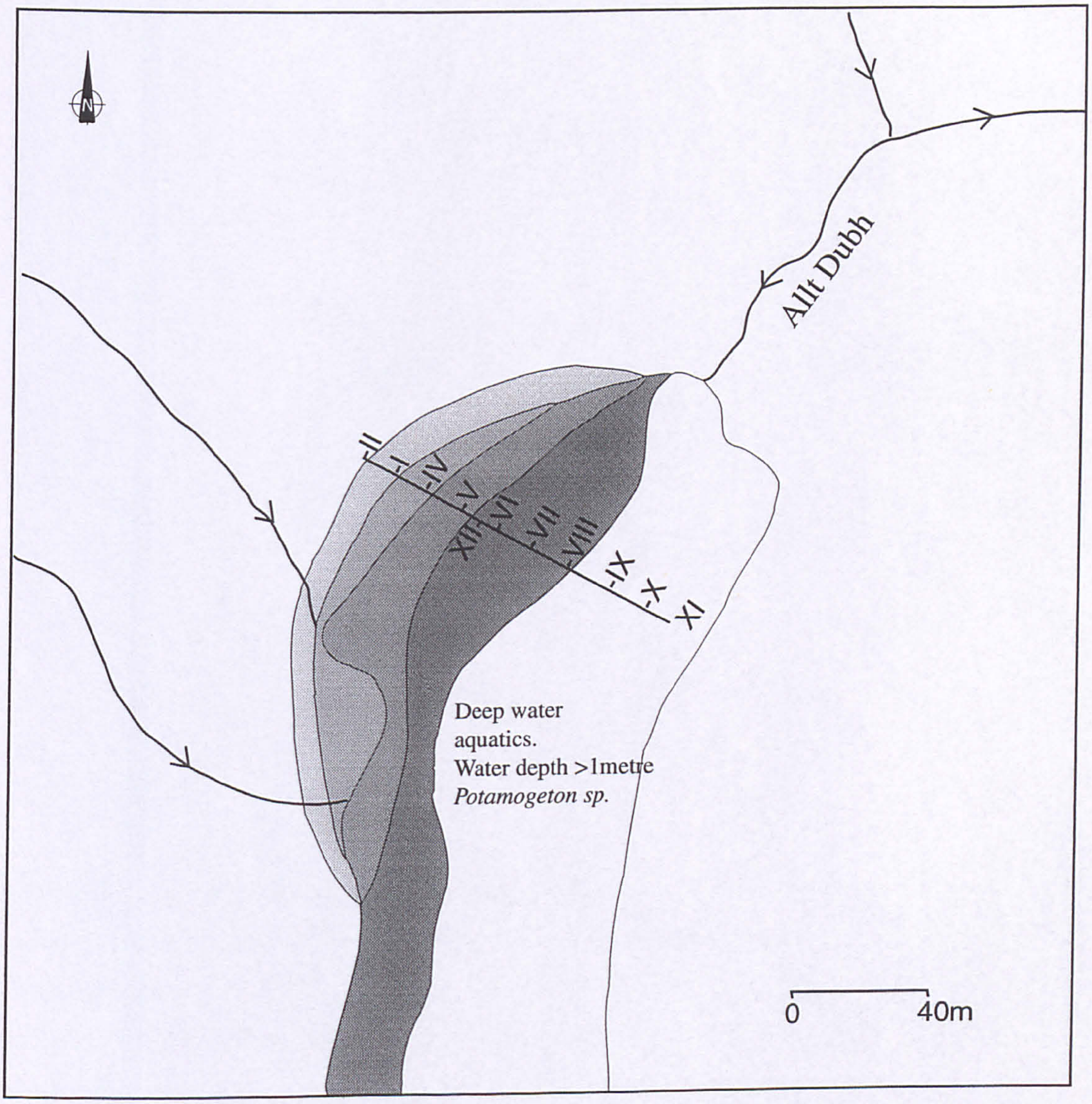
CHAPTER 5
METHODS

5.1 CORING STRATEGY

Site selection, as outlined in Section 4.9, identified within Loch Coulavie a sheltered, well vegetated embayment in the north-east section. This contained a sedimentary record subjected to minimum disturbance from non-climatic sedimentary controls. Evidence within this sedimentary record would indicate fluctuations in lake-level (Sections 3.3 and 3.4).

Digerfeldt (1986) outlines a coring strategy using a transect of cores taken from the shore (basin edge) across the basin. Digerfeldt (1986) suggests that distance between cores largely depends on the character and slope of the lake margin but the difference in depth, to bedrock, between each core should be about one metre. A transect of cores should ensure that any changes in lake-level will have been recorded by more than one core (*cf.* Bradbury and Dean 1993; Edwards and Whittington 1993; Gaillard and Digerfeldt 1991; Hannon and Gaillard 1997; Magny 1992).

A transect was positioned across the embayment in the north-east section of Loch Coulavie running south-east to east from the shore out across the lake (Figure 5.1). Along the transect at 10 metre intervals the water depth and sediment thickness were determined using a 1.0 m Eijelkamp gouge corer. Each site was marked by a survey pole. Sites were levelled and surveyed and used the lake surface as a datum. Sediment thickness, lake water depth and bedrock slope (Figure 5.2) were used to determine the coring sites for laboratory sediment analysis.



Equisetum sp., *Myriophyllum alterniflorum, Menyanthes trifoliata* with some *Carex rostra.* Water depth ~1m.

Predominantly *Carex rostra.* Water depth ~40 cm.

Carex echionata, Juncus squarrosus, Myrica gale and *Sphagnum* sp. Water depth 10-20 cm.

Marsh/fen blanket peat transition, increasing amounts of acid mire plants such as *Erica tetralix* and *Eriophorum angustifolium.*

Figure 5.1 Location of sites used to determine water depth (Figure 5.2) and sampling sites (LCII, LCI, LCIV, LCXII and LCVII) along a transect across the north-east section of the Loch Coulavie basin (see Figure 4.3). Zones of marsh and fen vegetation types across the basin as discussed in Section 4.6 are shown.

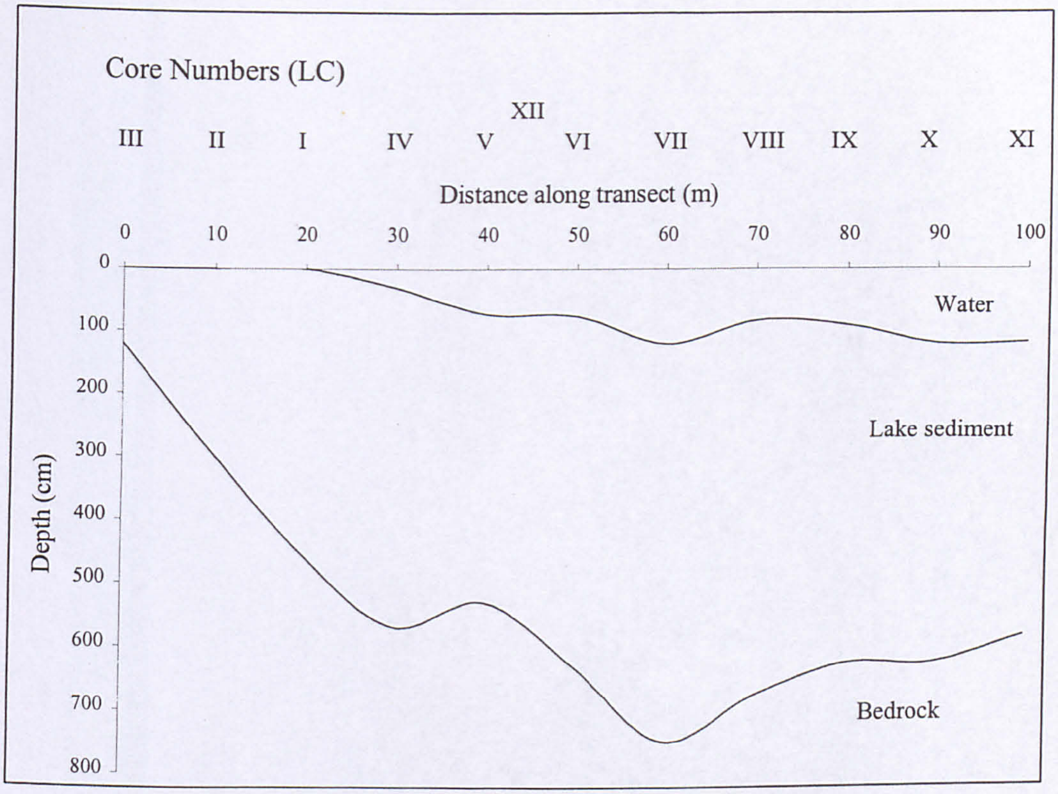


Figure 5.2 Loch Coulavie basin morphometry along coring transect within the lake (see text and Figure 5.1).

Using the basin morphometry (Figure 5.2) sediment cores for laboratory analysis were taken from sites LCII, LCI and LCIV as these sites represented sedimentation at the basin margin. Site LCVII is the deepest part of the basin with sediment cores for laboratory analysis from this site representative of deeper water sedimentation. To define the transition from marginal to deeper water sedimentation a site equidistant between LCIV and LCVII was sampled and labelled as LCXII (Figures 5.1 and 5.2).

All the sites were cored using a Russian corer of 1.0 metre in length and 6.0 cm internal diameter (Jowsey 1966). Overlapping cores and duplicate core sets were taken to ensure good core correlation and an adequate volume of sediment for analysis. Sample cores were placed in lengths of guttering, wrapped, labelled and once in the laboratory stored in the dark at 4°C. Coring of deeper water lake sites, LCXII and LCVII was carried out from a purpose-built raft anchored to the lake sediment (Figure 5.3).

Figure 5.3 The raft constructed as a platform to core deeper water sediments; the small dingy was used to store core samples.

Within these deeper water cores there were difficulties in sampling the surface metre of sediment which consisted of well-rooted aquatic vegetation and no sediment. To ensure consistency of the surface datum, lake cores were sampled in the same summer (LCXII and LCVII sampled on the same day) and against an arbitrary lake-edge datum.

5.2 SEDIMENT STRATIGRAPHY

Detailed sediment logs, X-ray analysis and magnetic susceptibility (Sections 5.2.1-5.2.3) further defined potentially synchronous sedimentary events within the stratigraphy, and allowed for accurate core correlation (Dearing 1986). The application of these techniques to the lacustrine sediments within Loch Coulavie is discussed in the following sections.

5.2.1 SEDIMENT DESCRIPTION

An adaptation of the Troels-Smith notation scheme (Aaby and Berglund 1986) was used. Field descriptions described the components, assessing sediment parts composition with a scale of 0-4, (0-100%) (Table 5.1). Traces (<10%) of other components elements were noted with a '+' symbol.

Field descriptions were further refined under laboratory conditions. After the sediment surface had been carefully cleaned sediment was described under fluorescent lighting, using the descriptive components as shown in Table 5.1 and the physical properties of the sediment, again using a scale of 0-4 (Table 5.2). The colour of the fresh sediment surface was noted together with a descriptive name for the sediment type and the humicity of organic sediment components.

5.2.2 X-RAY ANALYSIS

X-ray analysis of soft sediments, such as soils and peats, reveals in greater detail and definition, structural and stratigraphic details that are often invisible or unclear to the naked eye (Butler 1992; Dörfler 1992; Saarnisto 1986).

	Sh	Substantia humosa	Humus substance, homogeneous microscopic structure
I Turfa	Tb	T. Bryophytica	Mosses
	Tl	T. lignosa	Stumps, roots, intertwined rootlets of ligneous plants +/- trunks, stems, branches, etc. connected with these.
	Th	T. herbacea	Roots, intertwined rootlets, rhizomes, of herbaceous plants +/- stems, leaves etc.
II Detritus	Dl	D. lignosus	Fragments of ligneous plants > 2 mm.
	Dh	D. herbosus	Fragments of herbaceous plants > 2 mm.
	Dg	D. granosus	Fragments of ligneous and herbaceous plants and sometimes of animal fossils <2mm > c.0.1 mm.
III Limus	Ld	L. detrituosus	Plant and animals (except diatoms, needles of spongi, siliceous skeletons etc.) of organic origin or fragments of these. Particles < c. 0.1 mm.
	Lso	L. siliceous organogenes	Diatoms , needles of spongi, siliceous skeletons etc. of organic origin or parts of these. Particles > c. 0.1 mm.
	Lc	L. calcareous	Marl, not hardened like calcareous Tufa: lime and the like. Particles < c. 0.1 mm.
	Lf	L. ferrugineus	Rust, non-hardened. Particles < c. 0.1 mm.
IV Argilla	As	A. steatodes	Particles of clay < 0.002 mm
	Ag	A granosa	Particles of silt 0.06 to 0.002 mm
V Grana	Gmin	G. minora	Particles of sand 2-0.6 mm
	Gmaj	G. majora	Particles of gravel 60-2 mm.

Table 5.1 Deposit elements of the modified Troels-Smith sediment description scheme used in Loch Coulavie sediment description (after Aaby and Berglund 1986).

Name	Abbreviation	Description
Nigor	nig	The degree of darkness, Nig0, light e.g. quartz sand Nig4, black e.g. humified peat
Stratificatio	strf	Stratification can be structural or visual, Strf0, homogenous sediment Strf4, sediment consisting of many minor layers
Elasticitas	elas	The ability to regain shape after deformation. Elas0, no elasticity Elas4, highly elastic
Siccitas	sicc	Water content Sicc0, pure water Sicc1 soft sediment, runs Sicc2, water saturated, slumps Sicc3, not water saturated with water retains its shape. Sicc4, air dry sediment
Limes	lim	Contacts with adjoining sediments, Lim0, diffuse, boundary area >1cm Lim1, very gradual, <1cm >2mm Lim2, gradual, <2mm > 1mm. Lim3, sharp, <1mm >0.5 mm. Lim4, very sharp, <0.5mm.

Table 5.2 Physical properties of sediment used in sediment description as adapted from the Troels-Smith sediment description scheme (after Aaby and Berglund 1986).

X-ray analysis is non-destructive and can be carried out prior to further sedimentary analysis. Differential absorption of x-rays is controlled by variations in composition and thickness of the sediment (Butler 1992).

On the x-ray image more minerogenic sediments, with individual mineral grains greater than 0.1 mm, show up as brighter areas with a different texture as compared to the darker more amorphous image generated by highly organic sediment (Butler 1992; Dörfler 1992).

Cores from sites LCI and LCII were selected for a trial of X-ray analysis. The analysis was carried out at the X-ray facilities of the Marine Geology and Operations Group, British Geological Survey, Edinburgh using a fully portable X-ray, ScanRay AC 120L cabinet with a Tungsten source.

The sample cores were unwrapped but left in plastic guttering. Cleaning the sample surface evenly ensured that any extraneous sediment was removed, and down-core sediment thickness was relatively constant. The 1.0 m sample cores were placed within a marked area in the X-ray cabinet ensuring correct alignment beneath the x-ray head. Three exposures each around 35 cm in length were required to X-ray the whole core. Strips of photographic film (c.43cm x 11cm) were loaded into light proof cases and one placed beneath the core section to be exposed. Lead lettering was used to identify the core and the core section. Minerogenic lake sediments were exposed at 3mA and 38 kV for 2.5 minutes and more organic-rich sediments for 3mA, 30 kV for 2.5 minutes. Films were developed using black-and-white photographic film developing techniques. Film negatives were analysed visually on a light table and changes in the sedimentary type and structure recorded as variations in light and dark transferred diagrammatically onto tracing paper (Figures 6.1 and 6.3).

5.2.3 MAGNETIC SUSCEPTIBILITY

The magnetic susceptibility (κ) of a sediment is a measure of the abundance of magnetic minerals. Ferromagnetic and ferrimagnetic minerals such as magnetite and iron, have a high positive magnetic susceptibility, whereas quartz, water and organic material are diamagnetic with a low negative susceptibility (Dearing 1994). Variations in the magnetic susceptibility indicate broad changes in the proportions of these types of materials within the sediment.

A Bartington MS2 magnetic susceptibility meter and a 135 mm diameter MS2C core scanning sensor were used in a 'quiet' environment, well away from magnetic material, electromagnetic fields and sources of intermittent heat (draughts and sunlight) that would alter air temperature.

Air measurements of magnetic susceptibility were taken up to an hour before analysis to ensure that the sensor was not affected by adverse factors such as those noted above. The sediment core was unwrapped and the surface cleaned of any extraneous sediment material. The plastic guttering was marked in 2.0 cm intervals. Volumetric susceptibility was recorded in SI units (κ=SI 10^{-5}). An initial air reading was taken. Readings were not taken < 6.0 cm from each end of the core (Dearing 1986). Once the whole core had been analysed a final air reading was taken.

When measuring weakly magnetized samples (diamagnetic sediment) small increments of instrumental drift between readings must be corrected for (Dearing 1994). An error air reading for each sample measurement is generated from a linear plot of the initial and final air reading and the volume susceptibility and thus the magnetic susceptibility value (κ) for each sample corrected. Once corrected the magnetic susceptibility data for each core were then plotted graphically as κ values with depth.

5.2.4 BULK DENSITY, PERCENTAGE MOISTURE CONTENT AND PERCENTAGE LOSS-ON-IGNITION.

Bulk density, moisture content and loss-on-ignition were determined as a sequence of sediment analysis techniques as outlined by Bengtsson and Enell (1986) and Dean (1974). Bulk density and water content vary due to differences in the rate of sediment accumulation, the degree of sediment compaction and bioturbation (Bengtsson and Enell 1986). Clymo (1985, 1991) used changes in the bulk density of peat samples to define structural changes within the peat stratigraphy (Section 8.1). Variations in the loss-on-ignition (organic content) within lake sediments reflects changes in allogenic and autogenic sedimentary processes within a lake.

Bulk density is often difficult to sample for accurately with compaction of the sediment during sampling distorting the volume of the sediment (Formula (a) below) (*cf.* Anderson 1996; Blackford 1990). To try to identify an accurate and precise sediment volume sampling technique two methodologies were tested: (i) water displacement and (ii) sampling a known volume of sediment using a cork borer.

The sediment used in the study was well humified amorphous and homogenous peat sampled from a monolith tin.

(i) water displacement method: 4.0 ml of distilled water was added to a 10.0 ml graduated cylinder and weighed (in grams) to 2 decimal places. Peat was cut using a clean, sharp scalpel from the monolith tin in slices and added to the water until the meniscus reached 6.0 ml. The cylinder was then re-weighed and bulk density calculated. Three replicate samples were taken at five depths.

(ii) the cork borer method: A stainless steel cork borer of 1.0 cm internal diameter was used to sample a cylinder of peat which was put into a clean, dry pre-weighed crucible, re-weighed and the bulk density calculated. Three replicate samples were taken at the same five depths as in method (i).

The results of the mean bulk density for the three replicates at different depths, together with the standard deviation on these means, are shown in Table 5.3.

Sample depth	Bulk density using water displacement g/cm^3	Standard deviation	Bulk density using cork borer g/cm^3	Standard deviation
83 cm	0.99	0.6	1.06	0.04
93 cm	0.96	0.01	1.07	0.04
103 cm	0.95	0.02	1.09	0.04
113 cm	0.95	0.005	1.06	0.01
123 cm	0.98	0.01	1.03	0.01

Table 5.3. Bulk density to 1σ standard deviation for two methods of determining peat bulk density.

The water displacement method compressed the sediment less and shows a lower value of bulk density at all depths. It also shows greater variability about the mean for each sample, suggesting that this method is difficult to replicate accurately. The borer tends to extrude the sediment compacting the cylinder of sediment before an accurate measurement of sediment volume was made, thus increasing the bulk density value. A borer of a wider diameter would reduce the amount of force (compaction) needed to extrude the sediment.

The experiment was carried out using well humified peat, a sediment type that is relatively easily sampled. Different sediment types may be less easy to sample; for example, saturated sediments may not be retained within the corer and very fibrous sediments may have to be cut with a scalpel before sampling. The cork borer method, however, requires less re-weighing of the sample and this reduces the measurement error, and as a sampling method was faster and more efficient.

Using the results from the pilot study, wet bulk density was determined for the lake sediments at a contiguous 2.0 cm sampling interval with samples of known sediment volume, 4.0 cm^3, taken using a 1.8 mm diameter stainless steel cork borer. Samples were placed in clean, dry pre-weighed crucibles and re-weighed. Bulk density (BD) (g/cm^3) was calculated as follows:

$$\frac{\text{(Wet sample weight + crucible weight)-crucible weight (g)}}{4\ (cm^3)} \qquad \text{(Formula (a))}$$

To calculate the water content of the same sediment sample, samples were oven dried at 80°C overnight (up to 16 hours). This ensures that volatile organic carbon compounds are not lost as they would be if dried at 105°C (L. Belyea pers. com. 1997). Samples were weighed and returned to the oven at the same temperature for 3-4 hours and then re-weighed to ensure a constant dry weight. Crucibles were cooled to room temperature, in dessicators to ensure no moisture gain from the atmosphere, and reweighed. The percentage water content is calculated as follows:

$$\frac{\text{Weight loss on drying}}{\text{Wet sample weight}} \times 100 \qquad \text{(Formula (b))}$$

To determine loss-on-ignition the weighed oven-dried samples in the same crucibles were placed in a preheated muffle furnace at 550°C for 4 hours to ensure complete combustion. Crucibles were placed in dessicators, cooled to room temperature and re-weighed. The percentage loss-on ignition was calculated as follows:

$$\frac{\text{Weight loss on combustion}}{\text{Dry sample weight.}} \times 100 \qquad \text{(Formula (c))}$$

All samples were weighed and calculations made to two decimal places. The replicate samples used to test the bulk density methodology (Table 5.2) were also used to determine the variability of the loss-on-ignition method as outlined above. The six replicate samples at five depths (a total of 30 samples) had a 1σ variation of ±3.15% for loss-on-ignition. The standard deviation value determined above was used to define the variability about the mean for this specific sediment type-that is highly, organic peat. Significant changes in loss-on-ignition could be defined as those with a variability greater than the standard deviation value (Sections 6.2.4, 6.3.2). For different sediment types (that is more minerogenic), within the sedimentary record the definition of significant changes in loss-on-ignition were defined using the 1σ standard deviation value calculated for those specific sediment types. Significant changes were defined as those that deviated above or below the 1σ standard deviation value (Sections 6.4.2, 6.5.2).

Weight loss during ignition is due to the combustion of organic material and evaporation of interstitial water contained within the mineral component, of the sample and in particular from within clay minerals (Dean 1974). Although this water loss is thought to occur at temperatures greater than 550°C a direct conversion of loss-on-ignition values to organic carbon contents is unreliable.

5.3 GRAIN SIZE ANALYSIS

Within the lacustrine sedimentary record grain-size analysis was carried out using particle size distributions generated by laser diffraction analysis. The principle of laser diffraction analysis relies on a particle placed in a laser beam scattering light by reflection, refraction and diffraction. Light is diffracted from a particle over a range of angles, the pattern and angle of light diffraction dependent on particle size. Larger particles diffract very high intensities of light but only at low diffraction angles, whereas small particles diffract much lower intensities of light but over a much wider range of angles. The laser diffraction instrument measures the intensity of light received at a range of detectors, corresponding to the angles of diffraction, and will compute the diffraction patterns (light intensities and angle of diffraction) from the range of different particle sizes within the sediment sample, producing a particle size distribution interpretation (Coulter 1990, 1994).

A Coulter LS230 laser diffraction instrument was used for grain size analysis. The LS series of instruments have 127 diffraction detectors, and use a 11lb fully protected solid state diode laser with an output of 5 mW and a wavelength of 750 nm (Coulter 1990, 1994). Within the instrument samples could be suspended in different fluid types. For this analysis sediment particles were suspended in water. The operation software version 2.09 was run through Microsoft ™ Windows.

Within conventional laser diffraction instruments the angle of diffraction for particles smaller than <0.4 μm is higher than the highest angle detector. The LS230 instruments also have a Polarisation Intensity Differential Scatter (PIDS) system that can detect particles in the size range 0.04 μm-0.4 μm. The PIDS measurements are based upon the interaction of very small particles with polarised light whose wavelength is greater than the diameter of particles (Bott and Hart 1990). Under polarised light electrons within sub-micron particles oscillate and the oscillations manifest themselves as scattered light waves which can be picked up by a second series of detectors (Bott and Hart 1990).

Particle size distribution analysis carried out on the Coulter LS230 instrument using laser diffraction, measuring grain sizes 0.4 μm-2000.0 μm, can be used in conjunction with or separate from the PIDS grain size measurements (0.04 μm-0.4 μm).
The laser diffraction method within the Coulter LS230 requires a specific range of sediment sample suspension density, defined as the 'obscuration index' which is a value of between 10-13% Total Obscuration. If the concentration of the sample suspension is too low the results are less reproducible and unreliable (see sample preparation). If the concentration is too high too much multiple scattering occurs and again the result is

unreliable (Burmann *et al.* 1997). The PIDS system requires a sediment sample suspension density of between 50-55% Obscuration to produce a reliable result. These values are monitored by the Coulter instrument and displayed on the computer monitor. Once these values have been reached, that is 10-13% Total Obscuration and 50-55% PIDS Obscuration, both particle size distribution instruments can process the sample.

Burmann *et al.* (1997) noted that with certain sediment types both particle size systems could not be used. Very clay-rich samples caused the PIDS Obscuration value to rise rapidly before the laser diffraction Total Obscuration value was high enough for a successful particle size distribution measurement. For example, with clay-rich samples the PIDS Obscuration value will reach between 50-55% but the Total Obscuration value will only be at 2-3%. If the sample is processed at these values Burmann *et al.* (1997) note that information on coarser (silts and fine sands) grain sizes is under-represented within the particle size distribution data produced. Burmann *et al.* (1997) also note that if the Total Obscuration value is increased to the correct value of between 10-13% then simultaneously for the PIDS system a value of >90% Obscuration is obtained. Under these conditions the PIDS system can no longer operate, with this fraction of grain sizes under-represented from the analysis.

Burmann *et al.* (1997) suggest that if this situation arises then an assessment must be made as to which particle size group is most valuable for the interpretation of sediment grain size. Lacustrine sediments are essentially silts and fine sands (Jones and Bowser 1978; Sly 1978) and the grain size data for these fractions was considered more important than the sub-micron (0.04 μm-0.4 μm) data generated by the PIDS system. Therefore if the above situation arose the PIDS system was switched off and only laser diffraction analysis carried out.

Particle size distributions were determined using the Coulter LS230 and the PIDS measurements at contiguous 2.0 cm sampling interval for cores LCI, LCIV and LCVII. Time constraints meant that grain size analysis was not carried out for core LCII. After grain size analysis from these three cores, specific sample depths from core LCXII were analysed. All the sediment samples were prepared for analysis using the methods outlined below.

5.3.1 LABORATORY PROCEDURES

(1) All organic material was removed from the sample. For this analysis the loss-on ignition residues were used (*cf.* Edwards and Whittington 1993). The loss-on-ignition residues used in this analysis were often 1-2g and were in some instances not sufficient to generate a reliable result (Point 8), in particular if the sample was organic-rich, >75% loss-on-ignition.

(2) Samples were prepared in clean screw-top glass vials using 1ml of dispersant (Calgon; 35.0g sodium hexametaphosphate, 7.0g sodium carbonate made up to 1litre) and 1-2 ml of distilled water. The dispersant ensured that the finer particles remained suspended in solution and flocculation of particles was minimised.

(3) The samples were placed in a sonic bath for 1 hour to ensure that particles, which had perhaps been baked during the loss-on-ignition analysis, were dispersed. Sonication also reduced the amount of flocculation within the sample (Burmann *et al.* 1997).

(4) Immediately prior to running the sample through the Coulter LS230 instrument samples were placed in the sonic bath for a further 15 minutes. This ensured that samples, which may have settled out on standing, were thoroughly mixed and particles well suspended.

(5) Samples were kept mixed by constant agitation with a micro-pipette and once fully mixed added to the water column within the COULTER LS230 instrument.

(6) Sample was added until the Total Obscuration value reached 10-13% and the PIDS Obscuration value reached 50-55%. Then the sample was run for particle size distribution analysis. Throughout the sample run period the water flow rate within the water column was kept high to keep large or dense particles suspended and prevent sample loss in the tubing.

(7) The sample was run three times during particle size analysis. The data displayed as percentage volume particle size distribution curve were calculated using the Fraunhofer optical model. This model is pre-programmed into the LS 230 software. The optical model can be altered within the software to account for differences in the refractive indices of both the particles and the suspension fluid (Burmann *et al.* 1997). Only when a sample contains a significant fraction < 10 μm is it recommended that a specific optical model be set up (Coulter 1990, 1994). The Fraunhofer model had been used successfully for soil samples containing a range of

mineral types and grain sizes (Coulter 1990, 1994) and thus the same model could be used to interpret the particle size distribution of lake sediments.

(8) Using the software the three sample runs were displayed as particle size distribution curves and were overlaid to ensure within-sample reproducibility. If the three curves did not overlay correctly the sample was considered unreliable and was run again and if still non-reproducible, considered in error.

(9) Once the sample was analysed the machine was rinsed out automatically and once clear another sample added. A high rate of water flow ensures efficient rinsing of the system and prevents sample 'carryover' (contamination of a sample by the previously run sample).

5.3.2 DATA ANALYSIS

The particle size distribution data were analysed using the percentage volume distribution curves and the raw percentage volume particle size data obtained from the Coulter LS230 software. The percentage volume data for each particle size were collated into the grain size groups using the Udden-Wentworth grain size classification scheme (Table 5.4) (Leeder 1982). These data were entered into the TILIA v.2 software (Grimm 1992), and presented and analysed as changes in grain size class with depth. Within TILIA zones that defined changes within the grain size data were visually constructed and interpreted as periods of fluctuating energy of deposition.

Mode and relative energy of deposition of lacustrine sediments was further defined through comparisons of particle size distribution curves. Average particle size distribution curves, plotted with a 2σ standard deviation curve, were generated for sediments typifying a particular mode and energy of deposition, such as low-energy lacustrine sedimentary processes. Particle size distribution curves of sediment samples from within the lacustrine sedimentary record were then compared to this average particle size distribution curve. If sample particle size distribution curves lay inside the 2σ distribution curve these sediments were considered to share a common energy and mode of deposition. Sediment grain size distributions lying outside the 2σ distribution curve were considered to have a dissimilar mode and energy of deposition. A coarser or finer grain size would indicate a higher or lower energy of deposition than sediments represented by the average distribution.

The shape of the particle size distribution curve indicated the degree of sediment sorting, with a grain size distribution curve over a narrow particle size range defined as well sorted. The degree of sorting exhibited by a sediment sample would indicate the effectiveness of the depositional medium in separating grains of different classes. Sorting of the sediment will reflect the sediment source, the grain size and the depositional mechanism (Section 3.3.2; Tucker 1991).

Sediment description	Grain size (μm)
Clays	<3.9
Very fine silt	3.9- <7.8
Fine silt	7.8- <15.6
Medium silt	15.6- <31
Coarse silt	31- <63
Very fine sand	63- <125
Fine sand	125 <250
Medium sand	250- <500
Coarse sand	500 <1000
Very coarse sand	1000 <2000

Table 5.4 Grain size classifications derived from the Udden-Wentworth classification (Leeder 1982)

5.4 HUMIFICATION ANALYSIS

Near-shore sediments at Loch Coulavie contained substantial (>2m) deposits of sedge-rich peat. The accumulation of organic material is likely to have been beneath fen or reedswamp communities (Walker 1970; Wheeler 1999).

Section 3.4.1 outlines a methodological approach whereby fluctuations in surface wetness (lake-level) control the degree humification within fen peats. It is proposed that, similarly, humification changes within the Loch Coulavie marginal sedge peat would contain a record of fluctuations in lake-level.

The colorimetric determination of peat humification provides a method to measure quantifiably the changes in the amount of peat humification (Aaby and Tauber 1975; Aaby 1986). The methodology is based on the alkali extraction of humic acids within a peat sample and the amount of light transmitted or absorbed by the humic acid-alkali solution. The more humified the peat the greater the amount of humic acids, a result of aerobic decay of the peat, the darker the solution thus a lower light transmission value. A modified version of this technique developed by Blackford and Chambers (1993) is used to determine the humification record of peat at Loch Coulavie. The absolute intensity of surface dryness or wetness cannot be directly inferred from percentage light transmission data, and as such the latter represent a semi-quantitative measure (Blackford and Chambers 1993; Charman *et al.* 1999).

5.4.1 SITE SELECTION

The sensitivity of the fen peat record to a hydrological control on peat humification is dependent on the variable depth to the water table across the fen mire surface (Section 3.4.1). Sites closer to the basin centre such as LCIV (Figures 5.1, 5.2) could have always been wetter with the peat surface only drying out (becoming increasingly humified) during very prolonged low water levels. Conversely sites nearer the shore and higher within the mire such as LCII (Figures 5.1, 5.2) would be drier, with peat only becoming wetter (more poorly humified) during very high lake levels. Site LCI (Figures 5.1, 5.2) is associated with seasonal fluctuations (up to 30 cm) in lake-level (Section 4.5). It is proposed that this site on the mire surface would be sensitive to fluctuations in the water table with variations in the degree of peat humification in response to both prolonged periods of high and low water-level. The application of this spatially variable sensitivity across the mire to differing magnitudes of water level is further discussed in Section 9.6.

5.4.2 LABORATORY PROCEDURES

Humification analysis was carried out on the sedge peat within core LCI as defined from the Troels-Smith sediment description (Section 5.2.1) and from loss-on-ignition analysis (Section 5.2.4) > 70% loss-on-ignition, thus ensuring that humification analysis was only carried out on organic sedge peat and not on minerogenic lake sediments.

Contiguous sampling allowed for the interpretation of variations in humification and rates of peat accumulation, the temporal resolution of which is controlled by the sampling interval. The sampling interval was chosen to represent a compromise between the potential temporal resolution of the humification record and a realistic number of samples to be processed within the time available. Sample preparation and analysis followed the methodology as defined by Aaby (1986) and Blackford and Chambers (1993).

1. The surface of the peat was cleaned prior to sampling by scraping horizontally with a scalpel across the peat surface.
2. The peat was sampled at contiguous 2.0 cm intervals using a clean stainless steel cork borer of diameter 2.0 cm and placed in clean dry crucibles. For each sample the presence of woody fragments and vegetation type was recorded.
3. Samples were oven dried overnight at 40°C (minimum of 16 hours).
4. The dried peat was ground with a pestle and mortar.
5. 100ml of freshly made 8.0 % sodium hydroxide solution (NaOH) was added to 0.2g of dried peat in a 250 ml beaker and once brought to the boil, on a hot plate simmered for one hour. The time of initial mixing was noted.
6. Once cooled the peat solution was poured into a 200 ml volumetric flask and made up to the mark using distilled water and shaken thoroughly.
7. The peat solution was filtered through No.1 Watmann paper into a 250 ml conical flask. 50.0 ml of filtered solution was added to a 100ml volumetric flask and made up to the mark using distilled water and shaken thoroughly.
8. Immediately after shaking this 1:1 solution was added to clean colorimeter cuvettes using a micropipette. Prior to each sample reading the Jenway 6061 colorimeter was calibrated using distilled water for 100% transmission and a black cuvette for 0% transmission. The cuvettes containing the sample were placed in the colorimeter and a percentage transmission of light at wavelength 540 ηm recorded. Three replicates were taken for each sample; within-sample variation was no greater than 0.5%. If greater then more replicates samples were taken. The mean of the replicates was calculated as the percentage transmission value for the sample.

For all samples the time from initial mixing to the percentage light transmission readings took less than four hours. This ensured that percentage transmission values were not affected by fading of the humic acid solution (Blackford and Chambers 1993).

Blackford and Chambers (1993) suggest that owing to the variety of humic acids produced by different peats under different conditions a direct translation of percentage transmission values to percentage humification values, as outlined by Aaby (1986), is questionable. The authors propose that all analysis of changes in humification and inferences on mire surface wetness at the time of deposition are made from percentage light transmission data.

5.4.3 DATA ANALYSIS.

It is assumed that changes in surface wetness, particularly lake-level, will control the degree of sedge peat humification (Section 3.4.1). However, Blackford and Chambers (1993) outline further controls on the degree of peat humification as indicated by colorimetric analysis, which are as follows:

(i) vegetation changes within the mire surface. Differing peat-forming vegetation types will breakdown under aerobic conditions at varying rates and produce variations in humification that have no surface wetness control.

(ii) the amount of minerogenic sediment. Within the peat sample analysed the proportion of peat and minerogenic sediment will control the humic acid content in the sample solution. For example a sample which contains a high percentage of minerogenic sediment will appear to have on analysis a lower humification value as much of the sample is made up minerogenic sediment that does not generate humic acids upon alkali extraction (Chambers 1984).

These controls on the value of the degree of humification are considered to be just as applicable to the peats noted in Loch Coulavie. Thus humification data generated were tested against variations in vegetation type as noted through Troels-Smith notation (Section 5.2.1) and the amount of minerogenic sediment as determined through loss-on-ignition analysis (Section 5.2.4).

A further control on the degree of humification within the peat profile is the potential for continued decay (increasing humification) with depth. Within ombrotrophic mire systems the greatest amount of decay takes place in the acrotelm. However, Clymo (1984, 1991) proposes that decay does take place within the deeper catotelm but at a much slower rate. The sedge peat beneath the acrotelm within Loch Coulavie is accumulating under essentially permanently waterlogged conditions similar to the catotelm: thus it is thought that decay at depth could occur. Fitting an exponential regression curve to the sedge peat percentage light transmission data points will allow for an assessment of the amount of decay at depth (*cf.* Anderson 1996, 1998; Blackford and Chambers 1995). If increased amounts of decay at depth are indicated the data set is then detrended. The equation of the slope of the regression line is used to generate residual values. These residual values are plotted graphically and represent data detrended for decay at depth. This potential decay at depth within the humification data set is further discussed in Chapter 8.

Once these controls on the degree humification have been identified within the percentage transmission data set the data must be defined as a series of significant shifts. These significant shifts can then be interpreted in terms of actual shifts in peat humification and thus in surface wetness.

Within the literature the identification of significant shifts within the percentage light transmission (humification) record is poorly defined. Many authors rely on the generation of positive and negative residual values from decay at depth and exponential regression curve analysis to identify 'significant' shifts (see above) (*cf.* Anderson 1998; Anderson *et al.* 1998; Blackford and Chambers 1995; Mauquoy 1997; Mauquoy and Barber 1999a). However, where no decay at depth occurs within the data set significant shifts are identified by often subjective observations on changes in the percentage light transmission curve (Blackford and Chambers 1991; Chambers *et al.* 1997). Blackford (1990) defines major changes in surface wetness as 'those that mark points beyond which the curve (percentage light transmission curve) does not return', and minor changes as 'curve of a less prominent nature but still detected by 3 or more readings'. For analysis of the percentage light transmission data from Loch Coulavie this approach was considered unsatisfactory and a more objective methodological approach was used. A mean and standard deviation to 1σ was calculated from the percentage light transmission data controlled by shifts in surface wetness (as outlined above). Fluctuations in percentage light transmission values above and below the 1σ value were considered as significant shifts to decreased and increased humification respectively. The start of a shift in humification was defined as the point from which percentage light transmission values started to increase or decrease. An inferred chronology (Section 5.6) of shifts in humification was based on interpolations of peat depths associated with the start of shifts, as defined above.

5.5 POLLEN ANALYSIS

The application of changes in aquatic vegetation as a methodology for identification of lake-level fluctuations is discussed in Section 3.3.1. Changes in aquatic vegetation are generally determined through the analysis of plant macrofossil remains (Digerfeldt 1986, 1997; Hannon and Galliard 1997). The interpretation of macrofossil analysis is based on the assumption that these plant remains are *in-situ* and thus are indicative of the water depth at the time of deposition (Section 3.3.1).

Macrofossil analysis of the marginal peats and deeper water lake sediments at Loch Coulavie was assessed as follows:

- a 6.0 cm thick (*c.* 50 g in weight) slice from the sediment core was washed through a 0.5 mm sieve. The organic material remaining was well preserved. Under a low powered (magnification x10) binocular microscope samples were noted to contain the fruits and seeds of aquatic species, *Juncus* type, and non-aquatic species such as *Betula* and *Pinus*. However, it was estimated that to obtain a statistically viable count of aquatic macrofossils would require up to 200-300 g of sediment. Within the cores a minimum sampling interval of 20.0 cm was therefore required, probably too coarse a resolution to identify shifts in aquatic vegetation. Within the deeper water sample, fewer plant macrofossils were found and for a statistically viable macrofossil analysis sediment of greater than 200-300 g would be needed.

Pollen analysis was considered as an alternative method to determine changes in aquatic vegetation. There are several concerns about this application of pollen analysis:

(1) Davis *et al.* (1971), Harrison and Digerfeldt (1993) and Walker (1970) considered aquatic pollen to be deposited close to the parent plant (Section 3.3.1). However, there is potential for pollen grains, more so than plant macrofossils, to be reworked and redistributed away from the parent plant within the lake (Peck 1973).

(2) Many species such as sedges and grasses, which would have dominated marginal vegetation, cannot be identified to species level using pollen.

(3) The limitations of obtaining quantifiable changes in water depth from aquatic plant taxa, discussed in Section 3.3.1, become even more apparent with the identification from the pollen record of a more restricted number of aquatic species.

(4) Pollen concentrations of aquatic pollen are often low, with key (Table 5.5) aquatic species such as *Typha* producing low amounts of pollen (*cf.* Bunting and Warner 1998).

5.5.1 SITE SELECTION

Section 5.4.1 discussed how the position of LCI on the fen mire ensured that this site was sensitive to both falls and rises in lake-level. Palynological data from fen peat within this hydrologically sensitive site should therefore contain a record of changes in aquatic vegetation associated with fluctuations in lake-level.

A palynological record of aquatic vegetation changes associated with a potentially deeper water environment was obtained from core LCXII.

5.5.2 LABORATORY PROCEDURES

Within core LCI the sampling resolution was based on proposed lake-level fluctuations identified within the fen peat from previous analysis, loss-on-ignition (Section 5.2.4) and humification (Section 5.4). A sampling interval of 50.0 cm was used to obtain a palynological record for the full sediment record in core LCXII.

The sampling technique and preparation of pollen samples followed the methodology laid out in Moore *et al.* (1991).

(1) Around 1cm^3 of sediment was sampled from a cleaned section of core, the volume of sediment determined using water displacement (Bonny 1971)

(2) *Lycopodium* tablets were added at the earliest stage for use in pollen concentration calculations (Stockmarr 1971). 1-2 ml of 10% Hydrochloric acid (HCL) was added to dissolve fully the *Lycopodium* tablets and remove potential carbonate minerals.

(3) Highly organic sedge peat samples were boiled in 4.0 ml 10% Potassium hydroxide (KOH) for 15-20 minutes to dissolve humic material, rinsed in distilled water and washed through a 150 µm sieve to remove coarse organic fragments. Cellulose was removed using Erdtman's acetlation. 4-5 ml of a fresh mixture of 9:1 acetic anhydride and concentrated sulphuric acid was added to the pollen sample, pre-washed with glacial acetic acid, and heated in a boiling water bath for 2-3 minutes. The pollen sample was stained after acetlation using a few drops of aqueous safranin solution washed through with 10% KOH solution. Samples were dehydrated using a wash of 1 ml of tertiary butyl alcohol (TBA) prior to the addition of a small amount of silicon oil. Samples were placed in a warm drying cabinet to ensure that all the TBA had evaporated.

(4) The treatment of more minerogenic sediment samples initially followed the procedure outlined above, until after treatment with hot KOH. To remove larger minerogenic particles, samples were washed through a 150 µm and a 10 µm sieve. Sediment retained on the 10 µm sieve was treated with 2.0 ml of 10% Hydrofluoric acid (HF) and placed in a hot water bath for 30 minutes, samples being stirred on an auto-stirrer every 10 minutes. Any remaining minerogenic material was removed with a further hot HF treatment. Samples were subsequently treated to Erdtman's acetlation, staining and silicon mounting as outlined above.

Microscopic slides were made up using a small amount of the residue mounted in silicone oil and a cover slip placed over the sample. Pollen grains were counted using a LEICA DMLS binocular microscope at x400 magnification (x10 eye piece and x40 objective lens). Pollen counts were made using a series of systematic unidirectional traverses across the full width of the slide. Counts made right across the slide ensured that any potential sorting of pollen grains during slide preparation did not bias the data set. Pollen grains were identified using the key provided in Moore *et al.* (1991) and a type slide reference collection.

A total land pollen sum (TLP) of 300 pollen grains was counted initially. A local pollen sum was then constructed to reflect lacustrine and lake-shore environments, with taxa such as Cyperaceae and *Calluna vulgaris*, and a total of 300 pollen grains counted. Pollen from tree and shrub taxa considered to represent a regional vegetation pollen signal such as *Pinus sylvestris* were excluded from the local pollen sum. Table 5.5 lists taxa included in both regional and local pollen sums. Within the local pollen sum water depths associated with aquatic species were determined from Spence (1964) and Haslam *et al.* (1975) (Table 5.5).

Total Land Pollen Sum	Local Pollen Sum
Trees-*Alnus glutinosa*	**Trees**- *Alnus glutinosa*
Betula	**Shrubs**-*Salix*
Fraxinus	**Heaths**-*Calluna vulgaris*
Pinus sylvestris	*Erica*
Quercus	*Vaccinium* type
Sorbus	
	Herbs-Cyperaceae
Shrubs-*Corylus* type	Gramineae
Salix	*Fillipendula* sp
	Galium sp.
Heaths-*Calluna vulgaris*	*Plantago lanceolata*
Erica	*Potentillia* type
Vaccinium type.	Rosaceae
	Ranunculus sp.
	Succisa sp.
Herbs-Cyperaceae	Umbelliferae
Gramineae	*Urtica* type.
Fillipendula sp.	
Galium sp.	**Spores**-*Dryopteris*
Plantago lanceolata	*Equisetum* (40-150 cm)
Potentillia type	*Polypodium*
Ranunculus sp.	*Pteridium*
Rosaceae	*Sphagnum*
Succisa sp.	
Umbelliferae	**Aquatics**- *Isoetes* sp. (2-250 cm)
Urtica type	*Pediastrum*
	Potamogeton sp. (12-260 cm)
	Menyanthes sp. (3-43 cm)
	Myriophyllum sp. (40-250 cm)
	Nuphar lutea (20-150 cm)
	Nymphaea alba (20-195 cm)
	Typha latifolia (0-60 cm)

Table 5.5 Plant species counted in the Total and Local Pollen sums (see text), potential water depths for aquatic species are also given.

The data on regional vegetation change produced from the TLP sum were used as a chronological control. The radiocarbon chronology developed for Loch Coulavie

(Section 5.6) was tested from this pollen stratigraphy and from the radiocarbon-dated pollen record generated within the Loch Coulavie region (Davies 1999; Section 4.7).

During pollen analysis an assessment of the amount of pollen corrosion was made. *Betula* and *Corylus* pollen grains exhibit very clearly and distinctively any effects of corrosion through aerobic bacterial activity (Havinga 1984). During construction of the TLP sum 50 grains of *Betula* and *Corylus* were recorded as corroded (> 25% corroded surface) or non-corroded. The data were converted into percentages and presented as a bar graph.

Pollen counts were entered into TILIA v.2 (Grimm 1992) and pollen percentage diagrams for local pollen sums, total pollen sums and pollen concentration constructed using TILIA GRAPH. Pollen zones within the percentage pollen diagrams were, based on a visual interpretation of vegetation changes.

5.6 CHRONOLOGY CONSTRUCTION

To establish synchroneity of stratigraphic events within the Loch Coulavie basin and with regional events a chronological framework was generated from radiocarbon dates. Radiocarbon dates were obtained on stratigraphic events identified within each sediment core (LCII, LCI, LCIV LCXII and LCVII) through sediment description (Section 5.2), loss-on-ignition (Section 5.2.4) and grain size analysis (Section 5.3). Radiocarbon dating was carried out using the Accelerator Mass Spectrometry (AMS) technique for the following reasons:

(1) the very small sample size required for AMS dating, 1 mg of organic carbon as compared to 5-10 g of organic carbon for conventional dating. The small sample size needed for AMS dating improves the dating resolution within the sedimentary sequence. Within Loch Coulavie radiocarbon dating controls were needed to define events within the stratigraphy with AMS dating of sediment lying immediately (within 1.0 cm) above an event. The AMS dating of a thin sediment slice ensures that the amount of time taken to accumulate that slice of sediment is less than a much thicker slice needed for conventional radiocarbon dating. This is particularly

important within lacustrine environments where sediment accumulation rates can be highly variable (Lowe 1991).

(2) The small sample size required for AMS dating allows for the direct correlation with sedimentological and palynological data generated from the sample cores.

5.6.1 SAMPLING PROCEDURES

Meticulous sampling procedures further secured the reliability of the radiocarbon ages obtained. These were as follows:

(1) the use of a closed-chambered Russian corer to sample the lake sediment.

(2) wrapped cores were stored for a restricted time prior to radiocarbon assay in a cold and dark environment reducing the amount of bacterial activity and fungal growth (Wohlfarth *et al.* 1998).

(3) potential points of sediment reworking such as directly below minerogenic inwashing events evident within the core stratigraphy were not sampled (Lowe 1991; Harkness and Lowe 1991). This avoidance during radiocarbon sub-sampling of reworked organic material within the lacustrine record is further discussed in Sections 6.2.8 and 6.4.5.

(4) samples of greater 25 % loss-on-ignition and 2.0 g in weight were submitted to ensure a sufficient amount of organic carbon was available for radiocarbon dating.

(5) the surface of the core was scraped clean with a clean stainless steel scalpel prior to sampling. A thin sample (<1 cm thick) was taken using a clean stainless steel scalpel to ensure the highest dating resolution within the sediment stratigraphy.

(6) within the sample any obvious living penetrating rootlet material within the sample was removed carefully using clean stainless steel tweezers.

5.6.2 LABORATORY PROCEDURES

AMS radiocarbon assays were obtained from the insoluble humin fraction. Standard radiocarbon laboratory sample pre-treatment procedures remove the more soluble humic fraction and also more obvious carbon contaminants. Rootlets and foreign matter are removed from the sample. Samples are digested in 2M HCL (Hydrochloric acid) (80°C for 10 hours) to remove carbonates in the sediment matrix. Samples are washed free of acid using distilled water and further treated with a digestion in 2M KOH (Potassium hydroxide) (80°C for 10 hours) to remove soluble humic acids. The digestion was repeated using distilled water until no further humics were extracted. The residue (humin content) was rinsed free of the alkali, digested in 1M HCL (80°C for 5 hours) and then rinsed free of acid, dried and homogenised. The total carbon as a known weight of the pre-treated sample was recovered as CO_2 (Carbon dioxide gas) by heating with CuO (Copper oxide) in a sealed quartz tube. The gas was recovered to graphite by Fe/Zn (iron/zinc) reduction. The age of the sample is obtained by comparing the $^{14}C/^{12}C$ ratio measured within the graphite sample as compared to the ratio within a material of known ^{14}C activity. The number of ^{14}C atoms within the graphite sample was counted using the principal of accelerated mass spectrometry.

The radiocarbon assay is reported from the laboratory as an age in radiocarbon years before present (BP). A plus and minus value on the radiocarbon age reflects statistical errors generated during the measurement process. The radiocarbon age was calibrated to calendar years BP (cal. BP), to correct for temporal variations in ^{14}C production through the Holocene (Lowe and Walker 1997), using the CALIB V.4.1.2 software (Stuiver and Reimer 1999). In Chapter 6 the results of the radiocarbon assays are presented as radiocarbon years before present (BP) with the statistical errors alongside the fully calibrated ages (cal. BP) at 1σ and 2σ ranges. During discussion and interpretation of the stratigraphic event chronology (Chapter 7) both radiocarbon ages (BP) and the midpoint of the calibrated age range (cal. BP) are quoted. Radiocarbon (BP) and calibrated (cal. BP) ages of events are interpolated from the AMS chronology; here an error of ± 150-200 radiocarbon years is assumed, with all ages generated described as approximate or *circa* (*c.*).

Table 5.6 summarises pollen zone chronology and regional vegetation changes as defined by Davies (1999) from a site Torran Beithe within the Loch Coulavie catchment (Section 4.7, Figure 4.5) used where applicable to test the internal consistency of radiocarbon chronologies obtained (Sections 6.2.8, 6.4.5).

Pollen zone	^{14}C Years BP	Calibrated Years BP	Vegetation change events.
Cyperaceae-Poaceae-*Calluna vulgaris*-*Pinus*	0-190	0-150 (cal. AD c. 1800-0)	Evidence for modern forestry species from c.70 BP (modern cal. BP).
Calluna vulgaris-Cyperaceae-Sphagnum	190-1290	150-1200	Cereal-type 410 BP(cal. AD1430-1630), Reduction in *Calluna* between 1320-1250 BP (1270-1170 cal. BP).
Calluna vulgaris-Cyperaceae	1290-3600	1200-3890	Permanent drop in *Pinus* from 3600 BP (3890 cal. BP), spread of blanket peat from c. 3600 BP (3890 cal. BP). *Calluna* dominant by 3200 BP(3390 cal. BP)
Pinus sylvestris-Betula-*Calluna vulgaris*	3600-6360	3890-7230	*Alnus* maximum at 5630 BP (6410 cal. BP), *Ulmus* maximum followed by decline at 5240 BP (5970 cal. BP), *Quercus* peak at 5410 BP (6250 cal. BP). Reductions in *Pinus* pollen from 5570-4960 BP (6360-5690 cal BP) and 4620-4390 BP (5310-4930 cal. BP), maximum mid Holocene values between 4390-3765 BP (4930-4110 cal. BP) with a subsequent abrupt fall in *Pinus* from 3765 BP (4110 cal. BP)
Pinus sylvestris-*Betula*-Cyperaceae	6360-6720	7230-7540	*Pinus* well established by 6550 BP (7390 cal. BP).
Betula-Pinus-Calluna vulgaris-Corylus avellana-type	6720-7890	7540-8590	Transition in woodland communities around 7890 BP (8590 cal. BP) increase in *Pinus*, *Quercus* and *Ulmus* with declines in *Betula* from around 6720 BP (7540 cal. BP).
Betula-Corylus avellana-type	7890-8800	8590-9740	*Corylus* expansion c. 8710-8300 BP (9780-9310 cal. BP) with a decline at 8300 BP (9310 cal. BP).

Table 5.6 Torran Beithe Pollen zone chronology and regional vegetation changes as defined by Davies (1999).

PART 2 LAKE-LEVEL FLUCTUATIONS

CHAPTER 6
LACUSTRINE SEDIMENT ANALYSIS: RESULTS

6.1 INTRODUCTION

This chapter presents the results of sedimentological analyses discussed in Chapter 5. Table 6.1 summarises analysis techniques applied to each core. Discussion within Chapter 5 outlines why some techniques were not applied to all cores. Results for each sediment core across the Loch Coulavie transect are presented as independent data-sets with data for the near-shore core presented first (Figures 5.1, 5.2). Only through treating each core as an independent data-set is the complexity of sediment infilling within the Loch Coulavie basin, with differing sedimentary processes and sediment accumulation rates for each core, fully defined.

Technique	**Core**				
	LCII	**LCI**	**LCIV**	**LCXII**	**LCVII**
Sediment description (Section 5.2.1)	6.2.1	6.3.1	6.4.1	6.5.1	6.6.1
X-ray analysis (Section 5.2.2)	6.2.2	6.3.2			
Magnetic susceptibility (Section 5.2.3)		6.3.3			
Bulk density (Section 5.2.4)		6.3.4	6.4.2		6.6.2
Water content/Loss-on-ignition (Section 5.2.4)	6.2.3	6.3.3	6.4.2	6.5.2	6.6.2
Grain-size analysis (Section 5.3)		6.3.5	6.4.3	6.5.3	6.6.3
Humification (Section 5.4)		6.3.6			
On-site pollen analysis (Section 5.5)		6.3.7		6.5.4	
Chronology (Section 5.6)	6.2.4	6.3.8	6.4.4	6.5.5	6.6.4

Table 6.1 Loch Coulavie sediment cores; summary of analysis techniques with reference to discussion of techniques, sections in Chapter 5 and presentation of results, sections in Chapter 6.

Within each core the results of sedimentary analysis are presented as series of 'stratigraphic events'. These events defined by loss-on-ignition, sediment description and/or grain size represent rapid and/or short-term fluctuations within the sedimentary record (*cf.* Walker *et al.* 1999; Whittaker *et al.* 1991). It also recognised that these 'stratigraphic events' are taking place against a background of 'normal' lacustrine sedimentary processes. Within this chapter, lacustrine sedimentary processes and the 'stratigraphic events' are presented as sedimentological descriptions based on the results of analysis techniques (Table 6.1). AMS radiocarbon dates were obtained to define a chronology for some but not all of these events. Where obtained these dates are presented as part of the core data-set and tested for reliability. A full chronology for all the events outlined in this chapter is achieved only after basin-wide correlation of 'stratigraphic events', interpolated ages being generated in Chapter 7 and developed into a chrono-stratigraphic framework. This permits a full interpretation of lacustrine sedimentary processes and the mode of deposition of 'stratigraphic events' in Chapter 7.

6.2 CORE LCII (Figure 5.1)

6.2.1 SEDIMENT DESCRIPTION

Basal sediments consist of laminated minerogenic silty sands (Ag2 Gmin2), with 1-2 mm thick laminae of better-sorted pale grey silts and darker grey fine sands (Table 6.2). These laminated minerogenic sediments are overlain abruptly (transition <1 mm) by organic-rich mud at 279 cm which gradually becomes increasingly organic and detritus-rich, culminating from 142 cm in the accumulation of a sedge peat. Flattened graminoid stems are recorded within the organic rich lake mud (211-142 cm).

	LCII
Depth cm	Sediment description
0-31	Brown poorly humified very fibrous sedge peat; Th4, Nig2, Strf0, Elas3, Sicc2, Lim1-2.
31-142	Humified dark brown sedge peat, living roots down to 60 cm, with herbaceous organic detritus and woody fragments; Th2 Dh1 Ld1 Dl+, Nig3, Strf0, Elas2, Sicc2, Lim1-2.
142-211	Dark brown organic detrital mud, graminoid type stems and woody fragments; Ld2 Dh2 Dl+, Nig3, Strf0, Elas0, Sicc2, Lim2.
211-235	Dark brown detritus-rich organic mud, woody and herbaceous plant fragments, mica flakes visible; Dh3 Ld1 Dl+ Ag+, Nig3, Strf0, Elas1, Sicc2, Lim2.
235-279	Brown organic-rich silt within this unit blocks of metallic grey sediments as noted below (Ag2Ga2) (235-237 cm); mica flakes visible with pieces of gravel (5mm) Ag2 Ld1 Dh1 Gmaj+, Nig2, Strf1, Elas0, Sicc2, Lim3
279-285	Silver grey coarse silts and sands, mica flakes visible, Ag2 Gmin2, Nig1-2, Strf4, Elas0, Sicc2.

Table 6.2 Core LCII sediment description using a modified version of Troels-Smith notation (Section 5.2.1).

The sedge peat shows variations in degree of decomposition and organic detritus content (Dh), with the top 31 cm noted a highly fibrous sedge peat (Th).

6.2.2 X-RAY ANALYSIS

X-ray analysis was carried out for the whole core. The organic-rich minerogenic sediments between 285-185 cm were noted to have a much more complex stratigraphy than had been suggested by sediment description; these sediments are discussed in the following section. Lighter (more minerogenic) and darker (more organic) sedimentary units are traced in Figure 6.1 to highlight bedding structures within these sediments. Individual grains of mica and quartz were noted throughout the core. The highly minerogenic basal sediments (285-279 cm) are shown as one single unit (A). These sediments were highly reflective and it was difficult to discern structure.

Above this unit are two horizontal minerogenic bands at 278 and 273 cm (B and C): both units have diffuse boundaries (>1 cm). Between 273 and 255 cm (D) the bedding structure is highly complicated with alternating organic-rich and minerogenic sediment layers, 0.3-0.7 cm thick, appearing convoluted and slumped. These alternating layers appear to have abrupt boundaries (<1mm), with other sedimentary structures such as graded bedding difficult to define due to sediment disturbance. Slumping of these sediments is coherent with bedding structure retained, suggesting limited downslope movement. Between 255 and 240 cm sediment structure appears less convoluted but there is a degree of slumping with sediments appearing stretched and discontinuous downslope (E). The highly minerogenic sediment inclusion noted at 237-235 cm (Table 6.2) is confirmed as a separate unit (F) within more organic-rich sediments (G). Above 235 cm sediment appears less stratified although, as suggested by the texture and number of mineral grains, it is still minerogenic (G). A distinct and well-defined narrow (2-3 mm) minerogenic unit is noted at 223 cm (H). Unlike those lower in the core, this unit is sub-horizontal. From the x-ray analysis it is difficult to discern if sediments below unit (H) were deformed during one or several slumping events. Above unit (H) sediment is less minerogenic and the texture changes to become more amorphous, suggesting increasingly organic sediment deposition (I) which from X-ray analysis appears to continue without interruption.

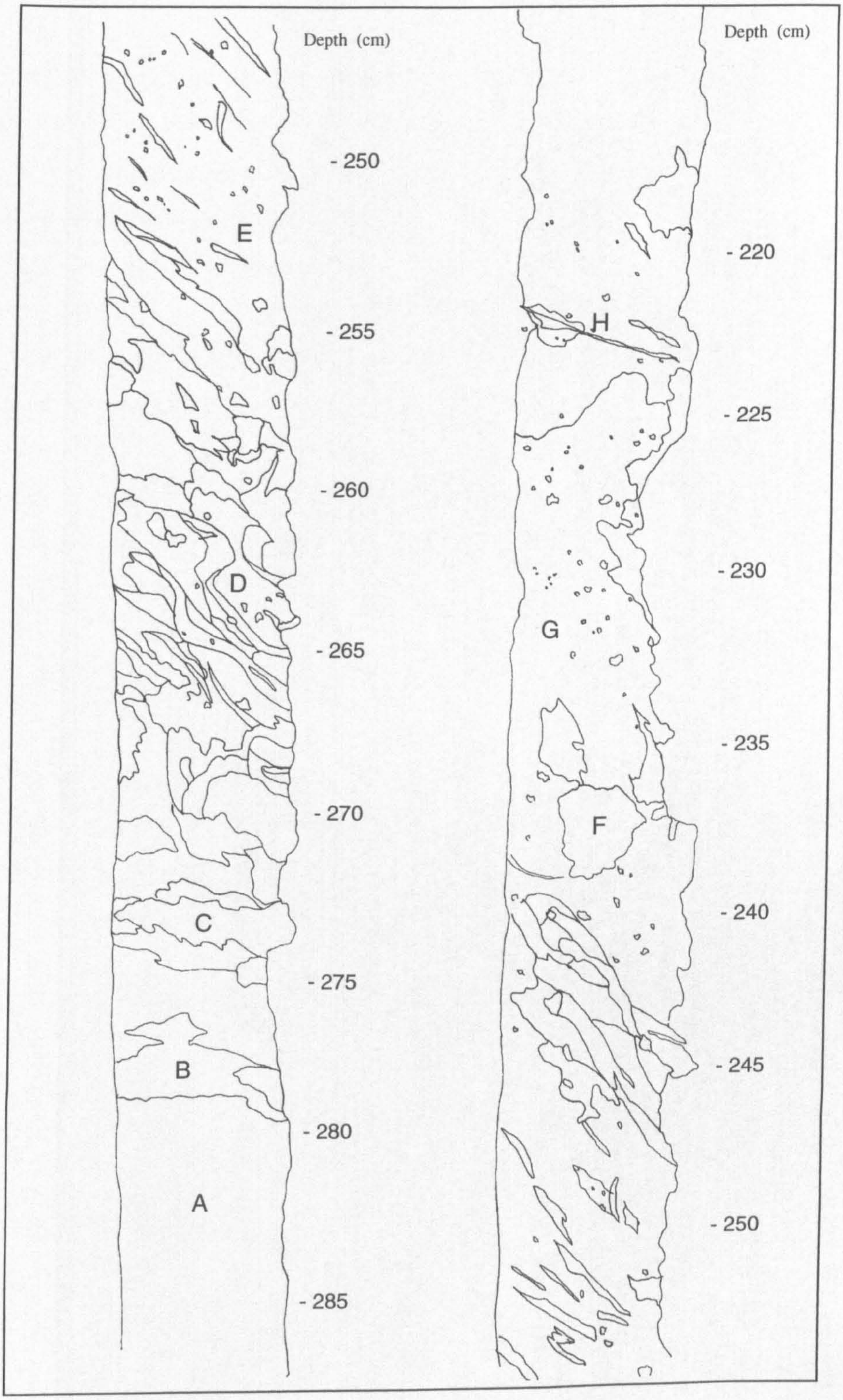


Figure 6.1 Core LCII x-ray 285-220 cm, Section 6.2.2

6.2.3 WATER CONTENT AND LOSS-ON-IGNITION ANALYSIS

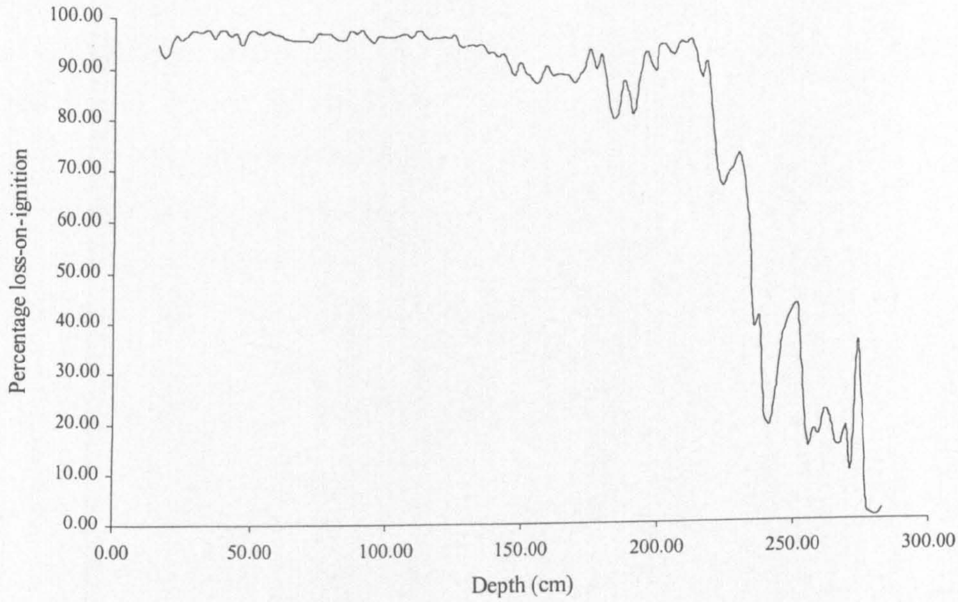
Variations in water content are related to organic content of the sediment (Figure 6.2a). Basal sediments are highly compacted and the high minerogenic content reduces porosity and water content of the sediment. Peats are less compacted and with the accumulation of highly organic peat, water content increases and varies little.
Basal minerogenic sediments (285-277 cm) have very low organic contents with a loss-on-ignition value of < 3% (Figure 6.2b). Sediment becomes rapidly more organic with loss-on-ignition values increasing over 2 cm to 19% at 275 cm. Between 275-235 cm loss-on-ignition values are highly variable (10-43%), reflecting the deposition of both organic and minerogenic rich sediments. A sustained increase in loss-on-ignition value from <38% at 235cm to >68% at 229 cm, culminates in the deposition of highly organic peat from 219 cm (> 80% loss-on-ignition), with little variation above this (between 80-97%).

6.2.4 CHRONOLOGY.

AMS radiocarbon dates define the start of (a) organic-rich mud sedimentation (275 cm) and (b) of sedge peat accumulation (219 cm) (Table 6.3). The sample at 275 cm was taken from undisturbed sediment immediately below disturbed sediments between 273-255 cm (Section 6.1.2).

Laboratory reference	Depth (cm)	C content (% by wt)	Conventional Radiocarbon age (yrs BP ± 1σ)	Calibrated age range to 1σ (yrs. BP)	Calibrated age range to 2σ (yrs. BP).
AA34225	275	1.6	10710 ± 70	12926-12641	12974-12419
AA34226	219	30	6980 ± 55	7919-7737	7937-7674

Table 6.3 Radiocarbon dates from core LCII.



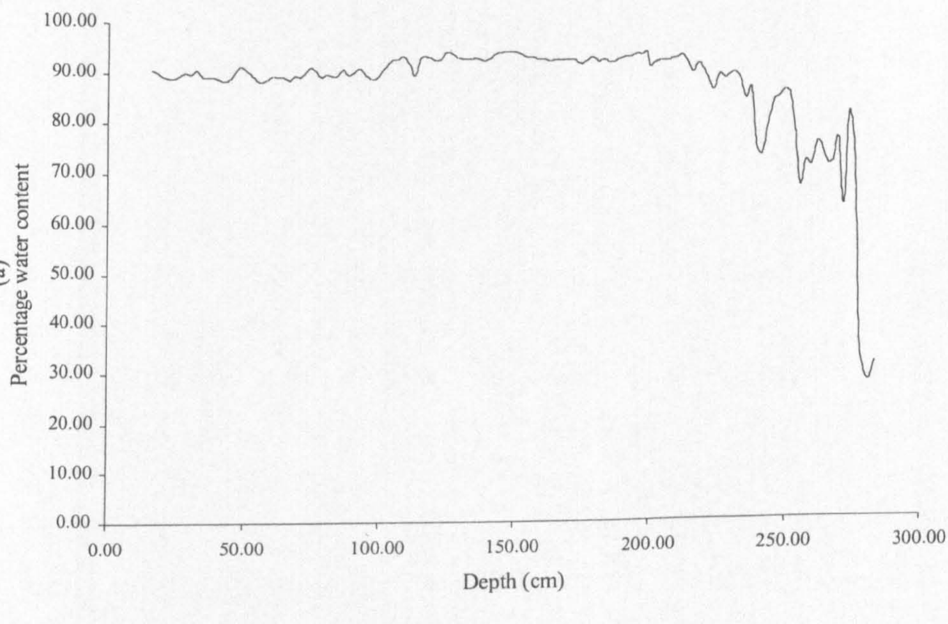


Figure 6.2 Core LCII percentage water content (a) and loss-on-ignition (b) plotted against depth

Within core LCII the radiocarbon date for the start of organic sediment accumulation, *c.* 10 710 BP (12 850 cal. BP) is considered too old, because at this date the basin is likely to have been beneath Loch Lomond Readvance ice (Section 2.3). The error in dating is unexplained. The carbon content (% by weight) value for this sample is very low (Table 6.3) and thus contamination with only a small amount of older carbon would affect the radiocarbon date (Lowe 1991). An interpolated date for the start of organic sedimentation at core LCII is estimated using radiocarbon ages obtained for a similar stratigraphic event within the remainder of the transect (Section 7.3).

6.2.5 SUMMARY OF STRATIGRAPHIC EVENTS

- Event LCII 1: deposition of laminated minerogenic silts and sands between 285-279 cm (Section 7.2).
- Event LCII 2: onset of organic sedimentation with an abrupt increase over 2 cm from <3% loss-on-ignition to >19% at 275 cm (Section 7.3).
- Events LCII 3 and 4: deposition at 278 cm and 273 cm of sedimentologically similar minerogenic units minerogenic with diffuse upper and lower boundaries (Section 7.4).
- Event LCII 5: deposition between 273-240 cm of highly stratified thin (0.3-0.7 cm thick) alternating organic and minerogenic rich units (Section 7.4).
- Event LCII 6: deposition at 237-235 cm of a block of highly minerogenic sediment similar to sediment recorded at Event LCII 1 (Section 7.4).
- Event LCII 7: sediment disturbance causing coherent slumping and overfolding of sediments recorded at events LCII 5 and 6 (Section 7.4).
- Event LCII 8: the onset of sedge peat accumulation after a sustained increase in loss-on-ignition values from<38% at 235cm to >68% at 229 cm. Sedge peat (>80 % loss-on-ignition) accumulates from 219 cm, with a radiocarbon age of 6980 ± 55 BP (7765 cal. BP). Sediment description suggests that during the accumulation of this consistently highly organic sedge peat there were changes in the degree of peat humification, but these are not well defined (Section 7.5).

6.3 CORE LCI (Figure 5.1)

6.3.1 SEDIMENT STRATIGRAPHY

Basal sediments (445-432 cm) consist of laminated (1-2 mm) highly minerogenic silts and sands, with laminations recorded as better sorted pale grey silts and darker grey sands (Table 6.4). An abrupt (over <1mm) deposition of organic-rich lake mud at 432 cm becomes increasingly detritus rich with the accumulation from 308 cm of a sedge peat. Prior to peat accumulation distinctive flattened graminoid type plant macrofossils are recorded within the lake mud (337-308 cm). The peat is predominantly composed of herbaceous and sedge type plant material with frequent large (>2 mm) woody fragments (Dl). Within the peat there are variations in the degree of decomposition as reflected by changes in the amount of coarse detrital (Dh) and finely decomposed organic material (Ld). Variations in plant material within the peat are recorded, such as at 235-229 cm and 219-215 cm where peat is composed of predominantly sedge type fragments.

Basal laminated minerogenic sediments were noted to be highly thixotropic and could not be retained in stratigraphic sequence, particularly near the base of the sequence (444-438 cm). Thus sediment analysis for this sedimentary unit is incomplete.

6.3.2 X-RAY ANALYSIS

X-ray analysis was carried out on the whole of core LCI. Sediment stratigraphy between 435-380 cm was shown to be more complex than was suggested by sediment description and is discussed in the following section. Bedding structures within sediment between 435-380 cm are shown in Figure 6.3a, a tracing of lighter (more minerogenic) and darker (more organic) sediment units. Highly minerogenic basal sediments (435-432 cm) are shown as one single unit (A). The contact between these highly minerogenic sediments and more organic sediments above (B) is shown as sharp (<1mm) and irregular. The irregular nature of the contact may in part be due to the thixotropic nature of the sediment (Section 6.3.1) and disturbance during transport. In the field this contact was noted to be sub-horizontal. The unit C running along the edge of the core is as a result of sediment disturbance.

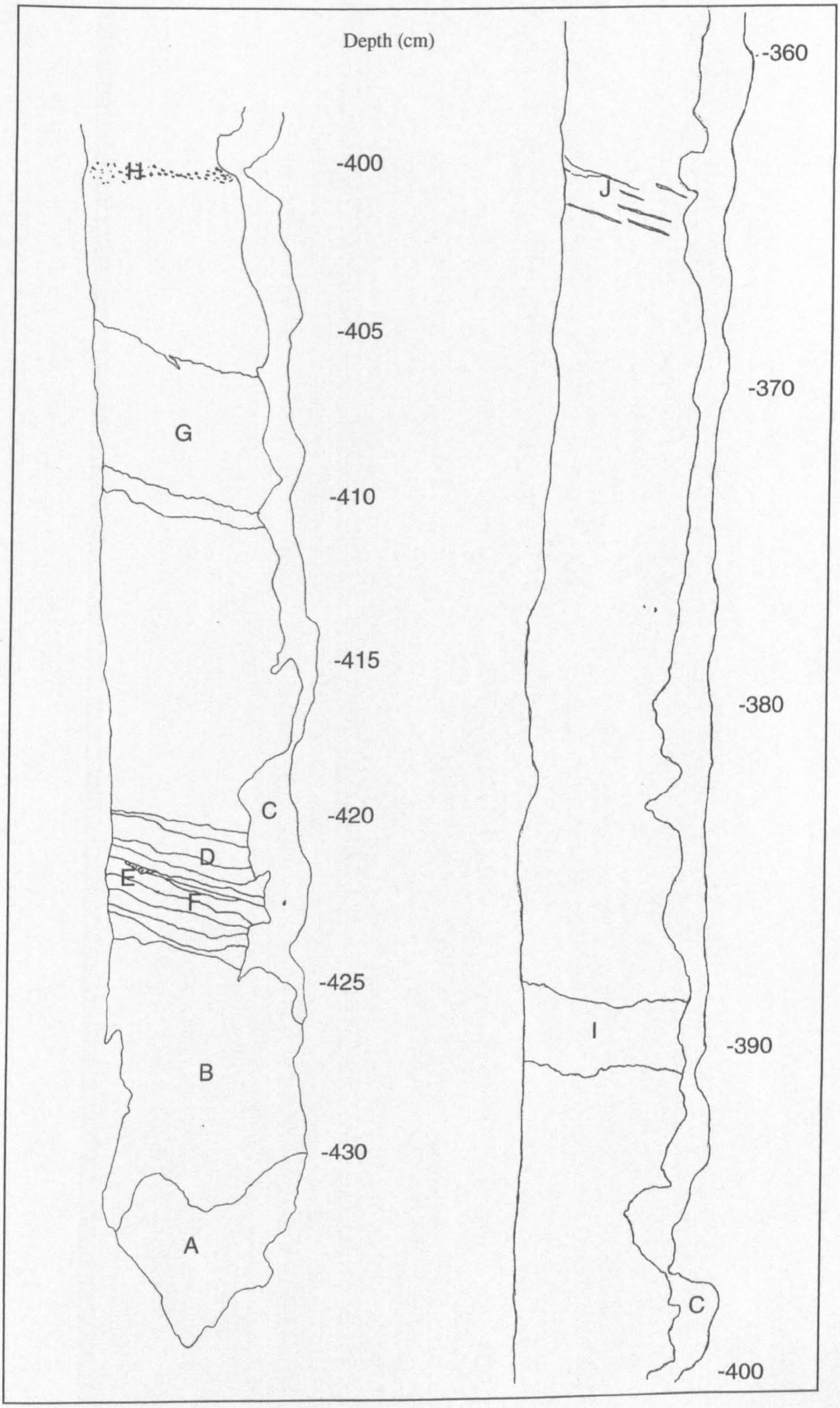


Figure 6.3a Core LCI X-ray 435-360 cm, Section 6.3.2

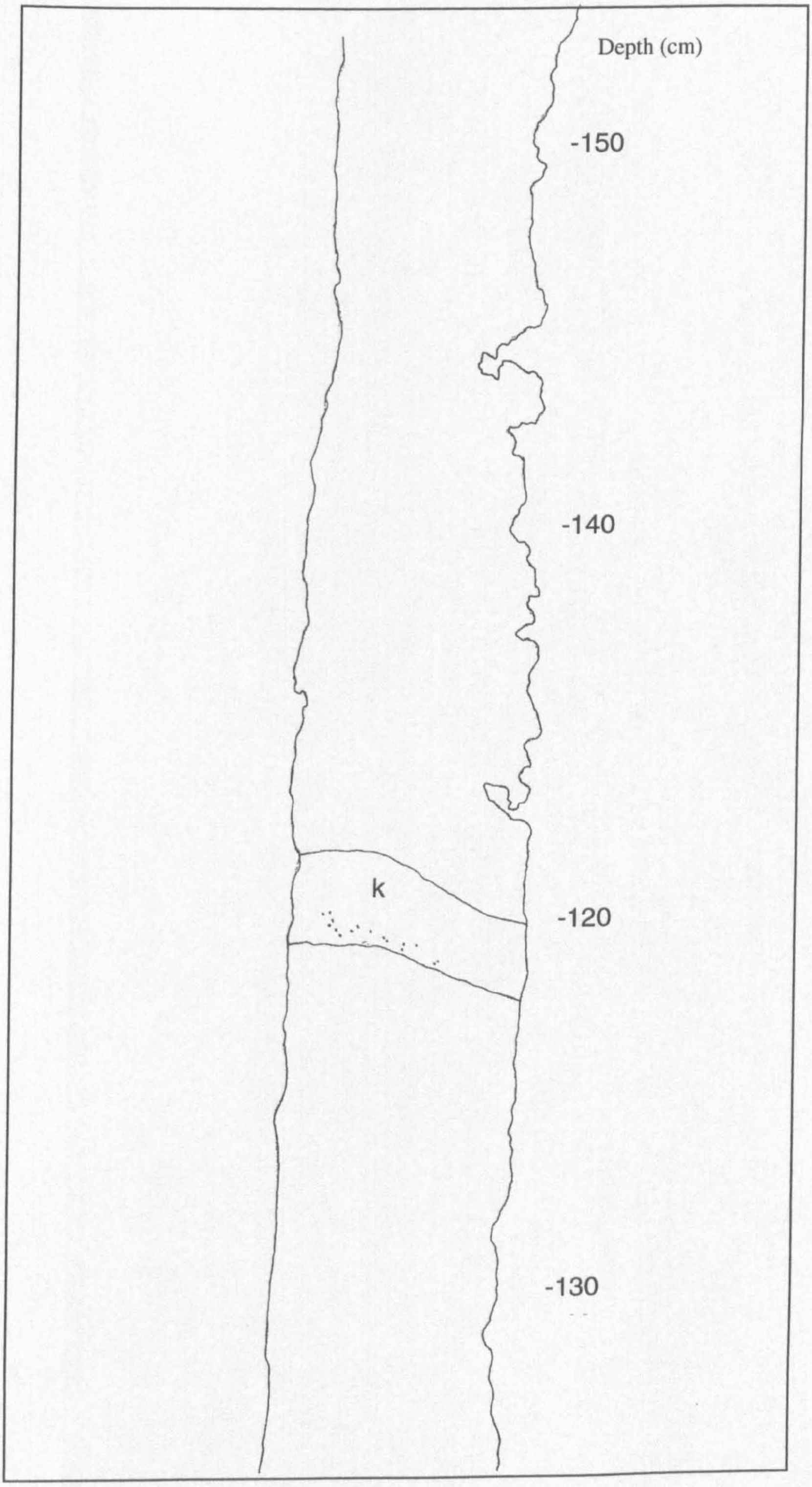


Figure 6.3b Core LCI x-ray image 135-155 cm, Section 6.3.2

Depth cm	LCI Sediment Description
0-61	Brown poorly humified fibrous sedge peat herbaceous, woody and sedge type plant fragments; Th4, Nig2, Strf0, Elas3, Sicc2, Lim1-2.
61-73	Dark brown humified fibrous sedge peat plant fragments more decomposed; Th3 Ld1 DL+, Nig3, Strf0, Elas2, Sicc2, Lim1-2.
73-111	Dark brown humified detrital peat; Dh2, Ld2 Dl+, Nig3, Strf0, Elas1, Sicc2, Lim1-2.
111-138	Finely decomposed, humified detrital peat, monocot plant fragments; Ld3, Dh1, Nig3, Strf0, Elas0, Sicc2, Lim1-2.
138-215	Dark brown humified detrital peat; large (>2mm) herbaceous plant fragments. Dh2, Ld2 Dl+, Nig3, Strf0, Elas1, Sicc2, Lim1-2.
215-219	Dark brown poorly humified fibrous sedge peat, predominantly sedge plant fragments; Dh3, Ld1, Nig3, Strf0, Elas2, Sicc2, Lim1-2.
219-229	Dark brown humified decomposed detrital peat; Dh2, Ld2, Nig3, Strf0, Elas1, Sicc2, Lim1-2.
229-235	Dark brown poorly humified fibrous sedge peat, predominantly sedge plant fragments; Dh3, Ld1, Nig3, Strf0, Elas2, Sicc2, Lim2.
235-263	Dark brown humified decomposed detrital peat; Dh2, Ld2, Nig3, Strf0, Elas1, Sicc2, Lim1-2.
263-273	Dark brown humified detrital peat; Dh3, Ld1, Nig3, Strf0, Elas2-3, Sicc2, Lim1-2.
273-308	Brown poorly humified very fibrous sedge peat herbaceous, woody and sedge plant fragments; Th4, Nig2-3, Strf0, Elas3 Sicc2, Lim2.
308-337	Dark brown organic-rich silt large plant fragments (>2mm) flattened black graminoid stems; Ag2 Ld1 Dh1, Nig3, Strat0, Elas0, Sicc2, Lim1-2.
337-345	Orange brown organic rich silt; Ag2 Ld1 Dh1, Gmin+, Nig2-3, Strat0, Elas0, Sicc0
345-351	Dark brown silt; Ag2 Ld2 Dh+, Nig3, Strf0, Elas0, Sicc2, Lim2-3.
351-357	Orange brown organic-rich silt; Ag2, Ld1, Dh1, Gmin+, Nig2-3, Strf0, Elas0, Sicc2, Lim2-3.
357-432	Dark brown organic-rich silt, small (<2 mm) organic detrital plant fragments; Ag3 Ld1 Dh+, Nig2, Strf0, Elas0, Sicc2, Lim4.
432-445	Silver grey coarse silts and sands, mica flakes visible and Ericoid plant fragments; Ag2 Gmin2, Nig1, Strf4, Elas1, Sicc1-2.

Table 6.4 Core LCI sediment description using a modified version of Troels-Smith notation (Section 5.2.1).

Between 420-424 cm a series of sedimentary units (D) of varying minerogenic and organic content is recorded. These units are 1-5 mm thick, sub-horizontal and well defined with sharp upper and lower boundaries. Within these units there are several micro-sedimentary structures: (E) a concentration of mineral grains (< 0.5 mm in size) in a cross cutting channel type structure (1.0 cm in width and 0.2 cm in depth) and (F) a similarly infilled more shallow cross cutting channel type structure (2.3 cm in width, 0.1cm in depth). Any further sedimentary structures such as grading within the units are difficult to ascertain from the X-ray image.

A sub-horizontal minerogenic unit with a gradual (<2 mm >1 mm) lower boundary and a diffuse (> 1 cm) upper boundary is recorded between 412.5 and 406.5 cm (G). The base of this unit, 412.5-413.5 cm appears to be more minerogenic. There is a concentration of minerogenic grains (< 0.5mm in size) between 399-5 and 400.5 cm (H). A further sub-horizontal minerogenic unit (I) with very gradual (<1.0 cm >0.2 cm) upper and lower boundaries is recorded between 390.5 and 386.5 cm. There is concentration of minerogenic grains between 366 and 364 cm (J).

Between 380 and 333 cm X-ray analysis suggests predominantly minerogenic sediment but, with decreasing amounts of mineral grains visible, sediment is increasingly organic-rich. From 333 cm to the top of the core the amorphous texture recorded in the x-ray image indicates the deposition of predominately organic-rich sediment. An increase in minerogenic content is recorded within the organic-rich sediments, between 122-120 cm (Figure 6.3 (b) (K), see Table 6.4). This sub-horizontal unit has a gradual lower boundary and a diffuse upper boundary, with a concentration of mineral particles (0.1-0.3 mm in size) near the base of the unit.

6.3.3 MAGNETIC SUSCEPTIBILITY

This core was used to test the applicability of magnetic susceptibility at Loch Coulavie (Section 5.2.3). Results from magnetic susceptibility analysis proved inconclusive. Sediment composition, water, organic sediment and minerogenic sediments of predominantly quartz and mica were negatively susceptible (diamagnetic) (Dearing 1994) (Section 5.2.3). It is suggested that the absence of ferro-magnetic minerals, highly positively susceptible, within the sediment renders the technique inapplicable to this catchment.

6.3.4 BULK DENSITY, WATER CONTENT AND LOSS-ON-IGNITION.

A high bulk density of >1.2 g/cm^3 at 432-436 cm (Figure 6.4a) reflects the highly minerogenic nature of the sediments (Table 6.4). Above this, bulk density is lower but variable. The data set is incomplete, reflecting difficulties in sampling accurately for bulk density (Section 5.2.4). The lack of variability within the bulk density record, despite changes in stratigraphy identified from both sediment description and loss-on-ignition analysis, suggests that bulk density values are not accurate or sensitive enough to identify variations in sediment density.

Percentage water content (Figure 6.4b) reflects changes in sediment organic content, with a rapid increase in sediment water content at 432 cm as organic content increases and minor fluctuations at 351-353 cm, 343-337 cm, 201-199 cm, 142-140 cm and 120-118 cm (see Table 6.5).

Loss-on-ignition data (Figure 6.4c) indicate that basal sediments (Ag2Gmin2; 434-444 cm) are highly minerogenic, <2%. There is subsequently a rapid increase in organic content with loss-on-ignition rising from <2% to 11% at 432 cm and 25% by 430 cm. Between 430 cm and 317 cm the organic content is highly variable, with variations in the amount of loss-on-ignition coinciding with changes in sediment stratigraphy (Table 6.4). An increase, to above 30%, between 381-355 cm, is followed by a fall to 12% at 344-338 cm. From 338-307 cm there is a sustained rise in loss-on-ignition with >70% by 307 cm indicating increasingly organic sediment, and above 307 cm sedge peat with a loss-on-ignition value of >82% accumulates. However, within the sedge peat several abrupt reductions in loss-on-ignition are recorded, described in Table 6.5.

Depth (cm)	**% loss-on-ignition.**	**Number of analyses**	**Event (LCI)**	**Notes**
88-92	76-78%	2	(8)	Loss-on-ignition residues noted to contain distinctive mica rich minerogenic sediments. During humification analysis (Section 6.2.6) this unit noted to contain minerogenic sediments. Reduction in percentage water content (Figure 6.4)
118-122	68-71%	2	(7)	Loss-on-ignition residues noted to contain distinctive mica rich minerogenic sediments. This unit is a well defined minerogenic sedimentary unit on x-ray analysis (Section 6.2.3). During humification analysis (Section 6.2.6) this unit noted to contain minerogenic sediments. Reduction in percentage water content (Figure 6.4).
198-200	74%	1	(6)	Loss-on-ignition residues noted to contain distinctive mica rich minerogenic sediments. Reduction in percentage water content (Figure 6.4).
252-254	60%	1	(5)	Loss-on-ignition residues noted to contain distinctive mica rich minerogenic sediments

Table 6.5. Description of units showing abrupt reductions in percentage loss-on-ignition within sedge peat, core LCI

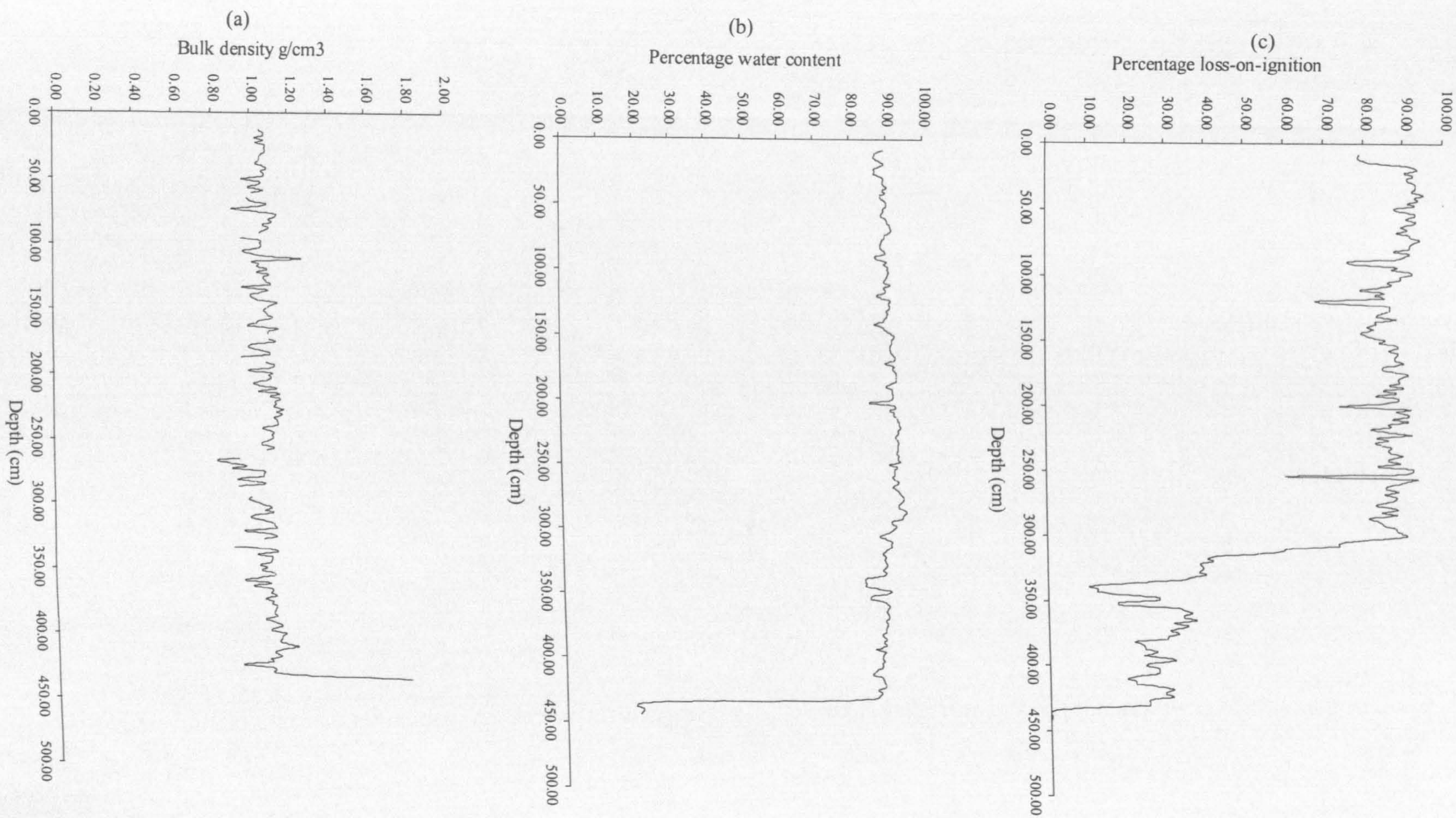


Figure 6.4 Core LCI bulk density (a), percentage water content (b) and percentage loss-on-ignition (c) plotted against depth (cm).

These reductions in loss-on-ignition value are considered as significant changes in the minerogenic content of the sedge peat, with the reduction in loss-on-ignition within these sediments exceeding the 1σ Standard Deviation, calculated for loss-on-ignition within this highly organic sediment (307-9 cm) (n=149; 87 ±5.3%). The abrupt reductions at 200-198 cm and 254-252 cm are only recorded by one analysis. Further evidence (Table 6.5) suggests that these reductions in loss-on-ignition are the result of increased mineral content within the sediment.

6.3.5 GRAIN SIZE ANALYSIS

Grain size data are presented as changes in percentage volume for grain size classes (Section 5.4, Table 5.4) plotted against depth (Figure 6.5). Zonation within Figure 6.5 is based on a visual interpretation of the data. Grain size analyses were not obtained for all sediments, in particular those with high organic content above 307 cm (Section 5.4). The basal highly minerogenic sediments (444-434 cm; zone LCIa) are very coarse grained, typically made up of sands and coarse silts, with 20-30% very fine sands, 10-12% fine sand, 4-6% medium sands and 2-5 % coarse sand. The transition from these highly minerogenic sediments to more organic sediments (432-309 cm) is marked by a gradual decrease in the amount of coarse silts and very fine sand and a more abrupt increase in finer grained sediments (zone LCIb). In general these more organic sediments show little variation in grain size made up of fine (24-30%) and medium silts (20-30%) and only small amounts of very fine sands (3-4%). However, during the accumulation of this more organic sediment there are several abrupt single-sample increases in very fine sand and coarse silt content, at 404 cm, 412 cm, 392 cm, 389 cm and 377 cm.

Grain size analyses (Figure 6.5) of the sediment units described as abrupt reductions in loss-on-ignition (Table 6.5) record similar grain size compositions to sediments between 432-309 cm discussed above and are predominantly made up of fine (23-28%) and medium silts (18-29%) with low amounts (5-6%) of very fine sands.

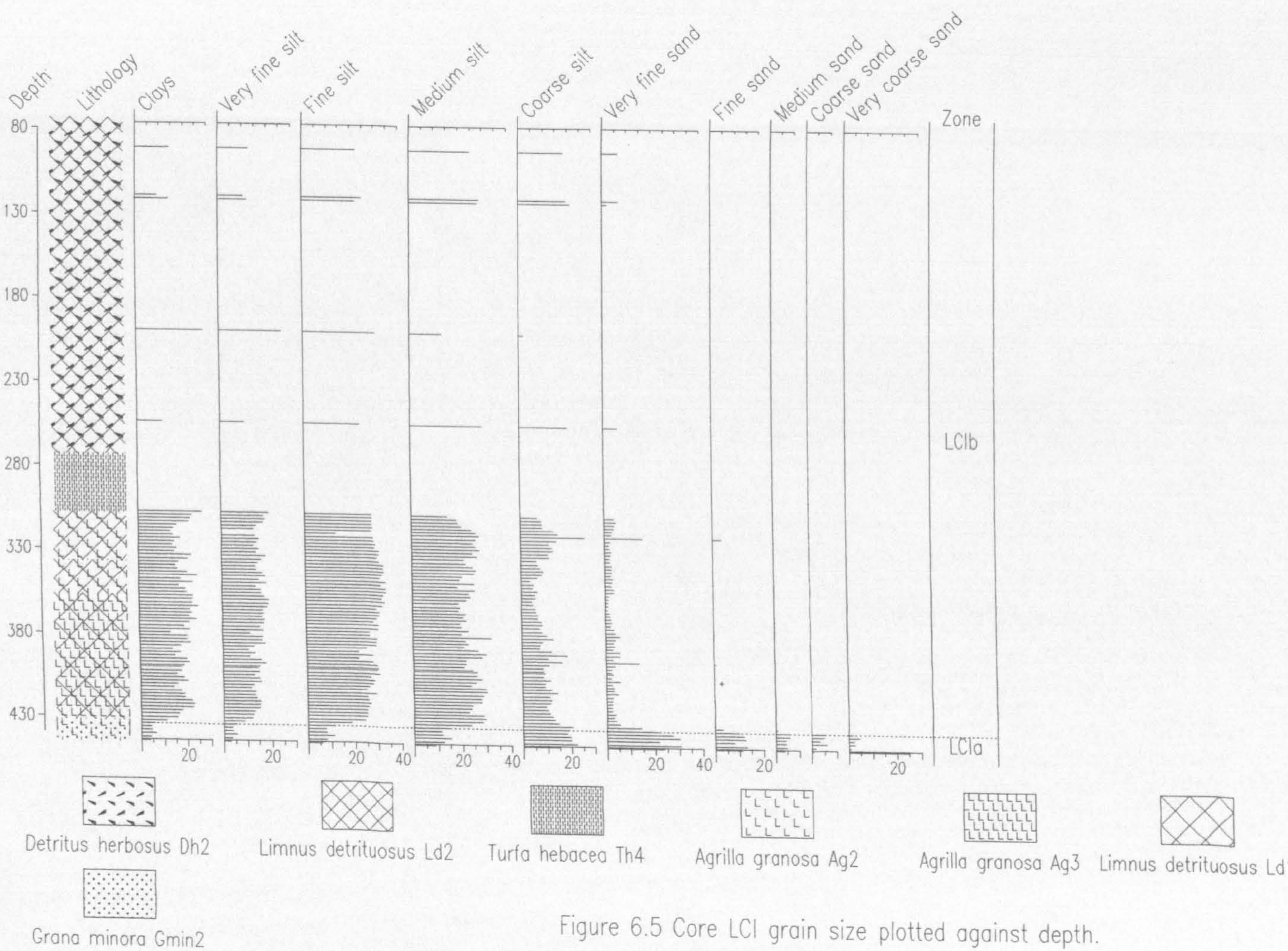
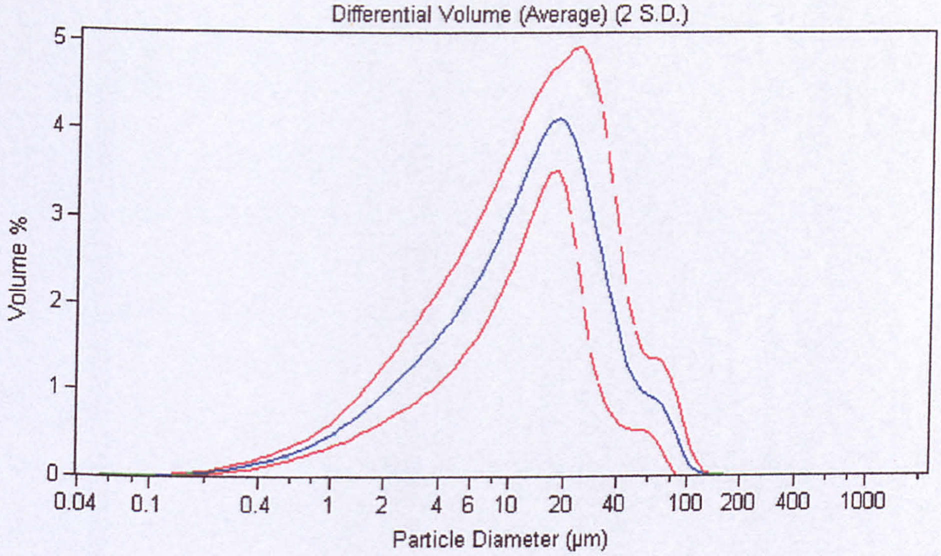


Figure 6.5 Core LCl grain size plotted against depth.

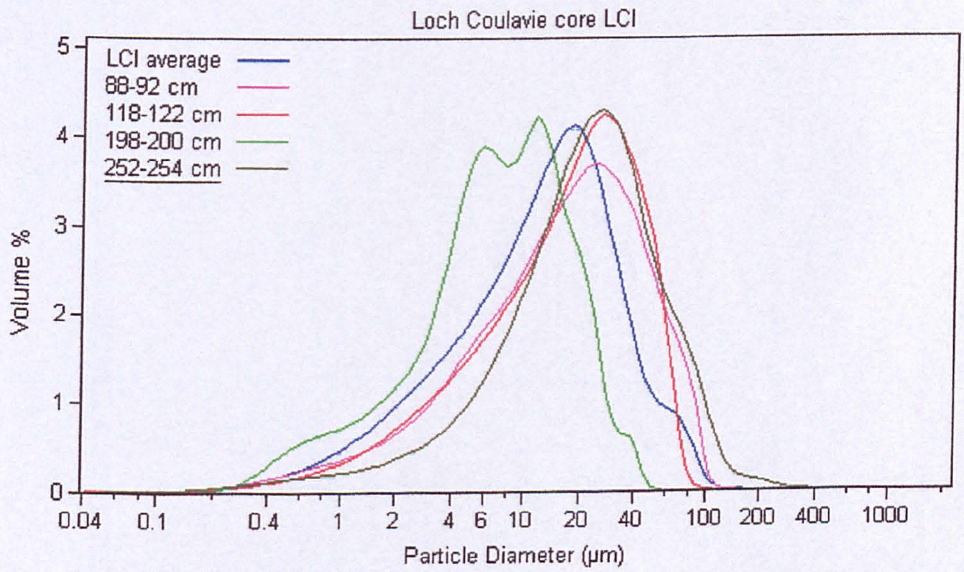
Further definition of the mode of deposition of sediment units described in *Table 6.5* was through examination of particle size distributions (Section 5.3.1). A mean particle size distribution considered typical of sediment deposited by lacustrine processes was defined using loss-on-ignition data (Figure 6.4), sediment description (Table 6.4) and grain size data (Figure 6.5); sediments between 432-309 cm were used. Particle size distributions for these depths were averaged (n=62) using the COULTER LS230 software to produce a typical or average particle size distribution curve plotted at 2σ standard deviation (SD) (Figure 6.6 (a)). Figure 6.6 (b) overlays the particle size distributions of the sediment units 92-88 cm, 122-118 cm, 200-198 cm and 254-252 cm (Section 6.2.9, events LCI (5-8)) with the distribution curve typifying lake sediments labelled as the LCI average.

The particle size distributions of minerogenic sediment units deposited within the sedge peat show comparable distributions to typical lake sediments. Three units 254-252 cm, 122-118 cm and 92-88 cm, lie within the $2\ \sigma$ SD for the sediment average. This indicates a common mode of deposition for all these finer grained sediments, one controlled by 'normal' lake sedimentation processes and not products of exceptional or extreme depositional processes. The unit 200-198 cm lies outside the $2\ \sigma$ SD showing that this unit is finer grained.

Particle size distributions for sediment units, 122-118 cm and 92-88 cm, which occur over 4 cm (2 grain size samples) are plotted in Figures 6.7 and 6.8. There was no significant variation in grain size during the deposition of each unit. This lack of sediment grading within the unit indicates no change in the energy of deposition during sedimentation and provides further evidence for a mode of deposition controlled by 'normal' lacustrine processes.



(a)



(b)

Figure 6.6 (a and b) LCI percentage volume particle size distributions. (a) Average particle size distributions for the basal organic sediments (see text for definition) to a standard deviation of 2 σ. (b) overlays of the particle size distributions of the units identified as reductions in loss-on-ignition (events LCI (5-8)) and the average particle size distribution (note here the 2 σ standard deviation on this average particle size distribution is not shown).

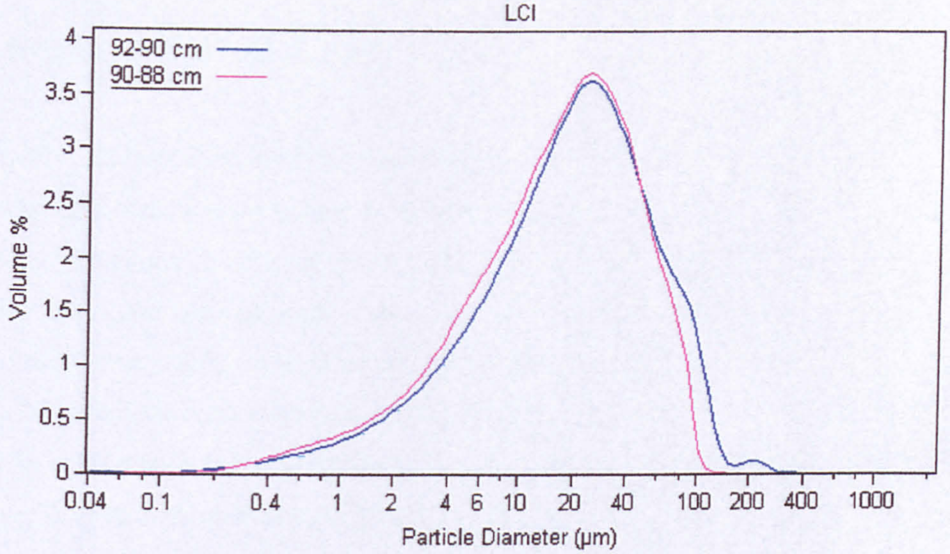


Figure 6.7 Core LCI particle size distributions for sediments within unit 92-88 cm (event LCI 8), see Section 6.2.5.

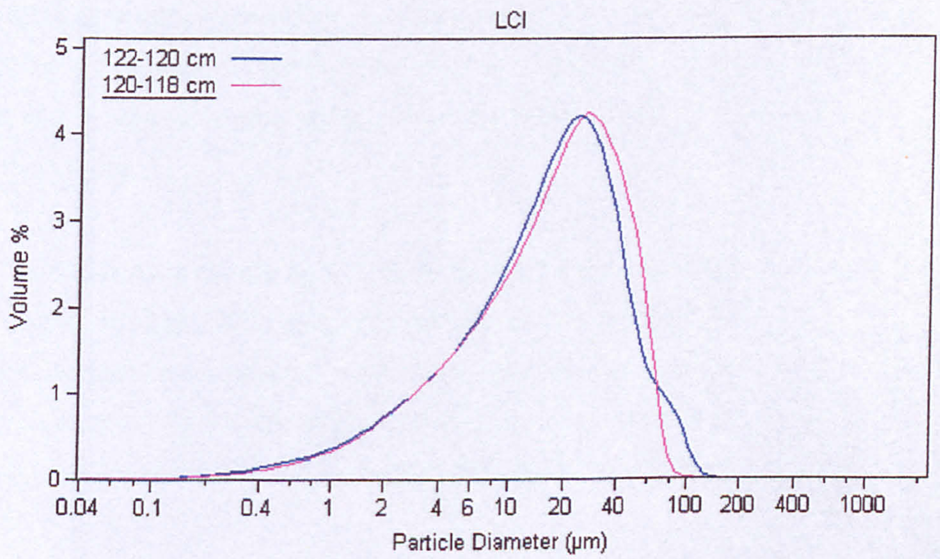


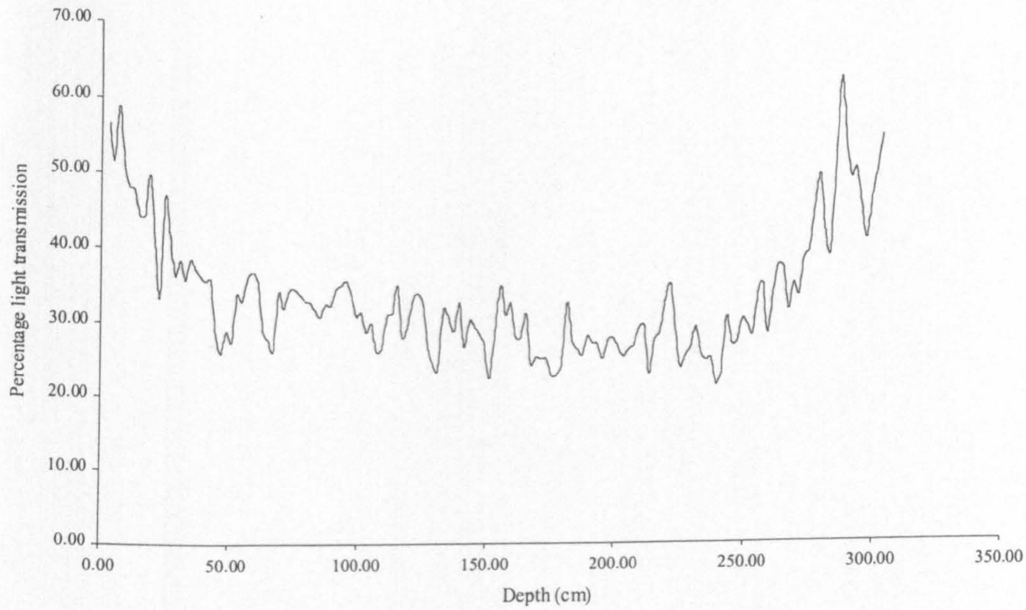
Figure 6.8 Core LCI particle size distributions for sediment within unit 122-118 cm , (event LCI 7) see Section 6.2.5.

6.3.6 HUMIFICATION ANALYSIS

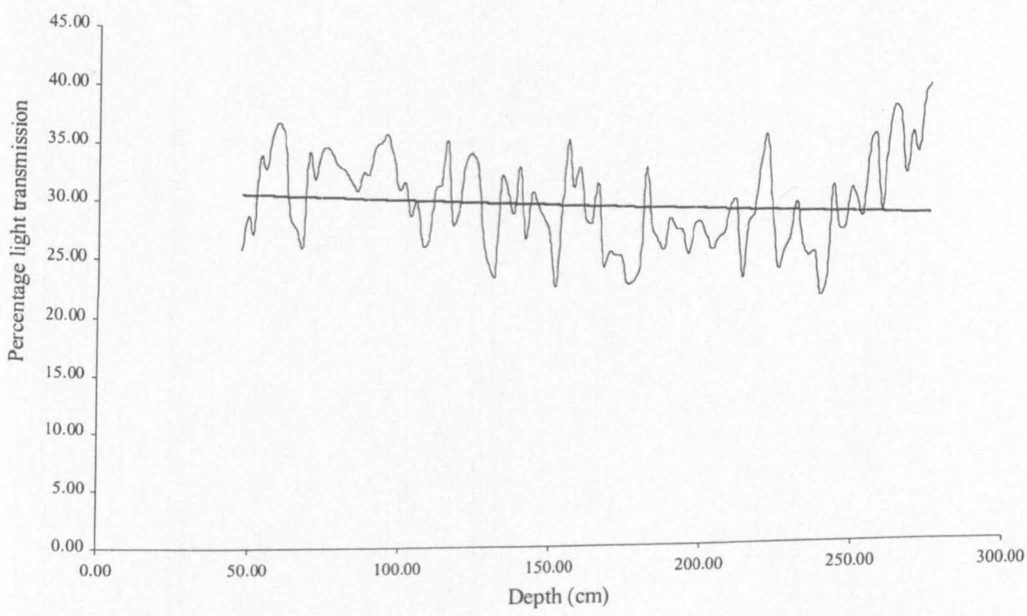
Humification analysis (Section 5.4) was carried out on sedge peat (307-3 cm); the percentage light transmission values are shown in Figure 6.9 (a). Humification data can be affected by changes in minerogenic content and vegetation type as much as by decay processes (Blackford and Chambers 1993). Percentage light transmission data were tested to show if variability within the data set could be explained by minerogenic content. Minerogenic content and percentage light transmission data sets produced a correlation coefficient value of 0.04, indicating that changes in minerogenic content do not appear to control the percentage light transmission value. Humification changes related to vegetation type were compared qualitatively when plotted against changes in percentage light transmission data (Figure 6.10). Notably higher percentage light transmission values recorded at 308-276 cm and 46-3 cm correspond with highly fibrous sedge peat (Th4). Variation in vegetation type could account for these higher percentage light transmission values and thus no inferences on changes other than hydroseral on the degree of humification are made. Throughout the remainder of the core changes in vegetation type appear to have no control on the percentage light transmission value.

Section 5.4 outlines a methodology to correct the peat humification data for potential decay at depth. In Figure 6.9 (b) an exponential regression curve fitted to the percentage light transmission data (excluding values controls by vegetation type; coarse fibrous peat units at 308-276 cm and 46-3 cm) indicates no trend within the data indicative of decay at depth: thus no attempt has been made to detrend the data.

The methodology for humification data analysis follows that outlined in Section 5.4. The mean percentage light transmission value for peat samples between 275-46 cm was calculated to 1σ Standard Deviation (n=115), Figure 6.9. Significant shifts were defined as those that extended above or below the 1 σ range (Table 6.6). The data-set was divided up into a series of shifts in percentage light transmission values (humification), with the beginning and end of each shift defined from the 'turning' point of the data curve (Section 5.4, Table 6.6).



(a)



(b)

Figure 6.9 (a) LCI raw percentage light transmission data for the whole sedge peat record (308-3 cm); (b) percentage light transmission data (275-46 cm) based on interpretation of Figure 6.10 (see text) percentage light transmission values between 46-3cm and 308-276 cm are disregarded. An exponential regression is fitted to this data set.

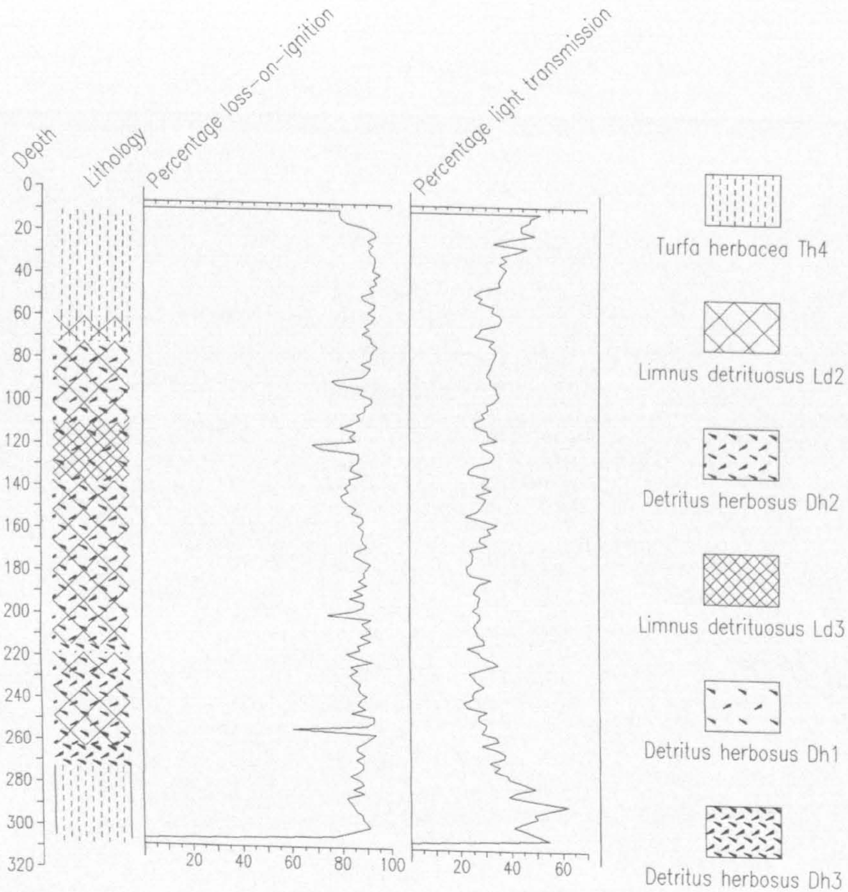


Figure 6.10 LCI %light transmission, LOI and stratigraphy.

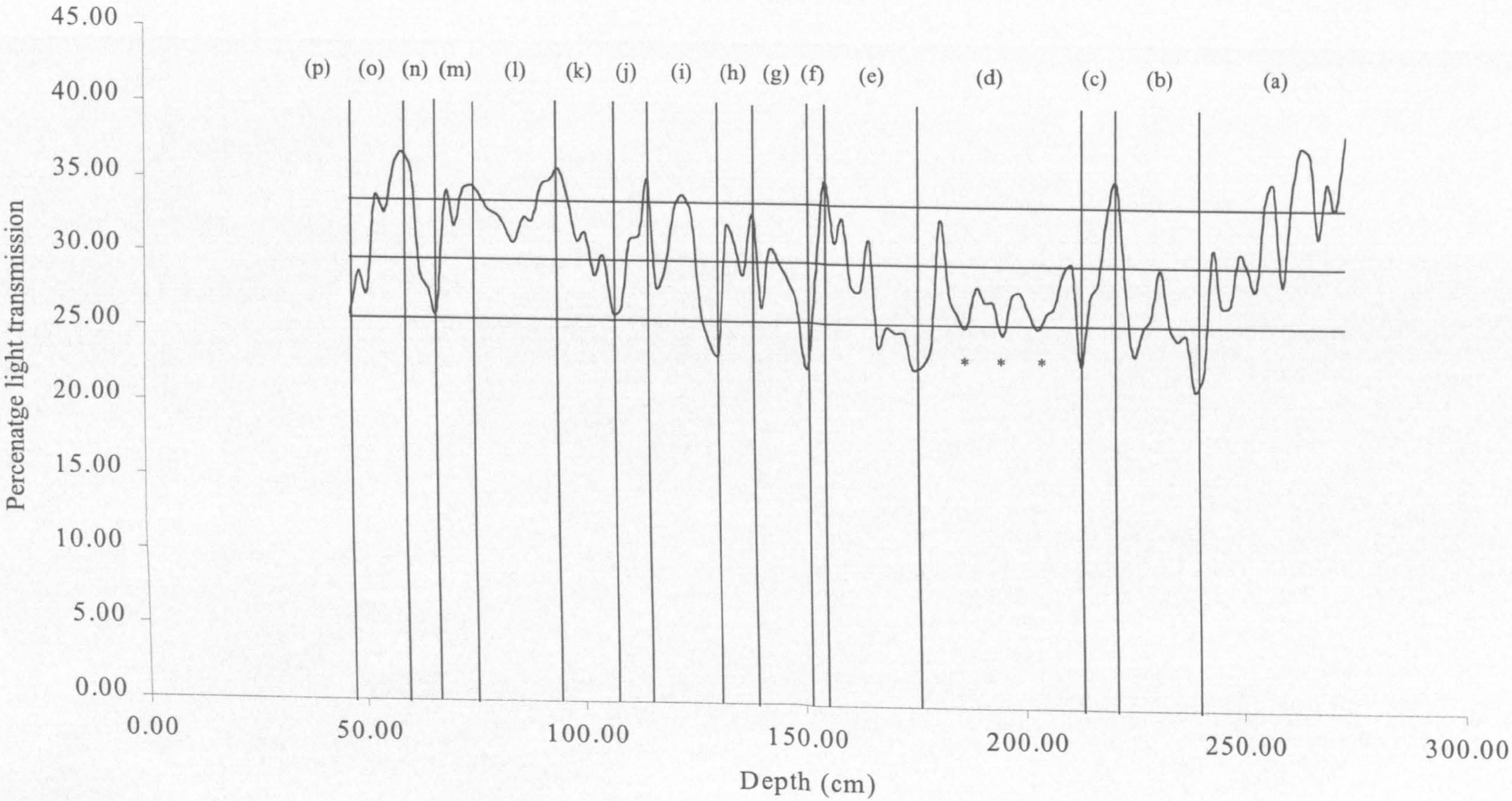


Figure 6.11 LC1 Percentage light transmission data (46-275 cm) plotted with mean value and 1σ standard deviation (see text). See text and Table 6.6 for further discussion of zones (a-p). * significant shifts discussed in Table 6.6.

Zone	Depth (cm)	Degree of Peat Humification
(a)	287-239	The general trend is a sustained increase in peat humification from significantly higher to significantly lower percentage light transmission values. However, trend of increasingly humified peat consists of a sequence of highly variable percentage transmission values.
(b)	239-221	A rise in percentage light transmission values culminating in a significant and abrupt decrease in peat humification.
(c)	221-213	An abrupt increase in the degree of peat humification with a shift from significantly higher to significantly lower percentage light transmission values.
(d)	213-197	A sustained period of lower percentage light transmission values, with significant shifts to humified peat at 203cm, 195cm and 187 cm. There is an abrupt though not significant shift to more poorly humified peat at 181 cm.
(e)	197-155	A gradual increase in percentage light transmission values from significantly lower to significantly higher values suggesting the accumulation of more poorly humified peat
(f)	155-151	Very abrupt fall to significantly lower percentage light transmission values suggesting an abrupt shift to more humified peat.
(g)	151-139	A gradual increase but not to significantly higher percentage light transmission values suggesting the accumulation of more poorly humified peat.
(h)	139-131	An abrupt fall to significantly lower percentage light transmission values suggesting an abrupt shift to more humified peat.
(i)	131-115	A period of fluctuating percentage light transmission values with a general rise in values with significant shifts to more poorly humified peat at 123cm
(j)	115-107	A fall in percentage light transmission values from significantly higher values indicating a shift to more humified peat.
(k)	107-95	A sustained rise to significantly higher percentage light transmission values suggesting the accumulation of more poorly humified peat.
(l)	95-75	A period of gradual fluctuations in high percentage light transmission values suggesting a period of poorly humified peat.
(m)	75-67	An abrupt fall from significantly higher percentage light transmission values suggesting a shift to more humified peat
(n)	67-59	An abrupt rise to significantly higher percentage light transmission values suggesting a shift to more poorly humified peat
(o)	59-46	A decrease in percentage light transmission values suggests a shift towards more humified peat.

Table 6.6 Core LCI Summary of humification analysis and description of zones of main shifts in surface wetness as shown in Figure 6.11.

Despite the complexity of the record, there are some general patterns to the nature of shifts in the humification record. There are several periods of gradually decreasing peat humification (increasing percentage light transmission values) ended by abrupt shifts to more humified peat (decreasing percentage light transmission values) as recorded in the zones labelled as (b)-(c), (e)-(f), (g)-(h), (i)-(j), (l)-(m) and (n)-(o). There are few sustained periods of either poorly or well humified peat. Only in (d) is there a period of relatively well humified peat and at (l) a period of poorly humified peat.

6.3.7 POLLEN ANALYSIS

Pollen analysis (Section 5.5) was used to identify changes in aquatic and local mire vegetation types during peat accumulation. The pollen stratigraphy generated was also used to test the radiocarbon chronology generated for core LCI (Sections 5.6 and 6.2.8). Samples for pollen analysis were taken at peat depths that represented shifts in peat humification (Figure 6.11; Table 6.6), described in Table 6.7.

Interpretation of changes in aquatic vegetation through the generation of a local pollen sum and the use of water depth-sensitive plant taxa is discussed in Section 5.5. Vegetation changes generated through this local pollen sum are shown in the pollen diagram (Figure 6.12), with pollen zones defined on visual inspection of major changes. Pollen from aquatic taxa at all sample points is however poorly represented, and it is difficult to draw any significant conclusions on changes in aquatic vegetation and water depth. The following section summarises the main vegetation changes generated from this local pollen sum (Figure 6.12).

6.3.7.1 Local Mire and Aquatic Changes

Figure 6.12 suggests some general changes in aquatic species through time. The presence of *Myriophyllum*, an obligate local aquatic species with a water depth requirement of between 40-250 cm (Section 5.5, Table 5.5) suggests that the mire prior to 210 cm was close to lake-level.

Sample Depth (cm).	Humification	Vegetation Type
11	Poorly humified, peat surface, percentage transmission 48%.	Th4, fibrous, vascular vegetation type. Organic content 79%.
47	Highly humified peat, percentage transmission 25%	Th4 fibrous, vascular vegetation type. Organic content 94%.
59	Poorly humified peat. Percentage transmission 37%.	Th3 Ld1 fibrous vegetation but becoming more broken down. Organic content 94%.
67	Well humified peat. Percentage transmission 26%.	Th3 Ld1 fibrous vegetation but becoming more broken down. Organic content 89%.
93	Poorly humified peat. Percentage transmission 34%. Corresponds to the deposition of minerogenic sediments just above the sample at 88-92cm, event LCI(8).	Dh2 Ld2 detrital vascular type vegetation becoming amorphous. Organic content 80%.
107	Well humified peat. Percentage transmission 26%.	Ld3 Dh1 amorphous peat with some plant fragments visible. Organic content 88%.
123	Poorly humified peat. Percentage transmission 34% Corresponds to the deposition of minerogenic sediments just above the sample at 118-122cm, event LCI(7).	Ld3 Dh1 amorphous peat with some plant fragments visible. Organic content 86%.
131	Well humified peat. Percentage transmission 23%.	Ld3 Dh1 amorphous peat with some plant fragments visible. Organic content 84%.
159	Poorly humified peat. Percentage transmission 33%	Dh3 Ld1 detrital peat with vegetation fragment well preserved. This change in vegetation type over 4cm. Organic content 89%.
175	Well humified peat. Percentage transmission 23%.	Dh2 Ld2 partially amorphous peat will well preserved fragments of vegetation. Organic content 92%.
198	Well humified peat percentage transmission 27%. This sample point is within a prolonged period of well humified peat. Also at a point after minerogenic sediments recorded at 198-200cm, event LCI (6).	Dh2 Ld2 partially amorphous peat will well preserved fragments of vegetation. Organic content 84%.
221	Poorly humified peat Percentage transmission 35%.	Dh2 Ld2 partially amorphous peat will well preserved fragments of vegetation. Organic content 92%.
239	Well humified peat. Percentage transmission 21%.	Dh2 Ld2 partially amorphous peat will well preserved fragments of vegetation. Organic content 88%.
252	Poorly humified peat percentage transmission 30%. This sample point also lies above minerogenic sediment input at 252-254 cm, event LCI (5).	Dh2 Ld2 partially amorphous peat will well preserved fragments of vegetation. Organic content 90%.
289	Very poorly humified peat. Percentage transmission 54%.	Th4 very fibrous woody/vascular type of vegetation. Organic content 85%.

Table 6.7. Pollen analysis sample description (Troels-Smith notation), percentage organic content (loss-on-ignition) and percentage transmission (alkali extraction) for fen peats within core LCI.

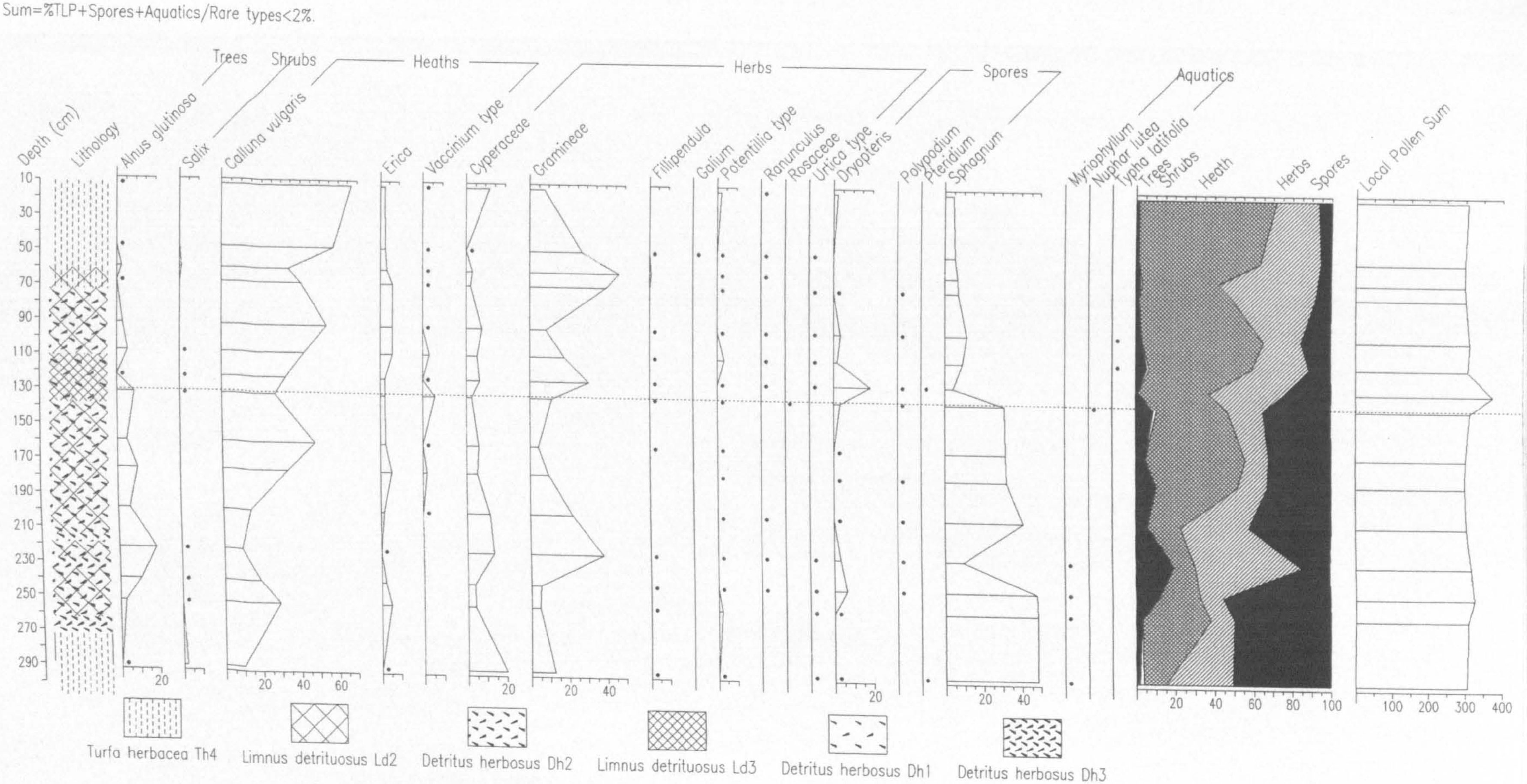


Figure 6.12 Core LCI local pollen percentage diagram.

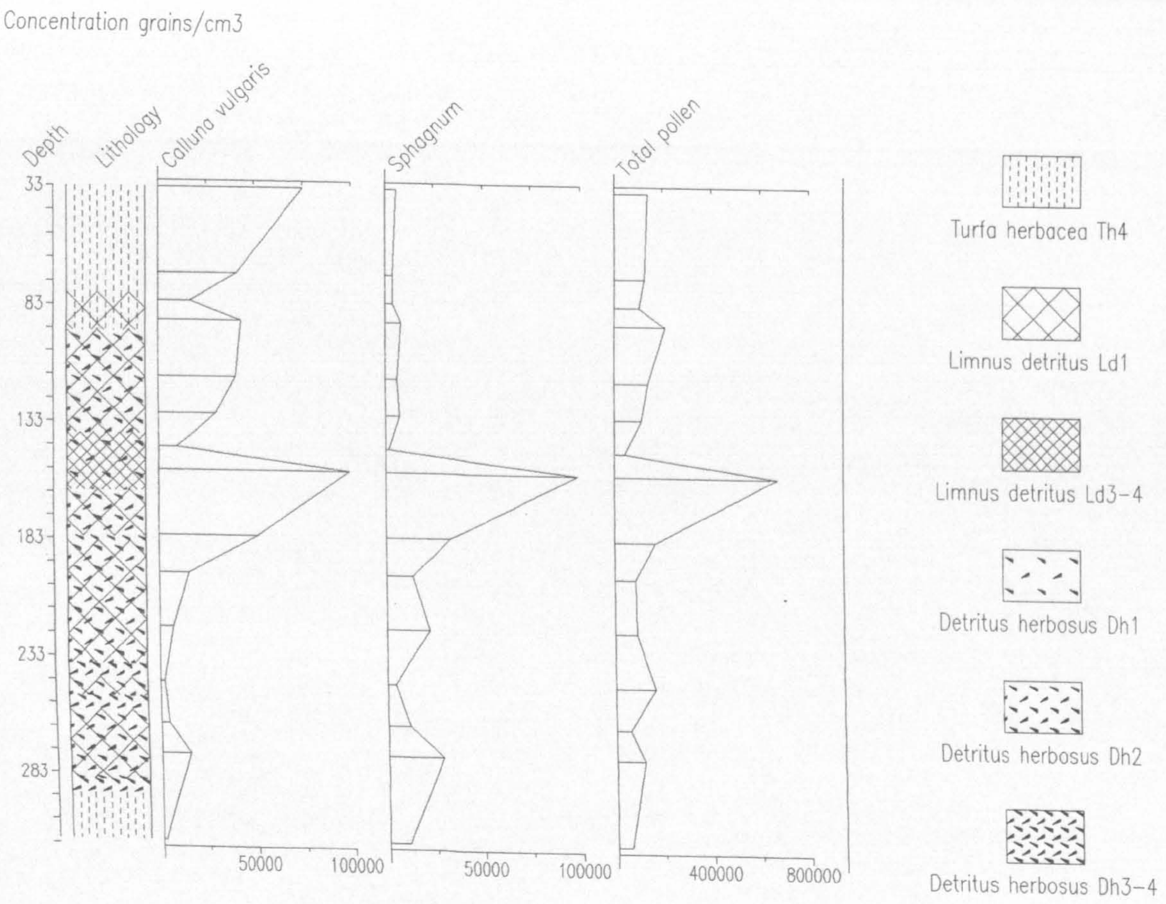


Figure 6.13 Core LCI Pollen concentration diagram

Above 210 no aquatic taxa are recorded until 130 cm, with the presence of *Nuphar lutea* and *Typha latifolia* species associated with shallower water 0-150 cm (Section 5.5, Table 5.5). Interpretation of these data is given in Section 7.6.2.1 Figure 6.12 suggests that during peat accumulation the local basin vegetation was dominated by Cyperaceae and Gramineae. These plant families would probably be associated with marsh and fen communities. *Calluna vulgaris*, *Erica* sp., *Sphagnum* and herbaceous plants such as *Potentilla* type and *Filipendula* are associated with damp, acid mire conditions around the basin margin. *Alnus,* present throughout the peat sequence in low amounts together with *Salix*, is associated with seasonally waterlogged habitats around the basin margin. During the accumulation of sedge peat, local vegetation change records a uni-directional vegetation change with the replacement of *Sphagnum* species with *Calluna vulgaris*, above 130 cm (Section 7.6.2.1).

A local pollen zone boundary at 130 cm indicates a poorly defined shift in mire hydrology. Below 130 cm, in particular below 210 cm, wetter mire hydrology is suggested by increased amounts of *Alnus* and *Sphagnum*, the presence of *Salix* and deeper water aquatic such as *Myriophyllum*. Above 130 cm there is a decline in *Alnus* and *Sphagnum* with an increase in herbs such as *Potentilla* and the presence of shallow-water aquatics suggesting a shift to drier mire conditions.

Figure 6.13 shows changes in *Calluna vulgaris* and *Sphagnum* concentrations alongside those of total pollen. These taxa represent contrasting hydrological conditions, with changes in concentrations interpreted as shifts in mire surface wetness. Major changes in local pollen concentrations are few. Of most significance are synchronous substantial increases in *Calluna*, *Sphagnum* and total pollen between around 180 and 130 cm. These parallel increases within these two taxa are thought to represent a reduction in peat accumulation rather than increases in abundance or pollen productivity of plants near the sampling site. The reduction in peat accumulation may relate to drier mire surface conditions. At around 130 cm the sharp reduction in concentration values may reflect an increase in peat accumulation rate through wetter mire conditions. Increases in *Calluna* concentrations above 130 cm probably reflect the increasing dominance of *Calluna* in the local mire plant community because total pollen concentrations do not increase significantly (Section 7.6.2.1).

Section 5.5 outlined a methodology for assessing the amount of corroded pollen, with increased amounts of corrosion associated with microbial decay under aerobic conditions within the peat (Havinga 1984). The corrosion data for both *Betula* and *Corylus* grains (Figure 6.14) is highly variable with few trends. The low sampling resolution necessarily employed limits analysis. Below 239 cm corrosion is high though not in both taxa and thus interpretation is difficult. Between 239-221 cm and 11-47 cm corrosion is consistently low which might suggest two phases of reduced decay in peats which were less aerobic and possibly wetter.

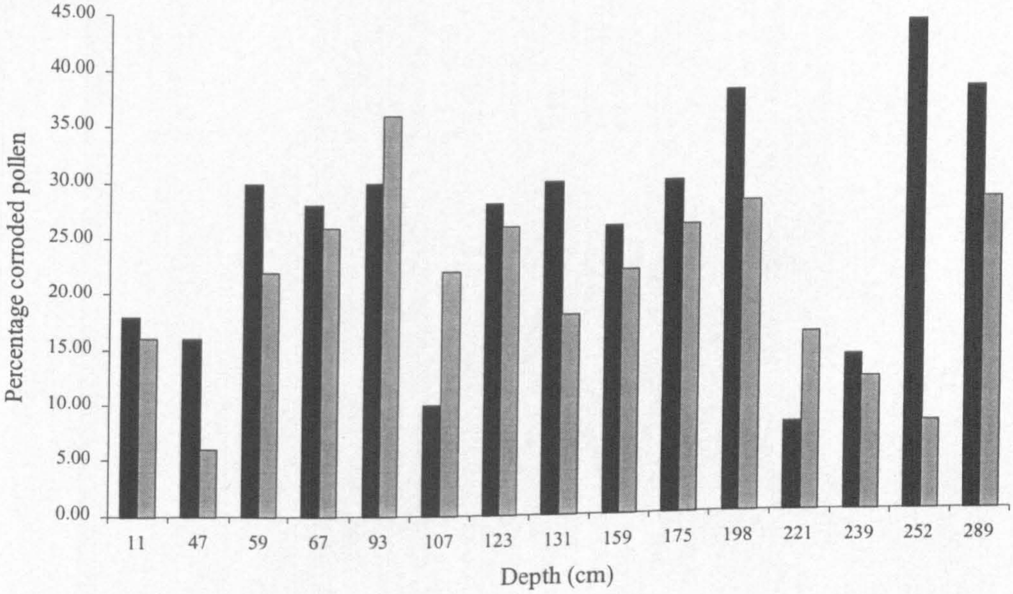


Figure 6.14 LCI Percentage corroded pollen (Section 5.5). Percentage *Betula* grains (black bars) and percentage *Corylus* grains (grey bars).

6.3.7.2 Regional Vegetation Change

The total pollen sum diagram (Figure 6.15) contains a more regional pollen signal with the major tree taxa included in the pollen sum. This total pollen sum was generated to cross-check the radiocarbon chronology (Section 6.3.8) against known-age biostratigraphic boundaries from Torran Beithe, a palynological site within the Loch Coulavie region (Davies 1999; Figure 4.5, Sections 4.7, 5.6; Table 5.6).

Figure 6.15 Core LCI Total pollen percentage diagram

Pollen zone boundaries have been placed on the diagram to define the main changes in vegetation (Figure 6.15). These are considered to be synchronous with vegetation changes described at Torran Beithe. Table 6.8 outlines an inferred chronology for Loch Coulavie pollen zones and vegetation change events based on pollen stratigraphic evidence from Torran Beithe.

Pollen zone	Depth (cm)	Inferred Chronology	Vegetation description and estimated event chronology.
Calluna-Betula LCIc	11-53	500 BP (500 cal. BP)-2000 BP (1950 cal. BP)	Predominantly *Betula* with increasing amounts of *Calluna*, declining *Pinus*, *Corylus*, and Gramineae. No rise *Pinus* associated with plantations in last *c*. 100 yrs.
Pinus-Betula-Calluna LCIb	53-210	2000 BP (1950 cal. BP)-5500 BP (6300 cal. BP).	Gradually declining *Pinus* with increasing *Betula* and *Calluna*. Low amounts of *Alnus* and *Corylus*. *Quercus*, *Sorbus* and *Ulmus* are present. Potential *Ulmus* decline at around 180 cm *c* 5250 BP (6000 cal. BP) and a gradual *Alnus* decline and increased *Calluna* from around 120 cm, *c*. 3200 BP (3600 cal. BP).
Pinus-Betula-Corylus LCIa	210-289	5500 BP (6300 cal. BP)-6600 BP (7400 cal. BP)	Predominantly *Pinus* with *Betula* and *Corylus*. With *Pinus* dominant at 289 cm *c* prior to 6600 BP (7400 cal. BP), but starting to decline at around 210 cm. Low amounts of *Alnus*, *Quercus*, and *Ulmus*. Potential Alnus rise at 250 cm around 6300 BP (7200 cal. BP).

Table 6.8 LCI Inferred chronology for pollen zones and vegetation changes (Figure 6.15) based on chronologically well defined regional bio-stratigraphic data (Sections 4.7 and 5.6).

The ages of vegetation change outlined in Table 6.8 are estimated as a low sampling resolution within LCI generates only approximate depths for chnages. For example *Pinus* shows no abrupt decline at Loch Coulavie and cannot be easily compared with Torran Beithe. However, bio-stratigraphic correlation indicates that sediments above 300 cm at Loch Coulavie are not early Holocene.

6.3.8 CHRONOLOGY

The chronology of core LCI was defined more precisely from a sequence of AMS radiocarbon ages (Section 5.6). Radiocarbon ages were obtained to define (a) the start of organic sedimentation (430 cm) and (b) the start of sedge peat accumulation (303 cm) as identified from loss-on-ignition data (Figure 6.4). Radiocarbon assays were also obtained to define the timing of minerogenic sediments deposited within the sedge peat

(88-92 cm, 118-122cm, 198-200 cm and 252-254 cm) as recorded by abrupt reductions in loss-on-ignition (Table 6.5).

The radiocarbon dates (Table 6.9) do not present a conformable sequence. One assay (AA-34427: 303.0 cm) is not in the correct stratigraphic order. Palynological comparisons (Figure 6.15; Table 6.8) with radiocarbon dated vegetation history at nearby Torran Beithe (Sections 5.6, 6.3.7.2) would suggest that assay AA-34427 is in error. The sediments at around 300 cm are likely to have an age of between 6500-6000 BP (7400-6800 cal BP) (Section 6.3.7.2, Table 6.8) indicating that assay AA-34427 of c. 1460 BP (1340 cal. BP) is too young. The reasons for this are unclear. Assay AA-34427, dating the onset of peat accumulation, is rejected from further analysis. The age of this sediment-stratigraphic boundary is estimated by linear interpolation between assays AA-34224 and AA-34233, Figure 6.16. There is a large depth interval and timespan between these assays (178 cm; *c.* 2900 radiocarbon years) and the age-estimate for peat inception at 303.0 cm, *c.* 7100 BP (8100 cal. BP).

Laboratory reference	Depth (cm)	C content (% by wt)	Conventional Radiocarbon age (yrs BP ± 1σ)	Calibrated age range to 1σ (yrs. BP)	Calibrated age range to 2σ (yrs. BP).
AA34236	88	27	2410 ± 50	2705-2350	2712-2341
AA34235	118	30	3060 ± 50	3354-3173	3379-3080
AA34234	198	35	5000 ± 80	5890-5623	5918-5592
AA34233	252	58	6280 ± 60	7266-7098	7317-7011
AA34227	303	44	1460 ± 40	1389-1306	1413-1291
AA34224	430	8.7	9130 ± 65	10378-10219	10481-10189

Table 6.9 Core LCI AMS radiocarbon dates and calibrated dates in cal. yrs. BP

Other radiocarbon assay in the series are considered correct. These assays lie in the correct stratigraphic order and are in agreement with the broad chronology suggested in Table 6.8 (Section 6.3.7.2).

The age-depth curve in Figure 6.16 indicates that sediment accumulation rates within core LCI have been relatively constant. Initial higher sedimentation rates, 0.62mm/yr are associated with more minerogenic sediments. Based on the dating resolution, accumulation rates within the sedge peat suggest a constant rate of accumulation with minerogenic sediment events not causing any change in the accumulation rate.

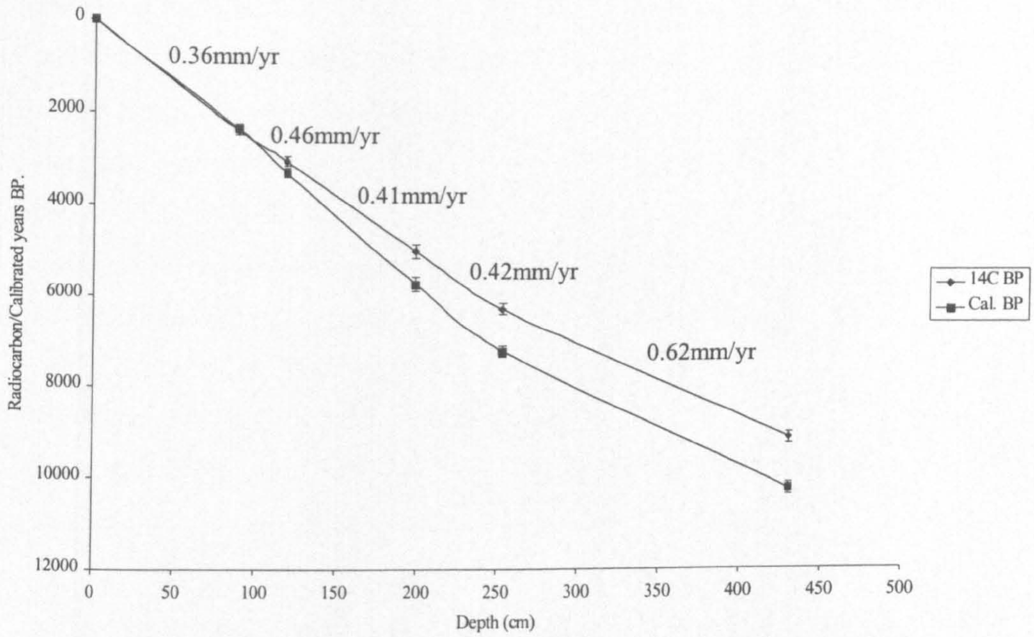


Figure 6.16 Core LCI Radiocarbon (14C BP) and Calibrated ages (cal. BP) years plotted against depth. Error values are to 2σ. Accumulation rates are calculated as mm per radiocarbon year.

6.3.9 SUMMARY OF STRATIGRAPHIC EVENTS

Sediment analysis has defined several significant sedimentary events within core LCI; these are summarised in Figure 6.17, outlined below and further discussed in Chapter 7.

- Event LCI 1: deposition of minerogenic laminated silts and sands, between 445-432cm (see Section 7.2).
- Event LCI 2: marks the start of organic sedimentation at 430 cm as defined from sediment description (Section 6.3.1) and loss-on-ignition (Section 6.3.4) and grain size analysis (Section 6.3.5). Sediment organic content increases rapidly with <2% loss-on-ignition at 434cm and 25% by 430cm. Radiocarbon assay suggests that onset of organic sedimentation occurs at 9130 ± 65 BP (10 240 cal. BP) (see Section 7.3).
- Event LCI 3: defined from x-ray analysis (Section 6.3.2) as a series of alternating thin (1-5mm) organic and minerogenic sediment units between 424-420 cm, with bedding structures including graded bedding and infilled cross-cutting channels (see Section 7.4).

- Event LCI 4: is a series of four minerogenic sediment units with diffuse boundaries identified from x-ray analysis (Section 6.2.2) at 412.5-406.5 cm, 400.5-399.5, 390.5-386.5cm and 366-364 cm. From Figure 6.17 these events also correspond to single sample increases in grain size (Section 6.2.5). This suggests that these events are significant changes in the energy of sediment deposition within the basin (see Section 7.4).
- Event LCI 5: marks the start of organic peat accumulation from 307 cm, with a sustained rise in loss-on-ignition (Section 6.2.4) from 12% at 338 cm to >70% by 307 cm. An age for the onset of peat accumulation is interpolated as *c.* 7100 BP (8100 cal. BP) (Section 6.3.8) (see Section 7.5).
- Event LCI 6: the deposition of a fine to medium silt, typical of lacustrine sedimentary processes during the accumulation of sedge peat between 254-252 cm dated to 6280 ± 60 BP (7215 cal. BP) (see Section 7.6).
- Event LCI 7: the deposition of a fine to medium silt, typical of lacustrine sedimentary processes during the accumulation of sedge peat between 200-198 cm, dated to 5000 ± 80 BP (5730 cal. BP) (see Section 7.6).
- Event LCI 8: the deposition of a fine to medium silt, typical of lacustrine sedimentary processes during the accumulation of sedge peat between 122-118 cm, dated at 3060 ± 50 BP (3300 cal. BP) (see Section 7.6).
- Event LCI 9: the deposition of a fine to medium silt, typical of lacustrine sedimentary processes during the accumulation of sedge peat between 92-88 cm, dated at 2410 ± 50 BP (2360 cal. BP) (see Section 7.6).
- The lack of control from either vegetation type or minerogenic content suggests a hydrological control for the highly variable sedge peat humification record (Section 6.2.6) (see Section 7.6). However, pollen analysis (Section 6.3.7) proved inconclusive with no change in aquatic vegetation associated with these proposed reductions in water depth. Only long term changes in mire surface wetness were palynologically recorded (see Section 7.6).

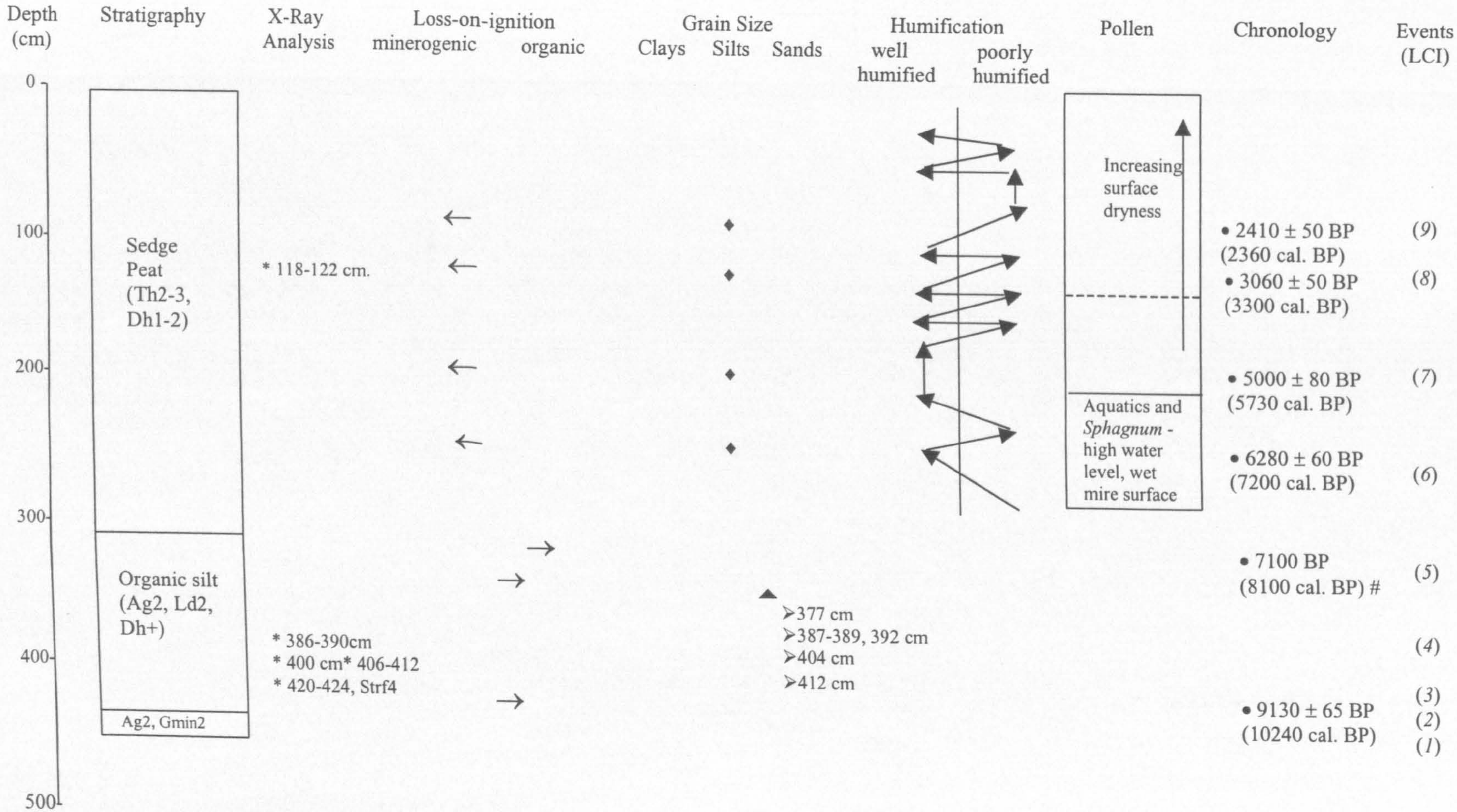


Figure 6.17 Core LCI summary of main sedimentary events. * minerogenic sediment units defined through x-ray analysis. ▲ decreasing grain size, ➢ units with increased grain size, ♦ sediments with grain size distributions similar to typical lake sediments, # interpolated radiocarbon date.

6.4 CORE LCIV (Figure 5.1)

6.4.1 SEDIMENT STRATIGRAPHY

Basal minerogenic sediments (Ag2, Gmin2) are laminated with 1-2mm thick laminae consisting of better sorted pale grey silts and darker grey sands (Table 6.10). These are overlain, with a sharp (<1 mm) lower boundary (577 cm), by a thin (2 cm) unit of grey green mica-rich silts. The accumulation of more organic-rich sediments is recorded by a further abrupt transition (575 cm) from grey/green silts to a dark brown organic-rich silt. This dark brown silt becomes increasingly detritus rich (Dh) (575-275 cm) with the accumulation of a peat (Dh3 Ld1) above 275 cm. Distinctive flattened black graminoid plant macrofossils are recorded between 429-390 cm.

Depth cm	LCIV Sediment Description
0-30	Water
30-155	Brown poorly humified fibrous sedge peat herbaceous, woody and sedge type plant fragments; Th4, Nig2, Strf0, Elas3, Sicc2, Lim1-2.
155-236	Dark brown slightly humified fibrous sedge peat; Th3 Dh1 Ld+, Nig3, Strf0, Elas2, Sicc2, Lim1-2.
236-243	Brown organic detrital silt, mica flakes visible; Ag2 Ld1, Dh1, Nig2, Strat1, Elas0, Sicc2, Lim 1-2.
243-275	Dark brown humified peat; Dh3 Ld1 Dl+ large wood fragments noted at 250cm and 270 cm. Nig3, Strf0, Elas1, Sicc2, Lim1-2.
275-325	Dark brown detrital rich organic mud, plant detrital material becoming increasingly humified; Dh2-3 Ld 1-2 Dl+, Nig3, Strf0, Elas1, Sicc2, Lim1-2.
325-355	Dark brown fine organic mud with plant detrital fragments; Ld2-3 Dh1-2 Dl+, Nig3, Strf0, Elas0, Sicc2, Lim1-2.
355-390	Dark brown fine organic mud; Ld3-4 Dh+, Nig3, Strf0, Elas0, Sicc2, Lim1-2
390-429	Dark brown fine organic mud with large (>2mm) flattened black Graminoid type plant macrofossils; Ld2 Dh2, Nig3, Strf0, Elas0, Sicc2, Lim1-2.
429-496	Dark brown fine organic mud, mica flakes visible; Ld4 Ag+, Dh+, Nig3, Strf0, Elas0, Sicc2, Lim2.
496-537	Dark brown fine organic silt; Ld3, Ag1, Nig3, Strf0, Elas0, Sicc2, Lim1-2.
537-575	Greenish brown organic rich silt, mica flakes visible; Ag2, Ld2 Dh+, Nig2, Strf1, Elas0, Sicc2, Lim3.
575-577	Dark grey green mica rich silt, mica flakes visible; Ag3, As1, Nig2, Strf1, Elas0, Sicc2, Lim3.
577-586	Silver grey coarse silts and sands, mica flakes visible; Ag2, Gmin2, Nig1, Strf4, Elas1, Sicc1.

Table 6.10 Core LCIV sediment description using a modified Troels-Smith notation (Section 5.2).

The accumulation of this detrital peat is interrupted by the deposition of organic-rich silt between 243-236 cm. Subsequent peat accumulation (236-30 cm) consists of fibrous peat, herbaceous and sedge type plant material with frequent large (>2mm) woody fragments (Dl).

Sediment analysis of sediments 577-586 cm was not undertaken because these highly thixotropic sediments could not be retained in stratigraphic sequence

6.4.2 BULK DENSITY, WATER CONTENT AND LOSS-ON-IGNITION.

Bulk density (Figure 6.18 (a)) values of 1.25 g/cm^3 at 576 cm suggests compacted minerogenic sediments. Below 254 cm the mean (n=163) bulk density value is 1.07 g/cm^3 with 1σ standard deviation of 0.05; above 254 cm the mean (n=108) bulk density value is 1.00 g/cm^3 with a 1σ standard deviation of 0.08. The increase in variability about the mean and decrease in the mean bulk density value above 254 cm reflects the accumulation of increasingly organic-rich sediment and sedge peat. The greater variability could reflect density variations within the accumulating peat or (as outlined in Section 6.2.4) is an artefact of difficulties in sampling this type of sediment.

Percentage water content (Figure 6.18b) increases at the transition from minerogenic sediments to increasingly organic-rich sediments. The remainder of the record varies little except for notable minor perturbations at 130 cm, 144 cm and 238-242 cm, suggesting increased mineral content.

Percentage loss-on-ignition values (Figure 6.18c) indicate the accumulation of increasingly organic-rich sediments from 576 cm. Sediments between 476-454 cm and 422-402 cm share a common pattern of changes in loss-on-ignition, with stepped declines followed by abrupt increases in value. This suggests for these sediments a common control on organic content (see Section 7.4). The deposition of mineral-rich sediment within the increasingly organic-rich sediment is detected by a decline in loss-on-ignition of 37% between 264 and 240 cm. The deposition of this more minerogenic unit (236-242 cm) is followed by the abrupt accumulation of organic sedge peat with percentage loss-on-ignition values increasing sharply to >75 % at 234 cm.

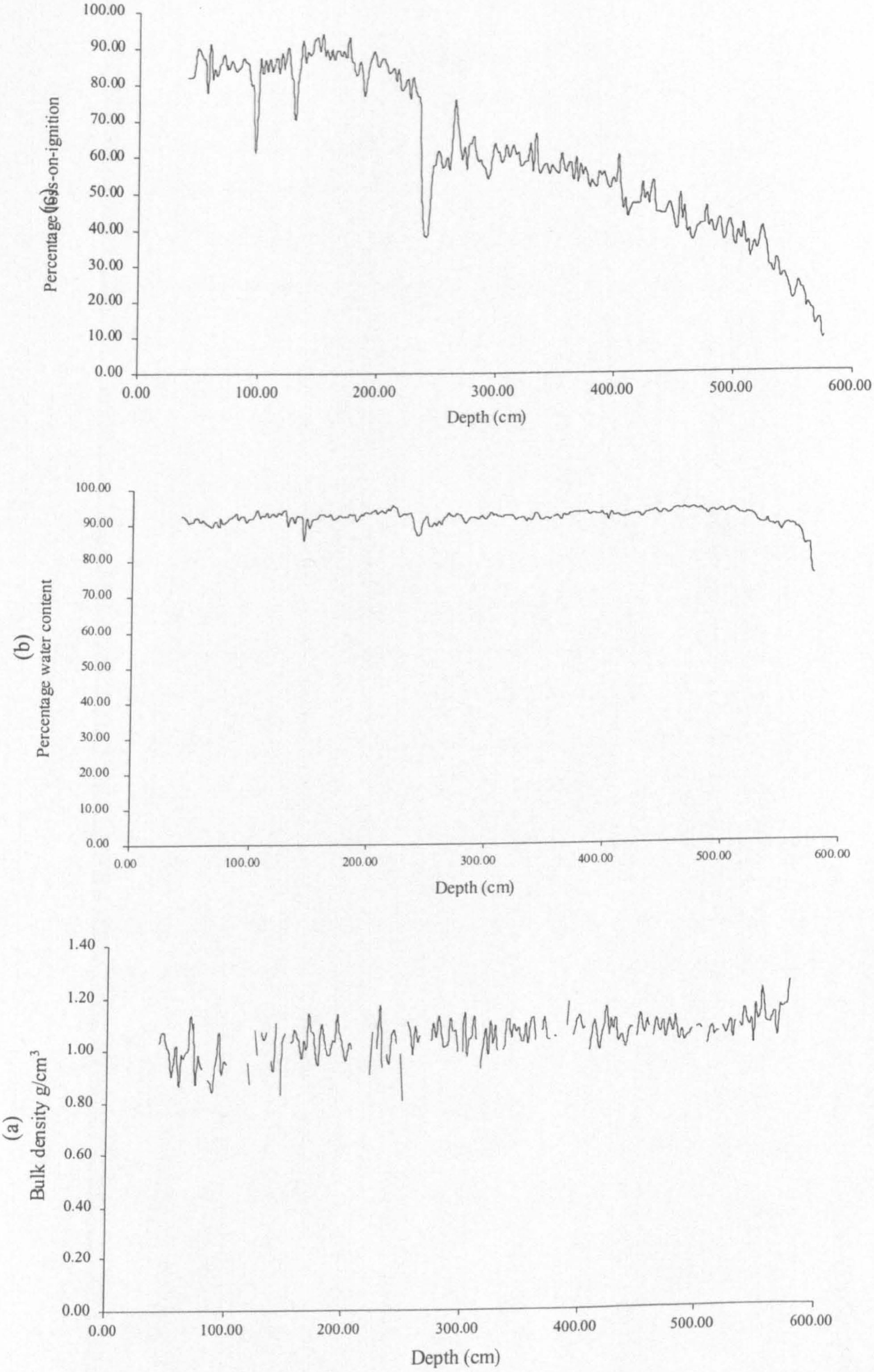


Figure 6.18 Core LCIV bulk density (a), percentage water content (b) and loss-on-ignition (c) plotted against depth.

The accumulation of these highly organic sedge peats, with a percentage loss-on-ignition value of > 84%, is interrupted again by the deposition of more minerogenic sediment units at 189-187 cm (77%), 131-129 (70%) and 99-95 cm (61-71%). These reductions in loss-on-ignition are single sample fluctuations except between 99-95 cm where one is recorded over two samples. However, the decline in loss-on-ignition within each of these more minerogenic sediments exceeds the 1σ Standard Deviation (n=97; 85 ± 4.8%) calculated for mean loss-on-ignition within the organic-rich peats (234-30 cm).

6.4.3 GRAIN SIZE ANALYSIS

The data are summarised as changes in grain-size with depth in Figure 6.19 and Table 6.11. Within the grain size data, sediments record several coarsening up or fining up sequences suggesting gradual and prolonged changes in energy of deposition (Table 6.11). During these periods of more gradual changes in grain size, abrupt changes in grain size are recorded and are described as a series of sedimentary events (Table 6.11). Many of these are single sample grain size events but could represent changes in the mode and energy of deposition (Section 7.4). Satisfactory grain size analysis could not be obtained for all sediments, in particular for organic-rich sediments (Section 5.3).

Grain size data between 414 and 256 cm are thus incomplete and it is difficult to evaluate prolonged trends within the data. In general the grain size of these sediments is similar to the unit 440-414 cm described in Table 6.11. Contiguous data obtained for sediments between 256 and 236 cm indicate increasingly coarser grained sediments, with decreasing clay (21 to 11%), very fine (18 to 10%) and fine silt (29-22%) contents and increasing coarse silt (17 to 22%), very fine sand (3 to 19 %) and fine sand (0 to 8%) with <2% medium sand recorded between 242-236 cm.

The more minerogenic units recorded within the peat at 189-187 cm, 131-129 cm and 99-95 cm (Figure 6.19) are predominantly medium (14-20%) to coarse (16-23%) silts with 10-26% very fine sand and at 129-131 cm up to 13% fine sand. These minerogenic sediment units appear to be as coarse as sediments recorded between 576-566 cm and 242-236 cm. However, the unit at 131-129 cm appears to be coarser than any other sediments within LCIV.

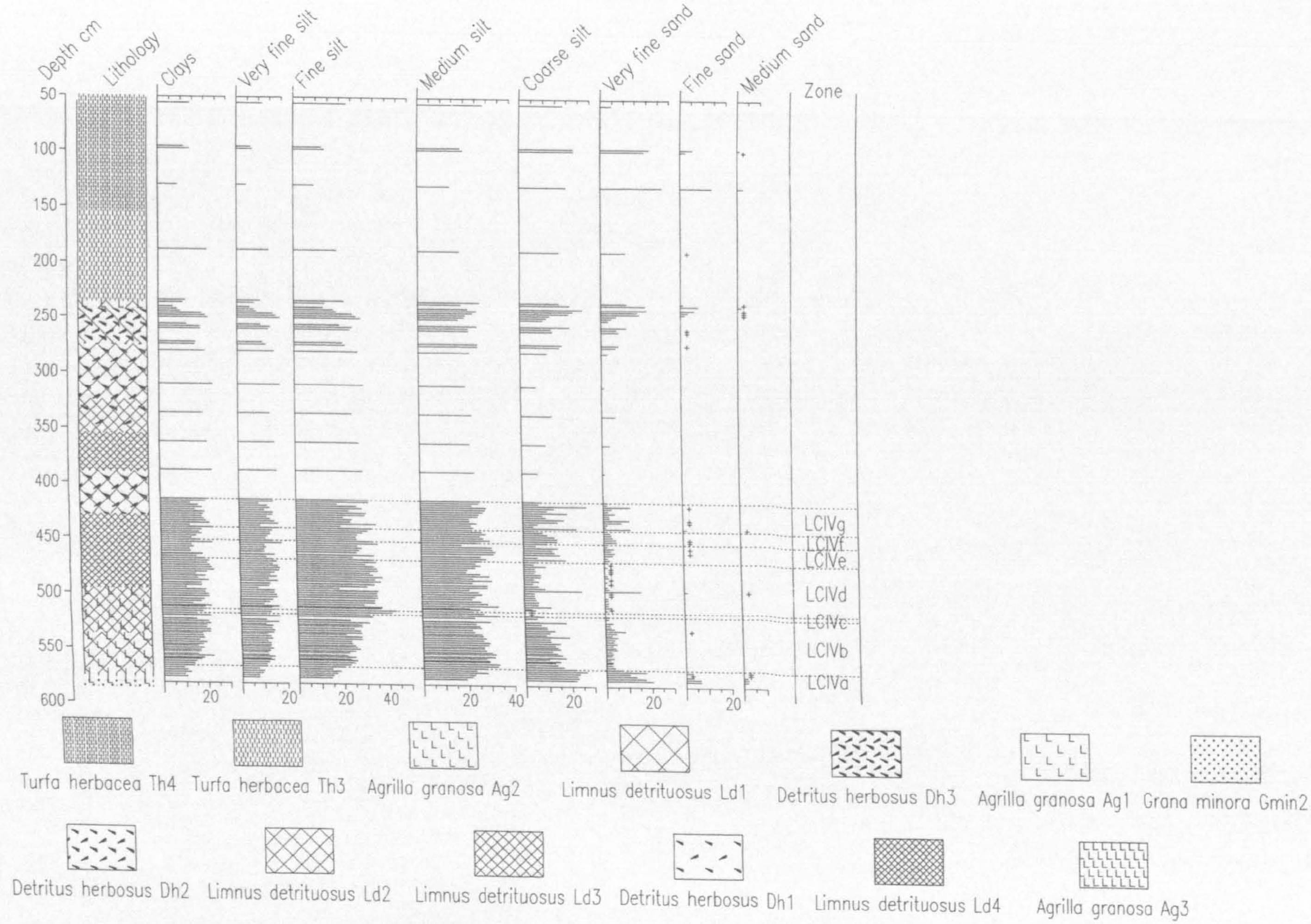


Figure 6.19 Core LCIV grain size plotted against depth.

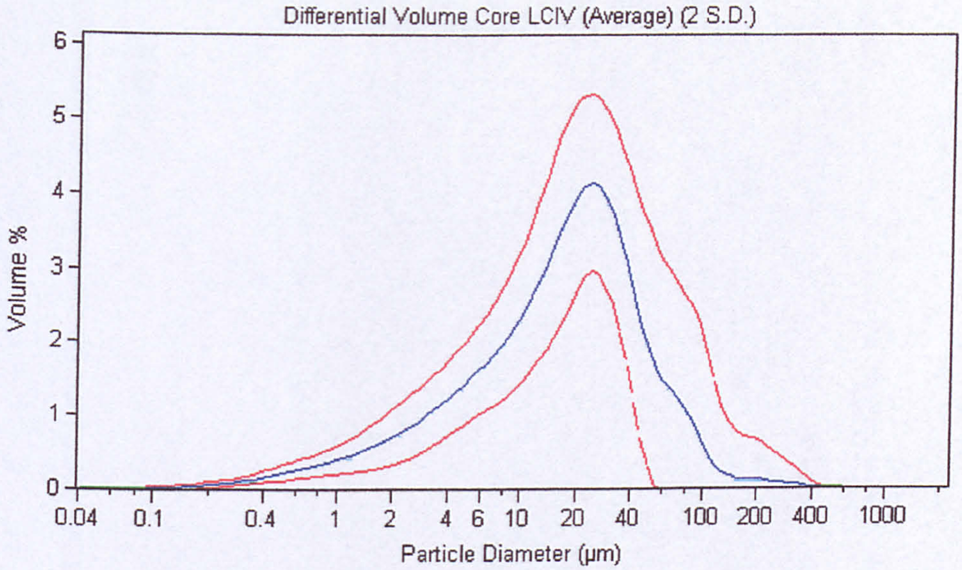
Depth (cm)	Zone	Sediment grain size description	Sedimentary Events LCIV
576-566	LCIVa	Silty sand. Fining up sequence with ↑ medium (19-27%) and fine (14-19%)silts with around 22% coarse silt. ↓ very fine (20-7 %) and fine sands (6-0%),.	
566-518	LCIVb	Fine to medium silt. ↑ fine silts (22-30%), with ↓ medium (32-22%) and coarse (16-7%) silts, with around 2-5% very fine sand.	536 cm abruptly ↑ coarse silts (19%) and at 530 cm abrupt ↑ very fine sands (8%)
518-514	LCIVc	Fine silt. Abruptly ↑ fine (43%) silts, with abruptly ↓ coarse silts (<1%) and an absence of very fine sands.	
514-468	LCIVd	Fine to medium silt. No trends within grain size but coarser than previous unit with predominantly 29-37% fine silts, 20-33% medium silts, <10% coarse silts and <2% very fine sand.	495 cm abruptly ↑ medium (21%) and coarse silts (24%) with 16% very fine sand and 1% fine sand.
468-452	LCIVe	Medium silt. Variable grain size data but coarser than previous unit, with 25-35% medium silts, 5-18% coarse silts and 1-5% very fine sand and <1% fine sand.	
452-440	LCIVf	Fining up sequence with gradually ↑ clays (13-22%), very fine (11-17%) and fine (26-33%) silts with gradually ↓ medium (30-21%) and coarse (14-4%) silts and very fine (5-1%)sand.	
440-414	LCIVg	Coarsening up sequence with↑ coarse silts and very fine sands, and is coarser than the unit below. Predominantly fine (29-35%) and medium silts (21-28%) with 4-12% coarse silt and <3% very fine sand	438 cm, 432-430 cm, 424 cm and 418cm (9) abruptly ↑ coarse silts (14-22%), very fine sands (5-11%) and with <1% fine sand.

Table 6.11 Core LCIV grains size summary for sediments 576-414 cm (Figure 6.19). ↑ denotes *increasing and ↓ denotes decreasing percentage grain size.*

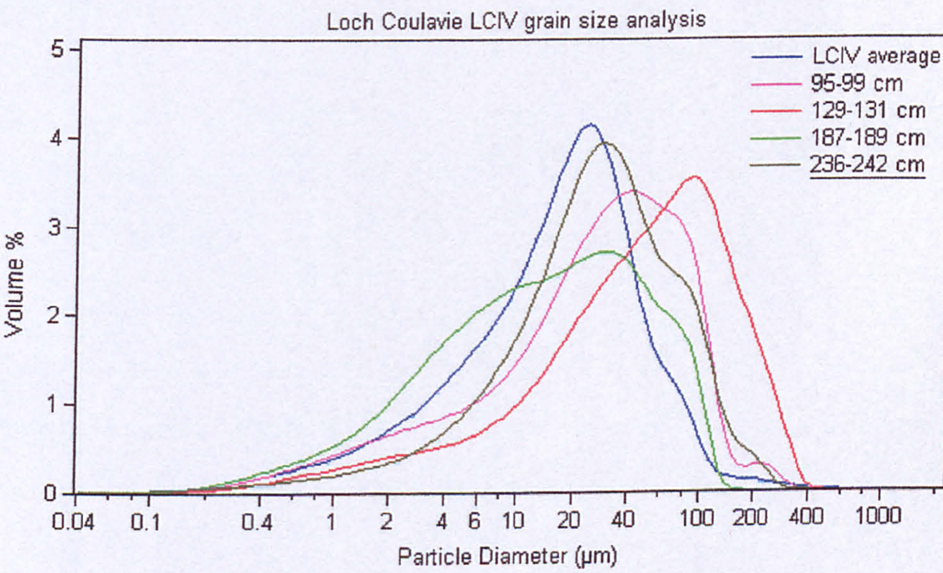
Further definition of the mode of deposition of the minerogenic units recorded at 242-236 cm, 189-187 cm, 131-129 cm and 99-95 cm (Section 6.4.5, events LCIV (11, 13-15)) was through examination of particle size distributions. Particle size distributions from these minerogenic units were compared with a mean particle size distribution of sediment considered typical of lacustrine processes. From the sediment description (Table 6.10), loss-on-ignition data (Figure 6.18 (c)) and grain size data (Figure 6.19) sediments deposited between 576-537 cm, greenish brown organic rich (<30 % loss-on-ignition) medium to coarse silts, were chosen. The particle size distributions of these sediments were averaged using the COULTER LS230 software to produce a mean (n=22) particle size distribution curve to two standard deviations, (Figure 6.20 (a)), *labelled LCIV average.* The particle size distributions for units 236-242 cm, 189-187 cm, 129-131 cm and 95-99 cm were compared with this average distribution curve Figure 6.20(b).

The particle size distributions for minerogenic units noted at 242-236 cm, 189-187 cm and 99-97 cm lie within the range of the core average. This suggests that these units share a common mode of deposition that is governed by 'normal' lacustrine processes. They are not viewed as exceptional sedimentological events. From Figures 6.19 and 6.20 (b) the grain size distribution for the minerogenic unit noted at 131-129 cm is significantly coarser which suggests that this unit may have a mode of deposition that is not primarily controlled by lacustrine processes but by higher energy processes such as fluvial or catchment erosional events (see Section 7.6).

Particle size distributions for sediment units, 236-242 cm and 99-95 cm which occur over 4 cm (2 grain size samples) are plotted in Figures 6.21 and 6.22. There was no significant variation in grain size during the deposition of each unit. This lack of sediment grading within the unit indicates no change in the energy of deposition during sedimentation and provides further evidence for a mode of deposition controlled by 'normal' lacustrine processes.



(a)



(b)

Figure 6.20 (a) and (b) Core LCIV percentage volume particle size distributions. (a) An average particle size distributions for the basal sediments (for definition see text) to a standard deviation of 2 σ. (b) particle size distributions of the units identified as reductions in loss-on-ignition (Section 6.4.5, events LCIV (11, 13-15)) and the core average particle size distribution (note the 2 σ Standard deviation for the average particle size distribution is not shown in this lower graph).

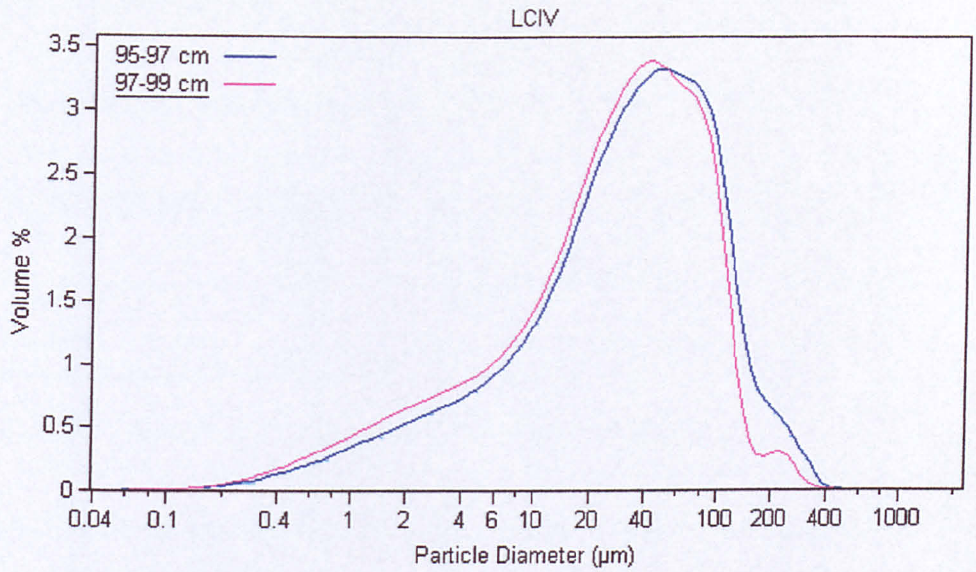


Figure 6.21 Core LCIV particle size distributions for sediments within unit 99-95 cm (event LCIV 15), see Section 6.4.3.

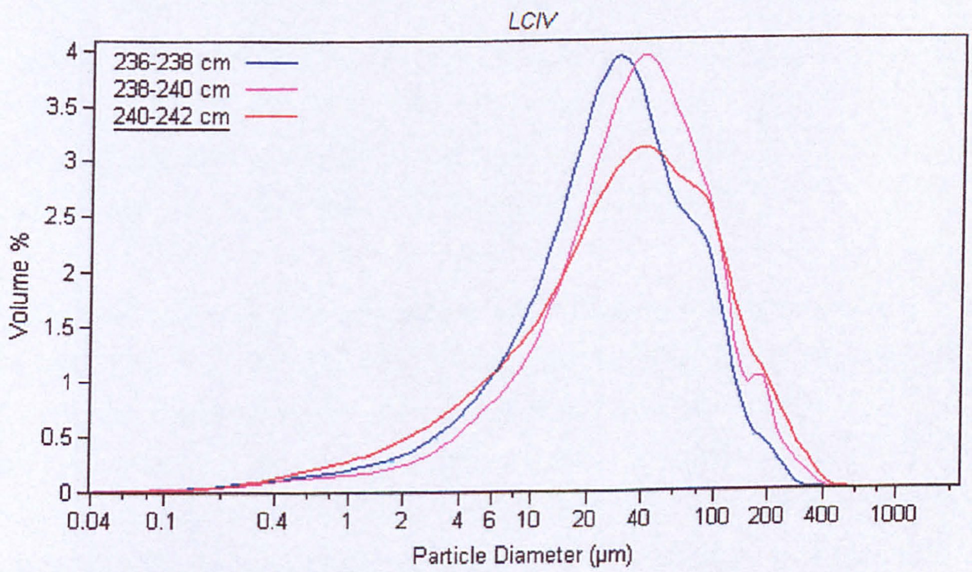


Figure 6.22 Core LCIV particle size distributions for sediment within unit 242-236 cm (event LCIV 11), see Section 6.3.3.

6.4.4 CHRONOLOGY

The chronology was defined using two AMS radiocarbon assays (Table 6.12).

Laboratory reference	Depth (cm)	C content (% by wt)	Conventional Radiocarbon age (yrs BP ± 1σ)	Calibrated age range to 1σ (yrs. BP)	Calibrated age range to 2σ (yrs. BP).
AA34228	234	39	6490 ± 55	7431-7327	7550-7273
AA34223	560	14	9015 ± 60	10222-10162	10241-9920

Table 6.12 Core LCIV AMS radiocarbon (BP) and calibrated (cal. BP) ages.

The AMS dates obtained are internally consistent and there are no grounds to reject these dates as incorrect.

Assay AA-34223 does not, however, date the earliest organic sediments but lies 17 cm above the actual start of organic sedimentation (Table 6.10; Figure 6.18 (c)). The very gradual increase in organic content made it difficult to define sediments sufficiently organic to date securely by radiocarbon. Organic sedimentation therefore predates 9015 ± 60 BP (10 200 cal. BP). Assuming constant sediment accumulation rates in the lacustrine sediments between 560 and 234 cm, organic deposits at 577 cm have an extrapolated age of approximately 9145 BP (10 250 cal. BP).

Assay AA-34228 at 234 cm may also not precisely define the onset of peat accumulation. There are uncertainties in defining the gradual transition to peat in this core. Troels-Smith description (Table 6.10) suggests that peat replaces lacustrine sediment at 275 cm, and an increased percentage loss-on-ignition at this depth (Figure 6.18 (c)) may support this interpretation, but after 260 cm increases in minerogenic content complicate the identification of this boundary. Because this minerogenic sediment may have contained re-mobilised organic matter, the radiocarbon assay was obtained from the earliest *in-situ* peat at 234 cm. Again assuming constant accumulation rates between 560 cm and 234 cm, which is unlikely given the changes in sediment type, peat inception could have occurred at around 6800 BP (7750 cal. BP) (see Section 7.5).

6.4.5 SUMMARY OF STRATIGRAPHIC EVENTS

Sediment analysis has defined several significant stratigraphic events within core LCIV and these are summarised in Figure 6.23.

- Event LCIV 1: deposition of laminated minerogenic sediments between 586-577 cm (Table 6.10) (see Section 7.2).
- Event LCIV 2: marks the onset of organic sedimentation at 577 cm as defined from sediment description (Section 6.4.1) and loss-on-ignition analysis (Section 6.3.2), with a gradual transition to increasingly organic-rich sediments, from 9% loss-on-ignition at 577 cm to >18% at 560 cm. The onset of organic sediment accumulation is dated as *c.* 9145 BP (10 250 cal. BP) (Section 6.4.4) (see Section 7.3).
- Events LCIV 3-10: are a series of abrupt (often single sample) changes in grain size (Section 6.4.3, Table 6.11) at 536 cm, 530 cm, 518-514 cm, 495 cm, 438 cm, 432-430 cm, 424 cm and 418 cm. These abrupt events occur during more prolonged coarsening or fining up grain size sequences (Table 6.11). This suggests that these more abrupt fluctuations in depositional energy occur during periods of more gradual change in the energy of deposition (see Section 7.4). There appears to be no correlation between these events and changes in organic content (Section 6.4.2, Figure 6.18(c)).
- Event LCIV 11: marks the deposition of a thick (7.0 cm) unit of organic-rich silty sand at 242-236 cm as defined from sediment description (Section 6.3.1), loss-on-ignition, percentage water content (Section 6.4.2) and grain size analysis (Section 6.4.3). Grain size analysis (Section 6.4.3) suggests that the unit was the culmination of a coarsening up sequence (Figure 6.19), but with a similar particle size distribution to basal lacustrine sediments (Figure 6.20 (b)) and no sediment grading within the unit (Figure 6.22), and may have been deposited by predominantly lacustrine processes (see Section 7.6).
- Event LCIV 12: marks the start of *in-situ* peat accumulation at 234 cm as defined from loss-on-ignition analysis and a rapid increase in organic content with <40% loss-on-ignition at 238 cm and >75% at 234 cm (Section 6.4.2). Radiocarbon assay suggest that this *in-situ* peat accumulated from 6490 ± 55 (7420 cal BP). However,

sediment description (Table 6.10) and loss-on-ignition (Figure 6.18(c)) indicates that peat accumulation could have started earlier from around 6800 BP (7750 cal. BP) (see Section 7.5).

- Event LCIV 13: is a thin (2 cm) minerogenic-rich silty very fine sand deposited at 189-187 cm during peat accumulation, as defined from loss-on-ignition (Section 6.4.2) and grain size analysis (Section 6.4.3). Grain size analysis (Figure 6.20 (b)) suggests that sediments within event LCIV13 were deposited by predominantly lacustrine sedimentary processes (see Section 7.6).
- Event LCIV 14: is a thin (2 cm) minerogenic rich silty very fine sand deposited at 131-129 cm during peat accumulation, as defined from loss-on-ignition (Section 6.3.2) and grain size analysis (Section 6.4.3). Grain size analysis (Figure 6.20 (b)) suggests that this event is significantly coarser, indicating a higher energy mode of deposition than lacustrine processes, such as fluvial processes. This event is also recorded as a reduction in percentage water content (Figure 6.18(b)) indicating an increase in minerogenic content (see Section 7.6).
- Event LCIV 15: is a thin (4 cm) minerogenic rich silty very fine sand deposited at 99-95cm during peat accumulation, as defined from loss-on-ignition (Section 6.4.2) and grain size analysis (Section 6.4.3). Grain size analysis suggests no sediment grading within the unit (Figure 6.21) and that this event was deposited by predominantly lacustrine sedimentary processes (Figure 6.20(b) (see Section 7.6).

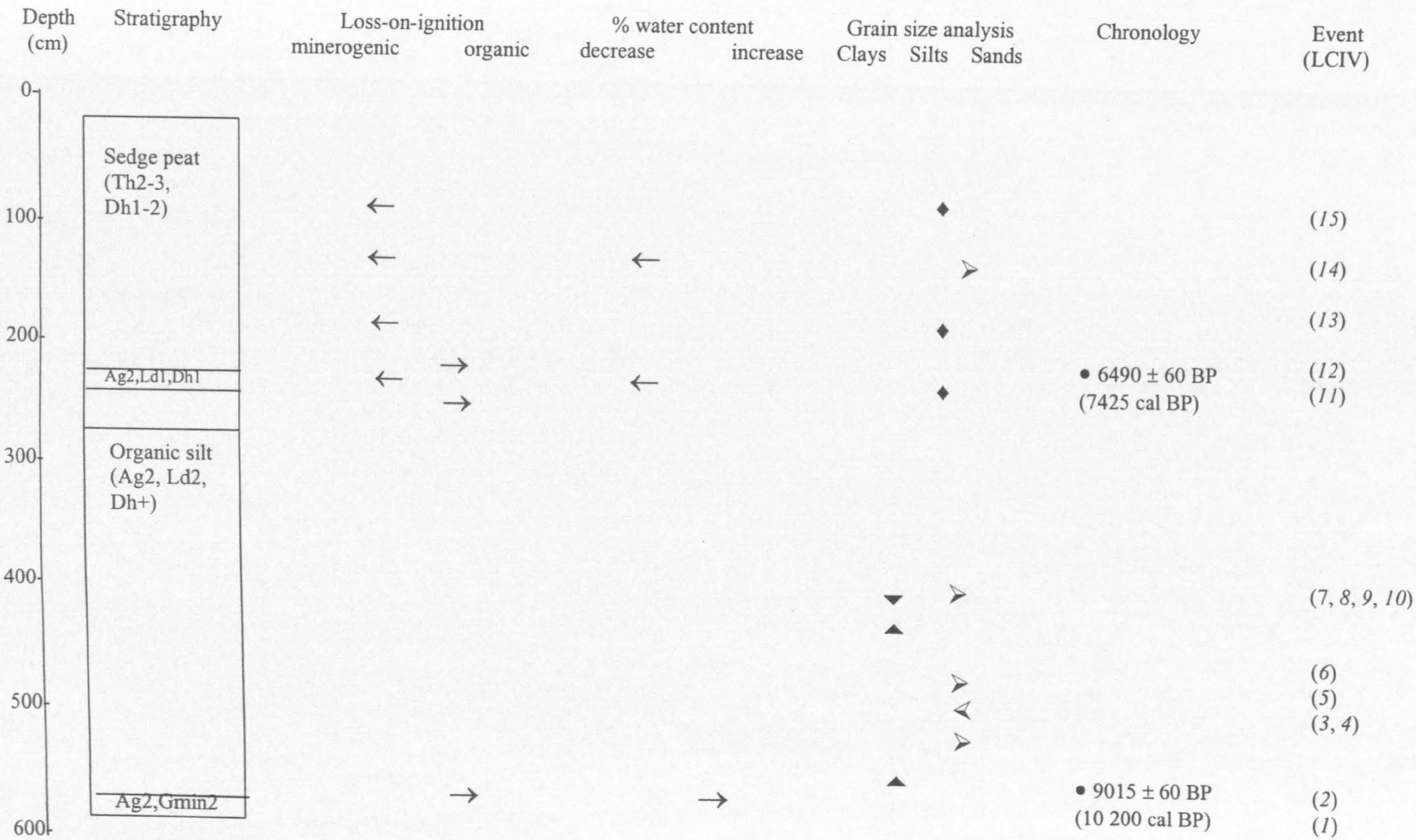


Figure 6.23 Core LCIV sediment analysis summary of main stratigraphic events. ← & → denote decreasing and increasing values ≺ units with decreased grain size, ≻ units with increased grain size, ▲ finning up grain size sequence, ▼ coarsening up grain size sequence ♦ sediments with grain size distributions similar to typical lake sediments.

6.5 CORE LCXII (Figure 5.1)

6.5.1 SEDIMENT DESCRIPTION

Basal sediments (707-688 cm), with laminae 1-2mm thick consisting of better sorted pale grey silts and darker grey sands, are overlain sharply (<1mm) by greenish brown organic-rich silts (688-620 cm) representing the start of organic sedimentation (Table 6.13). The subsequent accumulation of organic silt (620-150 cm) with increasing amounts of organic detrital material (Dh) is interrupted by the deposition of a series of sandy sediment units. Units recorded at 599-597 cm, 541-539 cm, 515-513 cm, 391-389 cm, and 273-271 cm are stratigraphically well-defined as thin (1-2 cm thick), pale brown sands. Minerogenic sediments recorded between 491-484 cm form a fining up sequence of medium-coarse sands (491-486 cm), coarse silts (486-485 cm) and silts and clays (485-484 cm). The final unit is a coarse, fibrous detritus-rich peat, above 150 cm.

	LCXII
Depth cm	***Sediment Description***
0-100	water
100-150	Dark brown coarse fibrous detritus rich peat; Dh3 Ld1, Nig3, Strf0, Elas2-3, Sicc2, Lim1-2.
150-271	Dark brown organic-rich mud with detrital herbaceous and woody plant fragments; Ld2-3, Dh1-2, Dl+, Nig3, Strf0, Elas2, Sicc2, Lim3.
271-273	Pale brown sand; Gmin4, Nig1-2, Strat2, Elas0, Sicc2, Lim3.
273-389	Dark brown organic-rich mud with detrital herbaceous and woody plant fragments; Ld2-3, Dh1-2, Dl+, Nig3, Strf0, Elas2, Sicc2, Lim3.
389-391	Pale brown sand; Gmin4 Nig1-2, Strf2, Elas0, Sicc2, Lim3.
391-396	Dark brown organic-rich mud with detrital herbaceous plant fragments; Ld2-3, Dh1-2 Nig3, Strf0, Elas2, Sicc2, Lim1-2.
396-484	Dark brown silt mica flakes visible and rare detrital plant fragments; Ld2, Ag2 Dh+, Nig3, Strf0, Elas0, Sicc2, Lim3.
484-485	Grey fine silt; As2, Ag2Nig2, Strf1, Elas0, Sicc2, Lim2-3
485-486	Dark grey silt; Ag4, Nig2-3, Strf1, Elas0, Sicc2, Lim3.
486-491	Pale brown coarse sand; Gmin3, Gmaj1, Nig2, Strf2, Elas0, Sicc3, Lim3.
491-513	Dark brown organic-rich silt; Ld2, Ag2, Dh+, Nig3, Strf0, Elas0, Sicc2, Lim3.
513-515	Pale brown sand; Gmin4, Nig1-2, Strf2, Elas0, Sicc2, Lim3.
515-539	Dark brown organic-rich silt; Ld2, Ag2, Dh+. Nig3, Strf0, Elas0, Sicc2, Lim3.
539-541	Pale brown sand; Gmin4, Nig1-2, Strf2, Elas0, Sicc2, Lim3.
541-597	Dark brown organic-rich silt, detrital plant fragments noted to be woody and fibrous; Ld2, Ag2, Dh+ DL+ Nig3, Strf0, Elas0, Sicc2, Lim3.
597-599	Pale brown sand; Gmin4, Nig1-2, Strf2, Elas0, Sicc2, Lim3.
599-620	Dark brown organic-rich silt; Ld2, Ag2, Dh+ Nig3, Strf0, Elas0, Sicc2, Lim2-3.
620-688	Greenish brown organic-rich silt; Ag3, Ld1 Dh+, Nig2-3, Strf1, Elas0, Sicc2, Lim3.
688-707	Silver grey coarse silts and sands, mica flakes visible; Ag2, Gmin2, Nig1, Strf4, Elas0, Sicc2.

Table 6.13 Core LCXII sediment description using a modified Troels-Smith notation (Section 5.2.1).

The basal laminated minerogenic sediments (688-707 cm) were thixotropic and were not retained.

6.5.2 WATER CONTENT AND LOSS-ON-IGNITION.

Difficulties in sampling accurately meant that bulk density could not be reliably established (Section 5.2.4).

Variations in percentage water content (Figure 6.24a) for core LCXII appear to reflect changes in minerogenic content of sediment with small but abrupt reductions in water content at 599-597 cm, 541-539 cm, 521-519 cm, 515-513 cm, 485-493 cm, 483-481 cm, 443-435 cm, 391-389 cm, 273-271 cm, 197-195 cm and 131-123 cm, corresponding to the minerogenic bands noted in the sediment description (Table 6.13) and fluctuations in loss-on-ignition (6.24b).

Organic content is highly variable (Figure 6.24b, c). An initial increase in organic content from 10% at 688 cm to 26% at 682 cm represents the start of organic sedimentation. Basal sediments (682-622 cm) have a fairly consistent organic content between 25-33% loss-on-ignition. The upper boundary of these organic-rich silts is well-defined with a rapid increase over 8.0 cm in organic content to >45% at 614 cm. The series of sand units described in Table 6.13 at 599-597 cm, 541-539 cm, 515-513 cm, 493-485 cm, 391-389 cm and 273-271 cm is recorded as abrupt reductions in percentage loss-on-ignition. Many of these reductions are single sample events, but correlation with sediment description and percentage water content (see above) suggests that they are significant increases in minerogenic content. Figure 6.24b shows substantial reductions in organic content not defined from sediment description (Table 6.13). Significant excursions from the mean percentage loss-on-ignition for lacustrine sediment between 614-194 cm (n=216, 37.1 $\pm$ 6.35%) are defined by their exceeding the 1σ Standard Deviation, (Figure 6.24c). Sediment units discussed above are recorded as significant reductions in loss-on-ignition with further units at 445-435 cm, 375-373 cm, 357-353 cm, 225-223 cm and 199-195 cm (Figure 6.24c).

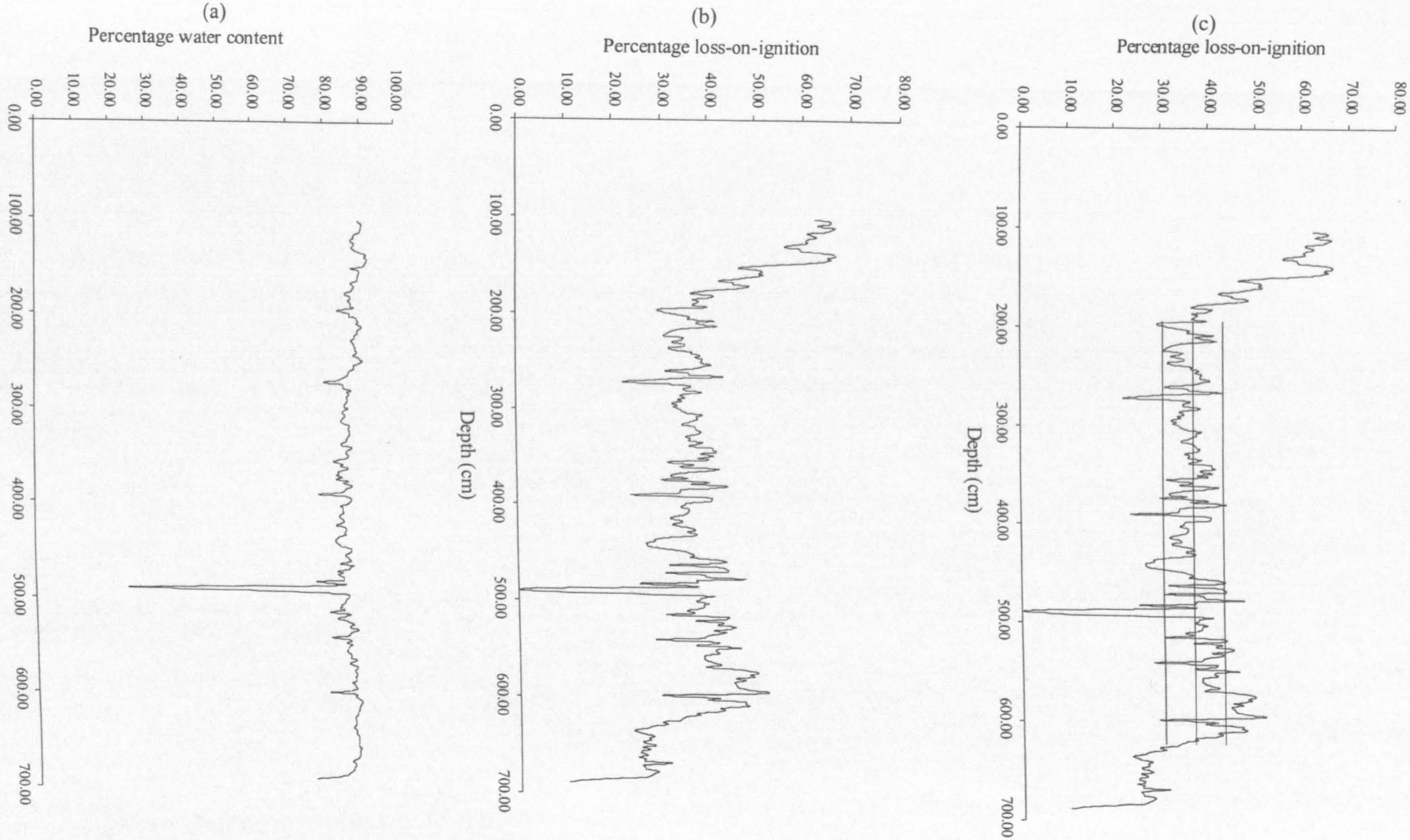


Figure 6.24 (a, b & c) Core LCXII percentage water content (a) and percentage loss on ignition (b and c) plotted against depth (cm). Graph (c) plots the mean loss-on-ignition value with the upper and lower standard deviation values, (see text).

Sediments of significantly higher organic content are recorded at 609-601 cm, 595-571 cm, 477-475 cm and 381-379 cm (Figure 6.24c).

A sustained rise in loss-on-ignition values above 194 cm indicate an increase in organic content with values exceeding 65 % at 144 cm. During peat accumulation there are variations in loss-on-ignition, with a reduction over 8.0 cm from 63% to 56% between 136 cm and 128 cm.

6.5.3 GRAIN SIZE ANALYSIS

Grain size data for core LCXII are incomplete with small sample sizes generating inaccurate grain size data (Section 5.3). Continuous data between 688-614 cm are zoned (Figure 6.25) and summarised in Table 6.14. For the remainder of core LCXII, grain size data for minerogenic units identified from sediment description (Table 6.13) and loss-on-ignition (Figure 6.24 b, c) are described in Table 6.14 (labelled as a series of events LCXII 4-11, see Section 6.5.6). Grain size data generated for sediment units at 375-373 cm, 357-353 cm and 225-223 cm (identified as significant reductions in loss-on-ignition, Section 6.5.2) were considered inaccurate owing to inadequate amounts of sample (Section 5.3).

The minerogenic sediment units outlined in Table 6.14 (events LCXII 4-11) were further characterised by examining their particle size distributions. Particle size distributions were compared with an average particle size distribution (mean and 2σ standard Deviations) for limnic sediments considered representative of uniform lacustrine sedimentation within this core. Sediments between 686-614 cm (n=36) have a fairly consistent loss-on-ignition value with no significant fluctuations in organic or minerogenic content (Figure 6.24b) and are well defined within the sediment stratigraphy (Table 6.13). These sediments, seen as 'normal' or typical lacustrine sediments unperturbed by major sedimentological events, are labelled LCXII average in Figure 6.26a. The grain size distribution curves of minerogenic units (events LCXII 4-11) described in Table 6.14 were compared with the typical lake sediment distribution curve (Figures 6.26b, 6.27 and 6.28).

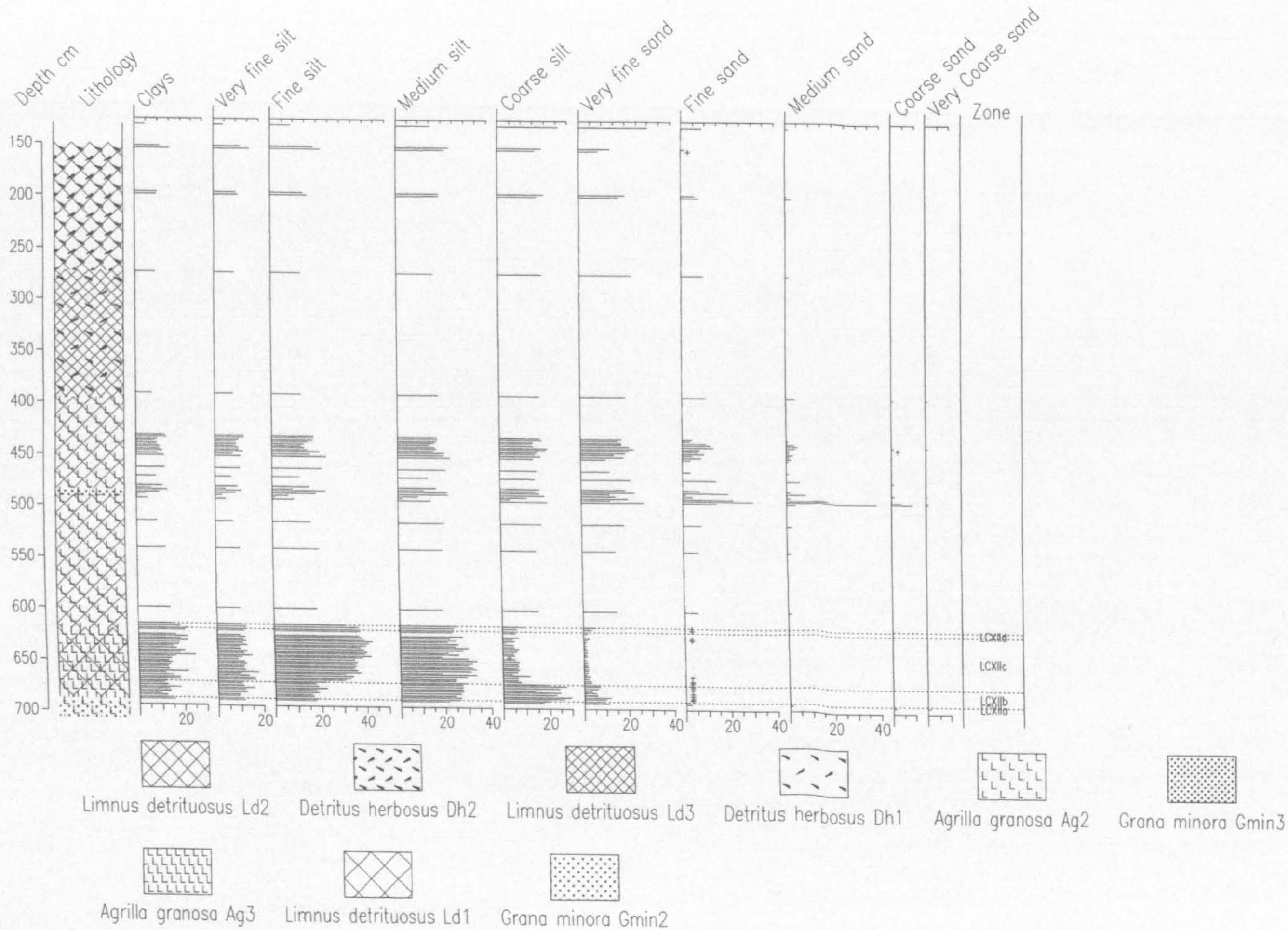
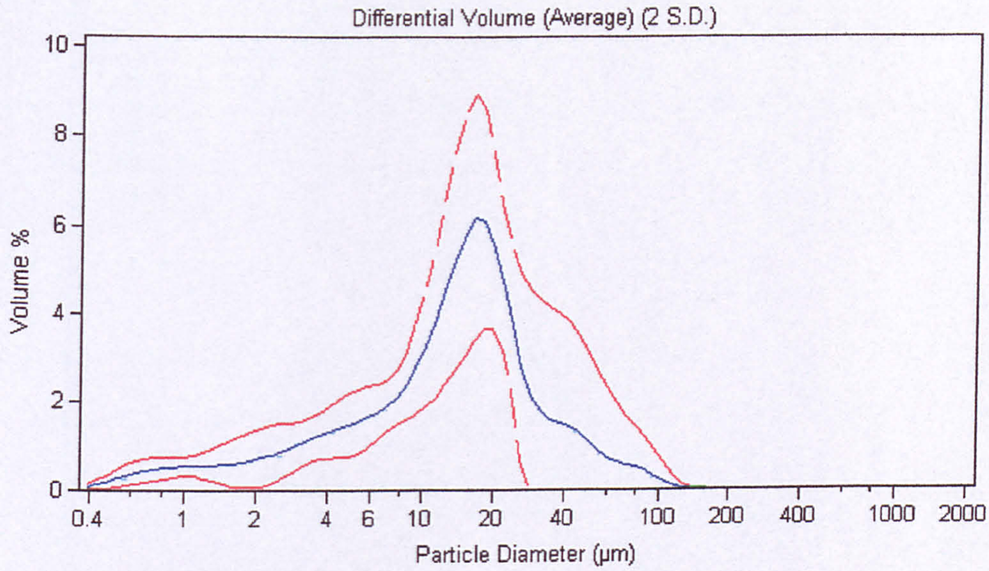


Figure 6.25 Core LCXII grain size plotted against depth

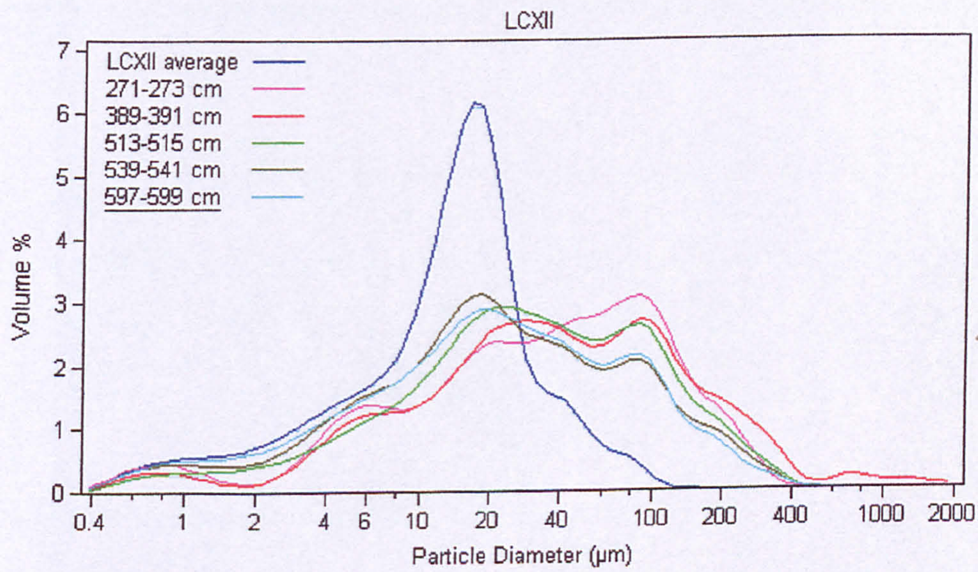
Depth (cm)	Sedimentary event	Grain size description
688-686 (LCXIIa)		Medium (24-33%) to coarse (20-29%) silty sands with 8-10% very fine sand and 1-1.5 % medium, coarse and very coarse sands.
686-670 (LCXIIb)		Fining up sequence with ↓ very fine sands (11-6%) but with similar amounts of fine silt (16-22%) and medium silts (27%) and variable coarse silts.
670-618 (LCXIIc)		Fining up sequence marked by an initial abrupt ↑ fine silts (20%->30%) and ↓ coarse silts (20%-<7%), very fine sands with no fine sands. Subsequent accumulation of fine to medium silts.
618-614 (LCXIId)		Coarser sediments with small ↑ in very fine sands (3-5%) and < 1% fine sand.
599-597	(4)	Silty very fine sand (15% coarse silt, 15% very fine sand, 5% fine sand, <2% medium sand).
541-539	(5)	Silty very fine sand (15% coarse silt, 15% very fine sand, 6% fine sand, <2% medium sand).
515-513	(6)	Silty very fine sand (17% coarse silt, 19% very fine sand, 8% fine sand, <2% medium sand)
493-491	(7)	Silty fine sand (20% coarse silt, 27% very fine sand, 14% fine sand, <4% medium sand)
491-489		Medium sand (12% very fine sand, 30% fine sand, 38% medium sand, 9% coarse sand, <1% very coarse sand)
489-487		Medium-fine sand (20% very fine sand, 22% fine sand, 17% medium sand, 5% coarse sand, <1% very coarse sand).
487-485		Silty very fine sand (19% coarse silt, 15% very fine sand, 4% fine sand, <2% medium sand)
445-435	(8)	Silty fine sand (14-20% coarse silt, 17-23% very fine sand, 7-13% fine sand, <5% medium sand, <1% coarse sand)
391-389	(9)	Silty fine sand (17% coarse silt, 19% very fine sand, 10% fine sand, <5% medium sand, <1% coarse and very coarse sand)
273-271	(10)	Silty very fine sand (18% coarse silt, 22% very fine sand, 9% fine sand, <1% medium sand)
199-195	(11)	Silty very fine sand (16-19% coarse silt, 17-20% very fine sand, 5-7% fine sand, <2% medium sand)

Table 6.14 Core LCXII Grain size summary of minerogenic sediments defined from sediment description (Table 6.13) and loss-on-ignition (Figure 6.24b,c). For zones LCXII a-d, see Figure 6.25.

The distribution curves for all the minerogenic units (events LCXII 4-11) lie outside the 2σ Standard Deviation curve for core LCXII. The minerogenic units are coarser and more poorly sorted than sediments characteristic of lacustrine processes. This suggests that these minerogenic units have a common mode of deposition but one that is not predominantly lacustrine in origin, such as catchment fluvial processes or surface erosion. The mode of deposition for these units is further discussed in Section 7.7.



(a)



(b)

Figure 6.26 (a) and (b) Core LCXII percentage volume particle size distributions. (a) Average particle size distribution for minerogenic sediments typifying lake sedimentation within core LCXII (for definition see text) to a standard deviation to 2σ. (b) overlays of particle size distributions for minerogenic units, events LCXII (4), (5), (6), (9) & (10) with the particle size distribution of typical minerogenic lake sediments (note the 2σ standard deviation is not shown in this graph). Note within LCXII no PIDS particle size data generated (Section 5.3).

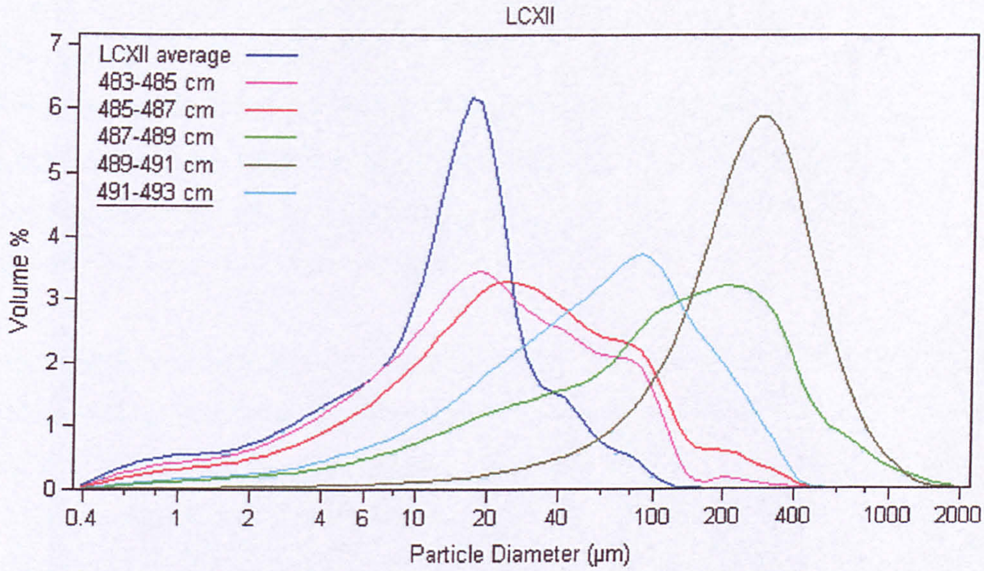


Figure 6.27 Core LCXII percentage volume particle size distributions. Overlays of particle size distributions for the stratigraphic sequence minerogenic units making up event LCXII (7) with the particle size distribution of typical minerogenic lake sediments (LCXII core average) (note the 2σ standard deviation is not shown, see Figure 6.26 (a)). Note no PIDS particle size data generated for core LCXII (Section 5.3).

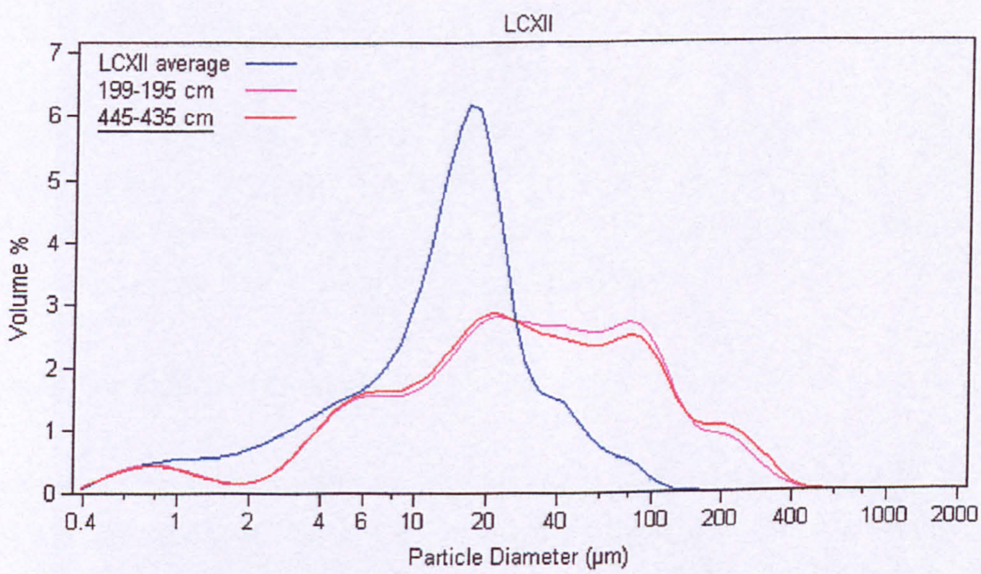


Figure 6.28 Core LCXII percentage volume particle size distributions. Overlays of particle size distributions for the minerogenic units noted from loss-on-ignition analysis, events LCXII (11) & (8) with the particle size distribution of typical minerogenic lake sediments (note the 2σ standard deviation is not shown, see Figure 6.26 (a)). Note no PIDS particle size data generated for core LCXII (Section 5.3).

Particle size distribution curves of the unit 493-485 cm (Table 6.14, event LCXII7) (Figure 6.27), suggest sediment grading indicative of changing depositional energy. Sediments at 493-491 cm consist of well sorted, silty fine sands coarsening up into very well sorted medium sands at 491-489 cm indicating increasing energy of deposition. After the deposition of these coarse sands, sediments fine up into well sorted medium-fine sands (489-487 cm) and into more poorly sorted silty very fine sands at 487-485 cm and 485-483 cm indicating a decreasing energy of deposition.

The particle size distribution curves for sediment units at 599-597cm, 541-539cm, 515-513cm, 445-435 cm, 391-389 cm and 199-195 cm (Table 6.14, events LCXII 4-6, 8, 9 and 11, Figures 6.26b, 6.28) are similar indicating that these poorly sorted silty fine and very fine sands share a common mode of deposition. Sediment unit 273-271 cm (Table 6.14, event LCXII 10, Figure 6.26b) is slightly coarser than the above units indicating a higher energy of deposition.

The minerogenic unit between 136-128 cm deposited during the accumulation of peat (Section 6.3.3) consists of a coarse silty very fine sand (Figure 6.25).

6.5.4 POLLEN ANALYSIS

Pollen analysis (Section 5.5) using a local and total pollen sum was used to identify changes in local aquatic and regional vegetation during sediment accumulation within a deeper water environment. The following sections summarise the main vegetation changes generated from the local and regional pollen sums.

6.5.4.1 Local Lacustrine Indicators

Interpretation of changes in aquatic vegetation through the generation of a local pollen sum and the use of water depth sensitive taxa is discussed in Section 5.5. The local pollen percentage pollen diagram (Figure 6.29) has a probable pollen source area consisting of the basin and basin margins (Section 5.5). The low amounts of aquatic pollen (< 2%) make it very difficult to draw any conclusions on how fluctuations in lake-level would have influenced aquatic vegetation.

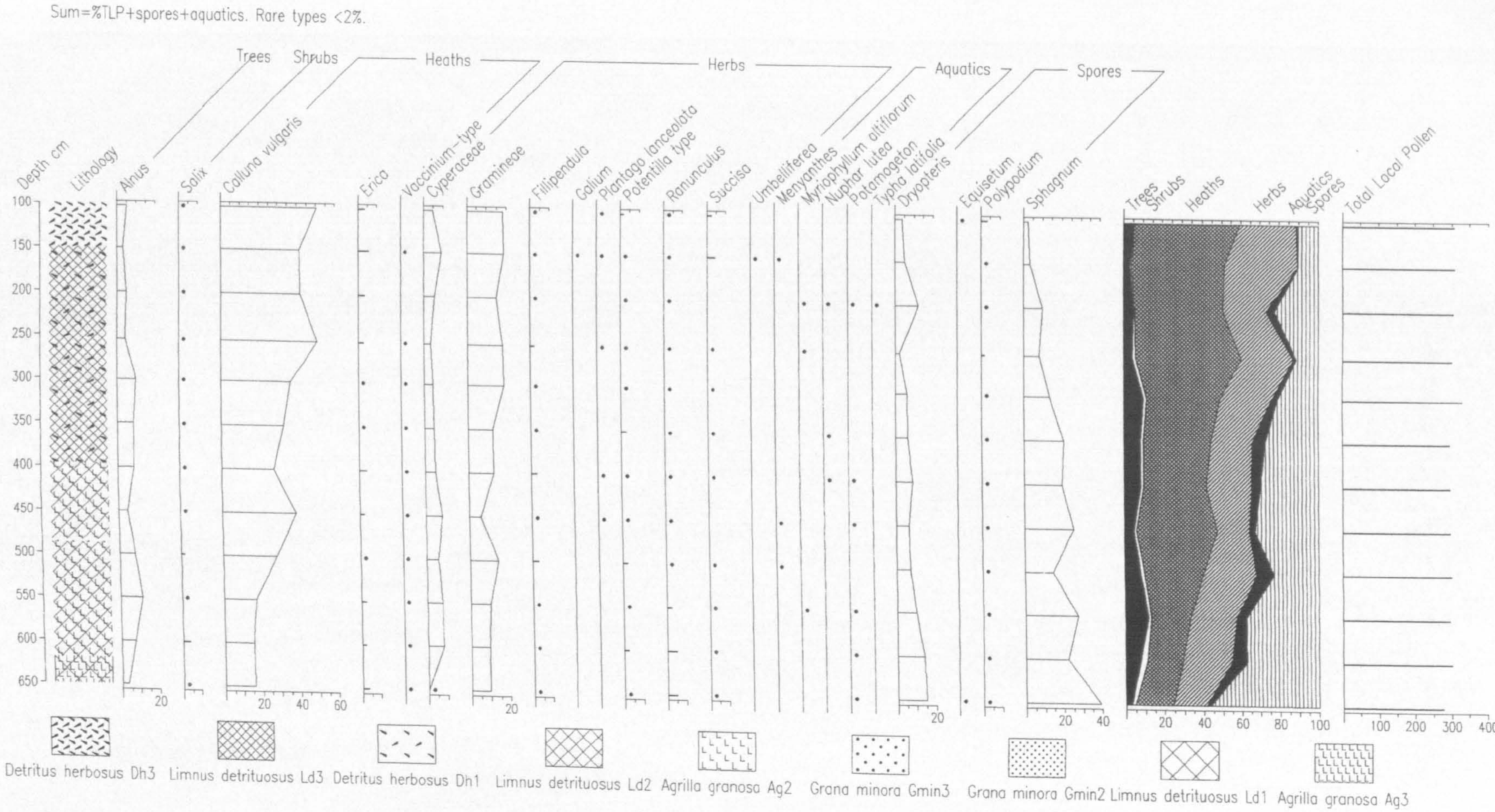


Figure 6.29 Core LCXII Local pollen percentage diagram

The continuous presence of aquatic species associated with a wide range of water depths (Table 5.5) could suggest that fluctuations in water level within the basin at this point were not of a great enough magnitude to cause significant and sustained changes in aquatic vegetation. Few trends are apparent, save a gradual unidirectional trend with the replacement of *Sphagnum* by heath taxa (550-300 cm), in particular *Calluna vulgaris* and other Ericaceae species.

The pollen corrosion data for key taxa *Betula* and *Corylus* (see Section 5.5) are shown as a bar graph in Figure 6.30. Basal sediments at 600cm and 650 cm contain exceptionally high amounts of corroded *Betula* and *Corylus* pollen. High proportions of corroded pollen can be associated with increased amounts of aerobic bacterial activity, and may relate to periods of sediment exposure to air (Section 5.5); see Section 7.4.2. Above this there are no clear trends in the data.

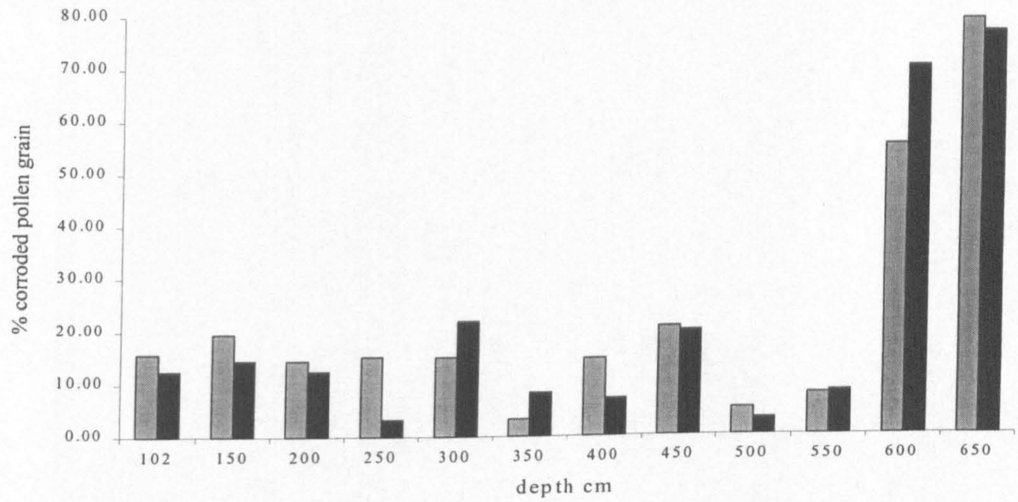


Figure 6.30 Bar graph of LCXII pollen corrosion data. Percentage of corroded *Betula* (grey bars) and *Corylus* (black bars) pollen recorded during pollen analysis (Section 5.5)

6.5.4.2 Regional Vegetation Change

The total pollen sum (Figure 6.31) is used to reflect more regional pollen recruitment (Section 5.5). Vegetation changes determined from this pollen sum were used to cross-check the radiocarbon chronology (Section 6.5.5) through a comparison with known-age biostratigraphic boundaries from a palynological record within the Loch Coulavie catchment, Torran Beithe (Davies 1999; Section 4.7 and Section 5.6).

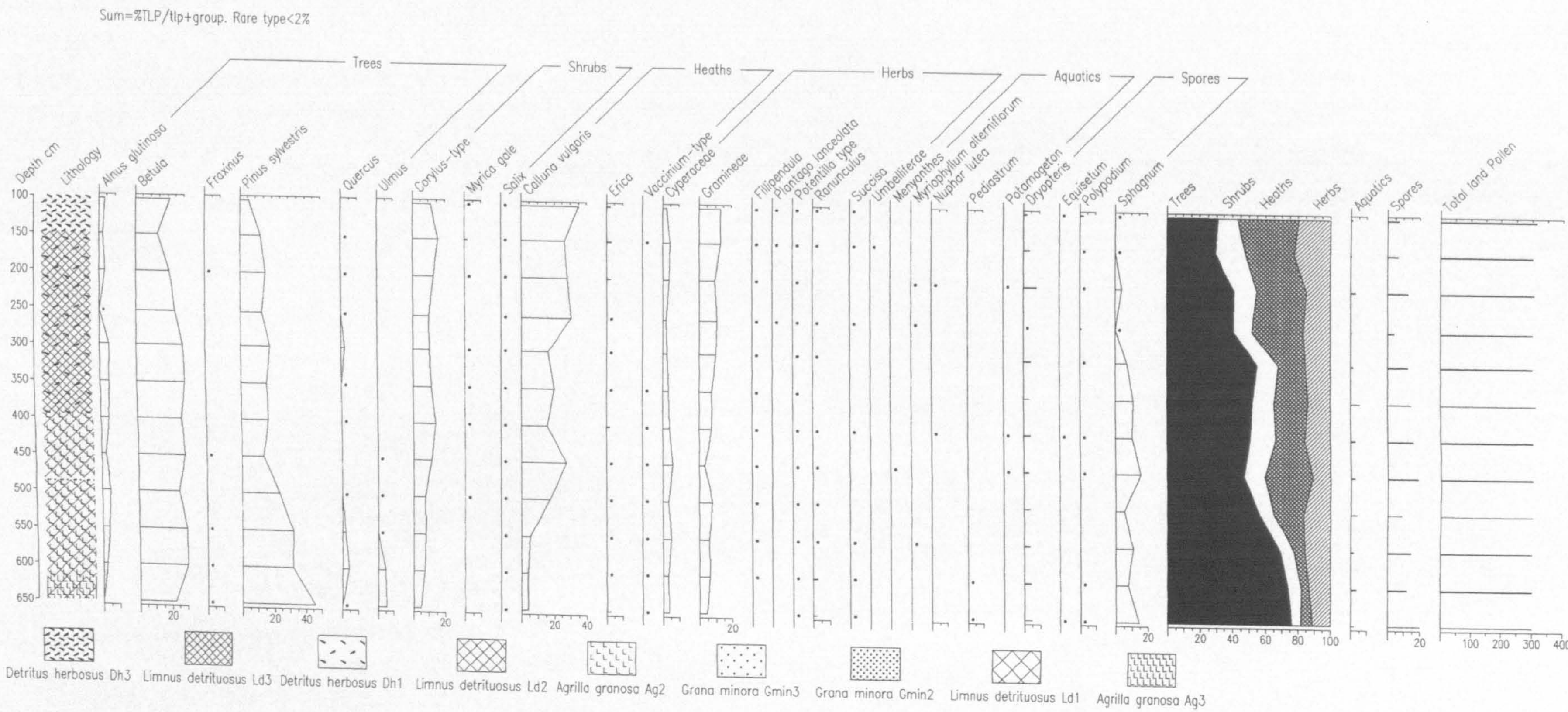


Figure 6.31 Core LCXII Total pollen percentage diagram

Vegetation changes are poorly defined at LCXII owing to the large sampling interval (50cm) and the chronological framework generated in Table 6.15 is thus restricted and approximate. The chronology presented in Table 6.15 suggests that sediments above 650 cm are younger than c. 6600 BP (7400 cal. BP). Further implications of this chronology are discussed in Section 6.5.5.

Vegetation Change	**Depth (cm)**	**Inferred chronology C14 years BP (Calibrated years BP)**
Abrupt reduction in *Ulmus*	550	*c.* 5240 BP(5970 cal. BP)
Gradually increasing *Alnus*, but *Alnus* never dominant so tentative indication of *Alnus* rise.	600-650	*c.* 6300 BP (7200 cal. BP)
Pinus sylvestris dominant with *Betula*, *Ulmus* and *Corylus avellana* type.	650	Older than *c.* 6550 BP (7390 cal. BP)

Table 6.15 Core LCXII inferred chronology for vegetation changes (Figure 6.27), based on chronologically well defined regional bio-stratigraphic data (Davies 1999; Section 5.6)

6.5.5 CHRONOLOGY

The chronology was further defined from a series of five AMS radiocarbon ages (Section 5.6). Radiocarbon ages were obtained to define (a) the start of organic sedimentation (682 cm) and (b) to define the timing of coarse minerogenic sediment units deposited within the lacustrine sediments as identified from sediment description (Table 6.13) and loss-on-ignition (events LCXII 4, 5, 6 and 7). To avoid contamination from reworked carbon, samples for radiocarbon assay were taken from sediment lying above these minerogenic units (Section 5.6). Results in radiocarbon years BP and calibrated years BP are shown in Table 6.16 and as an age-depth curve in Figure 6.32.

The radiocarbon assays are internally consistent and lie in the correct stratigraphic order. The regional pollen chronology generated in Section 6.5.4.2 indicated that sediments lying above 650 cm are younger than 6600 BP (7400 cal. BP) and the radiocarbon assays agree with this. The *Ulmus* decline (550cm) as inferred from data at Torran Beithe was dated as around 5240 BP (5970 cal. BP). The radiocarbon assay at LCXII (539 cm) provides an age of 5120 ± 55 (5910 cal. BP) indicating a slight but insignificant mismatch within the chronology. However, the depth for the *Ulmus* decline is approximate. The radiocarbon assays generated are assumed to be correct.

Laboratory reference	Depth (cm)	C content (% by wt)	Conventional Radiocarbon age (yr BP ± 1σ)	Calibrated age range to 1σ (yr. BP)	Calibrated age range to 2σ (yr. BP).
AA34232	484	3.7	4900 ± 55	5659-5593	5738-5491
AA34231	513	5.0	4850 ± 55	5612-5492	5660-5473
AA34230	539	8.0	5120 ± 55	5926-5754	5988-5735
AA34229	597	5.9	6205 ± 65	7234-7004	7266-6904
AA34221	682	5.0	9395 ± 75	10728-10505	11059-10404

Table 6.16 Core LCXII AMS radiocarbon (BP) and calibrated (cal. BP) ages.

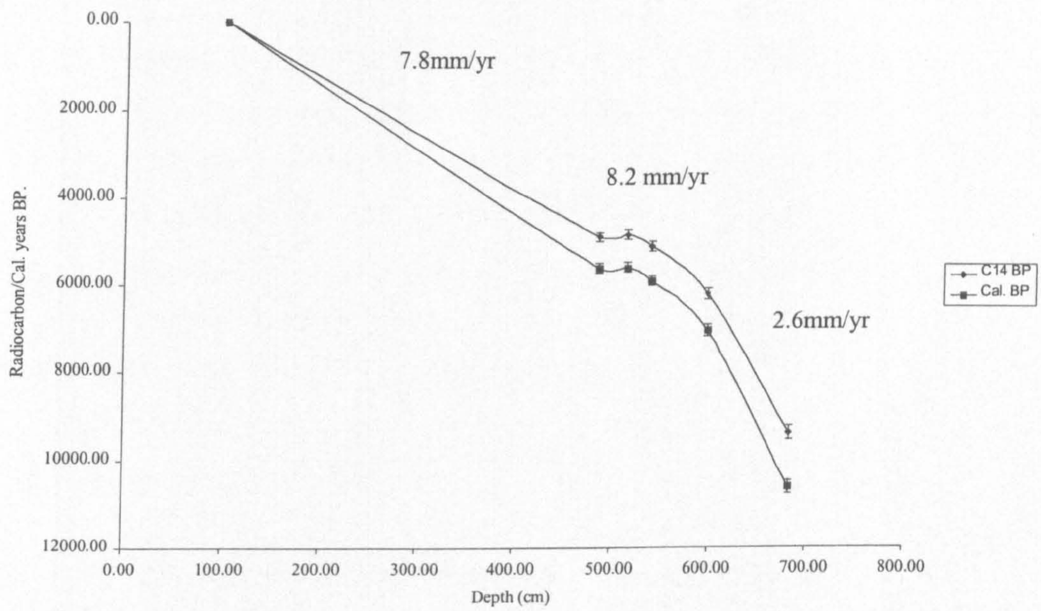


Figure 6.32 LCXII Age depth curve with radiocarbon BP and cal. BP ages plotted against depth (Table 6.15), with errors plotted to 2 σ. Sediment accumulation rates in mm per radiocarbon year are also shown.

The start of organic sedimentation is thought to have occurred immediately prior to 9395 ± 75 BP (10613 cal. BP). The deposition of coarse minerogenic sediment bands occurred as discrete events. The first event occurred at 6205 ± 65 BP (7060 cal. BP) (LCXII 4) with three further events (LCXII 5, 6 and 7) occurring at 5120 ± 55 BP (5910 cal. BP), 4850 ± 55 BP (5600 cal. BP) and 4900 ± 55 BP (5610 cal BP). A 2σ error of *around 150 radiocarbon years means that events LCXII (5, 6 and 7) cannot* be defined as chronologically significantly separate events, and thus are described as three stratigraphically separate events occurring between *c.* 5120-4850 BP (5910-5600 cal. BP).

The age-depth curve (Figure 6.32) suggests that within core LCXII there are marked changes in sediment accumulation rates. The sediment accumulation rate for basal sediments between 682 and 597 cm is 2.6mm/ year ^{14}C, which increases to 8.2 mm/ year ^{14}C between 597 and 484 cm. Interpolation of the age-depth curve suggests an accumulation rate of 7.8 mm/ year ^{14}C between 484 and 0 cm. The variable sediment accumulation rates probably reflect changes in sedimentation processes, discussed fully in Chapter 7.

The interpolation of the age-depth curve to the sediment surface may be inaccurate owing to the depth over which age is interpolated and likely variations in sedimentation rates associated with changes in sediment type. It was hoped that further chronological definition of sediments above 484 cm (4900 ± 55 BP (5610 cal. BP) would be achieved through pollen analysis data (Section 5.5); however, the poor sampling resolution (Section 6.5.4) indicates that there are probably no regionally synchronous palynological events defined.

6.5.6 SUMMARY OF STRATIGRAPHIC EVENTS

Sediment analysis has defined several significant sedimentary events within core LCXII; these are summarised in Figure 6.33, outlined below and further discussed in Chapter 7.

- Event LCXII 1: the deposition of basal laminated silts and sands between 707-688 cm.
- Event LCXII 2: marks the start of organic sedimentation at 682 cm as defined from sediment description (Section 6.5.1) and loss-on-ignition analysis (Section 6.5.2), with a rapid increase from 10% loss-on-ignition at 688 cm to 26% at 682 cm. The onset of organic sedimentation occurred at 9395 ± 75 BP (10613 cal. BP).

- Event LCXII 3: is a sustained increase in loss-on-ignition over 8 cm with <33% at 622 cm and >45% at 614 cm (Section 6.5.2, Figure 6.24 (c).
- Event LCXII 4: a unit of thin (<2 cm) well defined sand (599-597cm) identified through sediment description (Section 6.5.1), loss-on-ignition, percentage water content (Section 6.5.2) and grain size analysis (Section 6.5.3, Figure 26a). Grain size analysis suggests that this sediment unit shares a mode of deposition from higher energies, such as fluvial processes. Radiocarbon assay suggests that this unit was deposited at 6205 ± 65 BP (7060 cal. BP).
- Event LCXII 5 a unit of thin (<2 cm) well defined sand (541-539cm) identified through sediment description (Section 6.5.1), loss-on-ignition, percentage water content (Section 6.5.2) and grain size analysis (Section 6.5.3, Figure 6.26a). Coarse grain sizes within this unit suggest a higher energy mode of deposition, with depositional energy greater than is associated with 'normal' lacustrine processes. Radiocarbon assay suggests that this unit was deposited at 5120 ± 55 BP (5910 cal. BP).
- Event LCXII 6: a unit of thin (<2 cm) well defined sand (515-513 cm) identified through sediment description (Section 6.5.1), loss-on-ignition, percentage water content (Section 6.5.2) and grain size analysis (Section 6.5.3, Figure 6.26a). Grain size distribution curve analysis suggests a higher energy of deposition than 'normal' lacustrine processes for this unit, such as fluvial processes. Radiocarbon assay suggests that this unit was deposited at 4850 ± 55 BP (5600 cal. BP)
- Event LCXII 7: is defined from sediment description as a highly minerogenic graded sedimentary unit (Section 6.5.1), a significant reduction in loss-on-ignition and as a reduction in percentage water content between 493-485 cm (Section 6.5.2). Grain size analysis (Section 6.5.3, Figure 6.27) indicates an initial coarsening up sequence culminating in the deposition of a medium, sand followed by a fining up sequence. Grading within this unit suggests an initial increasing energy of deposition followed by a decline in depositional energy. Radiocarbon assay suggests that this unit was deposited at 4900 ± 55 BP (5600 cal. BP).
- Event LCXII 8: is identified as a significant reduction in loss-on-ignition analysis and a reduction in percentage water content between 445-435 cm (Section 6.5.2). Grain size analysis (Section 6.5.3, Figure 6.28) suggests that this unit is graded, with an initial coarsening up sequence followed by fining up, indicating a similar mode

of deposition to event LCXII (7), but grain size is finer and particle size distributions do not suggest such well defined grading for this unit.

- Events LCXII 9: (391-389cm) and 10 (273-271cm) are thin (<2 cm) well defined sand units identified through sediment description (Section 6.5.1), a significant reduction in loss-on-ignition, a reduction in percentage water content (Section 6.5.2) and grain size analysis (Section 6.5.3). Grain size analysis suggests that all these units share a similar mode of deposition with a higher energy than that of lacustrine processes, such as fluvial processes.
- Event LCXII 11: (199-195cm) is defined as a reduction in percentage water content and as a significant reduction in loss-on-ignition analysis (Section 6.5.2). Grain size data (Section 6.5.3, Figure 6.28) suggest that this unit shares a similar mode of deposition with events LCXII 4, 5, 6, 8 and 9.
- Event LCXII 12: a sustained increase in loss-on-ignition (Section 6.5.2) from 194 cm suggests increasingly organic sediment with detrital peat accumulating from 150 cm (Section 6.5.1).

Changes in aquatic vegetation and associated water depths as determined from pollen analysis proved inconclusive (Section 6.5.4.1). Pollen corrosion data suggest that pollen deposited between 650-600 cm may have been subjected to increased amounts of aerobic decay (Section 6.5.4.1).

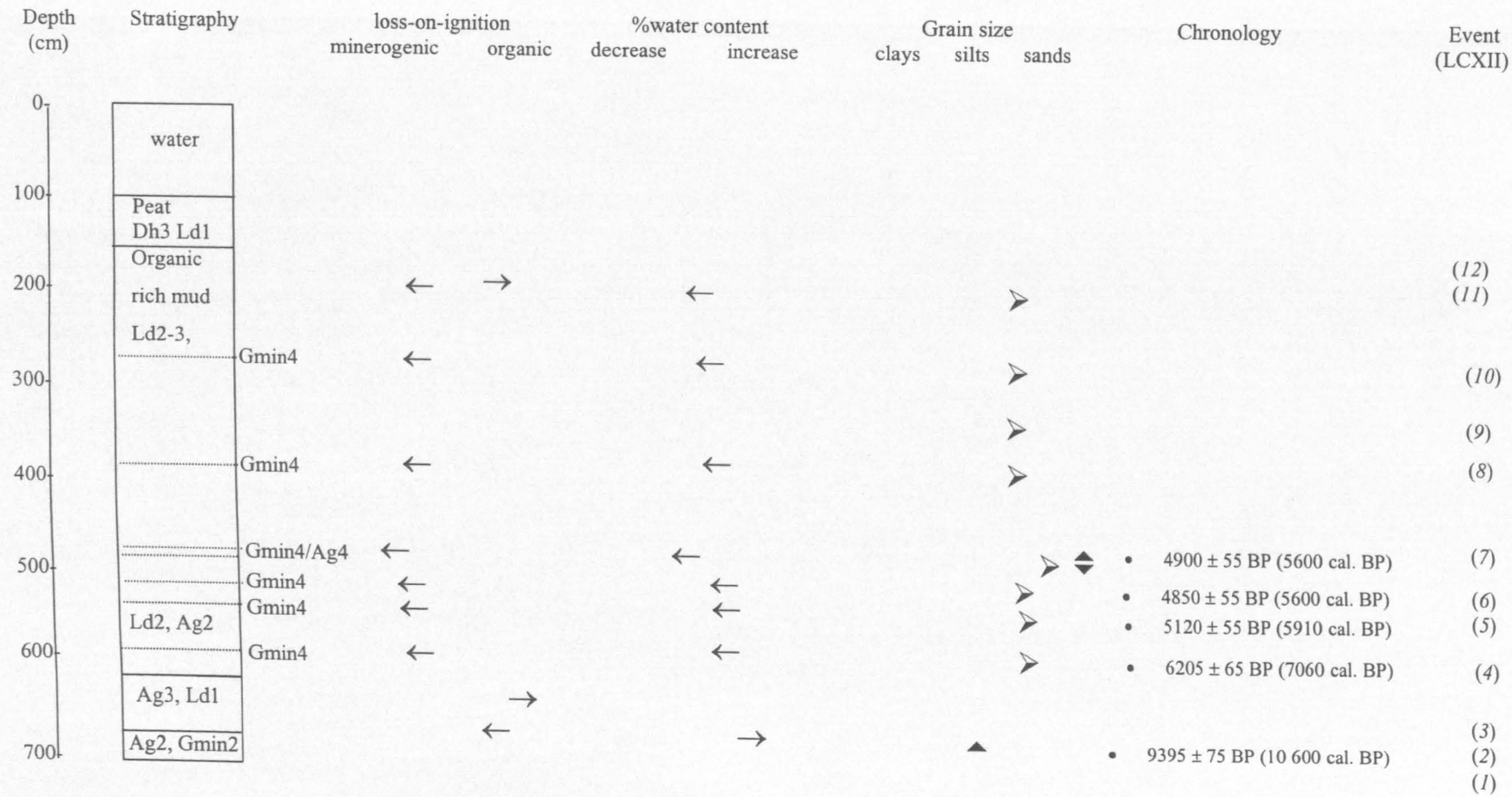


Figure 6.33 Core LCXII sediment analysis summary of stratigraphic events. ← & → denote decreasing and increasing values, ▲ denotes fining up sequence, ➤ denotes increased grain size and ⬍ denotes a coarsening up and fining up sequence.

6.6 CORE LCVII (Figure 5.1)

6.6.1 SEDIMENT STRATIGRAPHY

Depth probing showed the bedrock contact to lie at 751 cm (Figure 5.2), but Russian core samples extend to only 703 cm. Basal sediments (703-687 cm) are laminated, consisting of 1-2 mm thick laminae of better sorted light grey silts and darker grey sands. There is a sharp (<1 mm) lithological boundary with organic-rich greenish brown organic silts between 687-659 cm. Above this, dark brown organic silts with varying organic detrital (Dh) content accumulate, which are interrupted by the deposition a series of well-defined, pale brown sand units around 2 cm thick at 606-604 cm, 564-562 cm, 548-546 cm and 479-477 cm. A major minerogenic sedimentary unit between 534-524 cm comprises a coarsening up sequence of clays passing up into coarse sands. Sediment becomes increasingly detritus-rich and organic with the final accumulation of detrital peat at 208-155 cm.

Depth cm	LCVII Sediment Description
0-97	Water
97-155	Sediment-water interface not sampled.
155-208	Dark brown fibrous detrital peat, herbaceous and woody plant fragments; Dh3, Ld1 Dl+, Nig3, Strf0, Elas2-3, Sicc2, Lim1-2.
208-334	Dark brown detritus rich silt, mica flakes visible with large >2mm herbaceous plant macrofossils; Dh2 Ld1-2, Ag1 Nig3, Strf0, Elas2, Sicc2, Lim1-2.
334-430	Dark brown organic-rich silt, mica flakes visible, woody plant fragments; Ld2-3, Ag1-2, Dh+, Nig3, Strf1, Elas1, Sicc2, Lim1-2.
430-477	Dark brown organic-rich silt; Ld2, Ag2 Dh+, Nig3, Strf1, Elas1, Sicc2, Lim3.
477-479	Pale brown sand band; Gmin4, Nig1-2, Strf2, Elas0, Sicc2, Lim3.
479-524	Dark brown organic-rich silt; Ld2, Ag2 Dh+. Nig3, Strf1, Elas1, Sicc2, Lim3
524-527	Dark grey silt; Ag4, Nig3, Strf1, Elas1, Sicc2., Lim2.
527-533	Pale brown coarse sand with plant fragments; Gmin3, Gmaj1 Dh+, Nig1, Strf2, Elas0, Sicc2-3, Lim3.
533-534	Grey clay; As4, Nig2, Strf2, Elas1, Sicc2, Lim3.
534-546	Dark brown organic-rich silt; Ld2, Ag2, Dh+, Nig3, Strf0, Elas0, Sicc2, Lim3.
546-548	Pale brown sand; Gmin4, Nig1-2, Strf2, Elas0, Sicc2, Lim3.
548-566	Dark brown organic-rich silt; Ld2, Ag2, Dh+, Nig3, Strf0, Elas0, Sicc2, Lim3.
566-568	Pale brown sand; Gmin4, Nig1-2, Strat2, Elas0, Sicc2, Lim3.
568-604	Dark brown organic-rich silt; Ld2, Ag2, Dh+, Nig3, Strf0, Elas0, Sicc2
604-606	Pale brown sand; Gmin4, Nig1-2, Strf2, Elas0, Sicc2-3.
606-659	Dark brown organic-rich silt; Ld2, Ag2, Dh+, Nig3, Strf0, Elas0, Sicc2, Lim3.
659-687	Greenish brown organic-rich silt; Ag3, Ld1 Dh+, Nig2-3, Strf1, Elas0, Sicc2, Lim3.
687-703	Silver grey coarse silts and sands, mica flakes visible, Ag2, Gmin2, Nig1, Strf4, Elas0, Sicc2.

Table 6.17 Core LCVII sediment description using a modified Troels-Smith notation (Section 5.2.1).

The basal minerogenic sediments (703-687 cm) were noted to be thixotropic and could not be sampled satisfactorily.

6.6.2 BULK DENSITY, WATER CONTENT AND LOSS-ON-IGNITION.

The bulk density record for core LCVII (Figure 6.34a) is inconclusive owing to the incomplete data set (Section 6.3.2), and is not discussed.

Percentage water content (Figure 6.34b) appears to be sensitive to changes in sediment minerogenic content. After an abrupt increase at 685 cm percentage water content value is remarkably constant. However, several abrupt reductions occur at 219-217cm, 237-235 cm, 479-477 cm, 533-525 cm, 548-546 cm, 568-566 cm, 600-598 cm and 606-604 cm. Reductions at 606-604 cm, 568-566 cm, 548-546 cm, 533-525 cm and 479-477 cm correspond to deposition of minerogenic sediment units described in Table 6.18. Other fluctuations correspond to abrupt reductions in percentage loss-on-ignition (see discussion below). Between 421-319 cm there is a gradual reduction in percentage water content with values showing increased variability, which corresponds to a period of decreased but variable organic content (Figure 6.34c). Although many of the above units are single-sample events the close correlation with changes in the percentage loss-on-ignition (Table 6.18) would suggest that percentage water content is a sensitive proxy for sediment minerogenic content. This is further discussed in Section 7.7.

Variations in the percentage loss-on-ignition (Figure 6.34c) indicate an initial abrupt rise in organic content with an increase from 5% to 23% over 4 cm at 687-683 cm. Between 683-633 cm, basal sediments have consistent values of between 16-24%. The end of this unit is marked by a sustained rise in loss-on-ignition values, with <25% at 633 cm to >37% at 625 cm Between 625-214 cm, the loss-on-ignition values and thus the organic content for these sediments is highly variable. Significant fluctuations from the mean percentage loss-on-ignition for lacustrine sediment between 625-214 cm (n=205, 37.4%) are defined by their exceeding the 1σ Standard Deviation, ±6.2%, (Figure 6.34c) in Table 6.18.

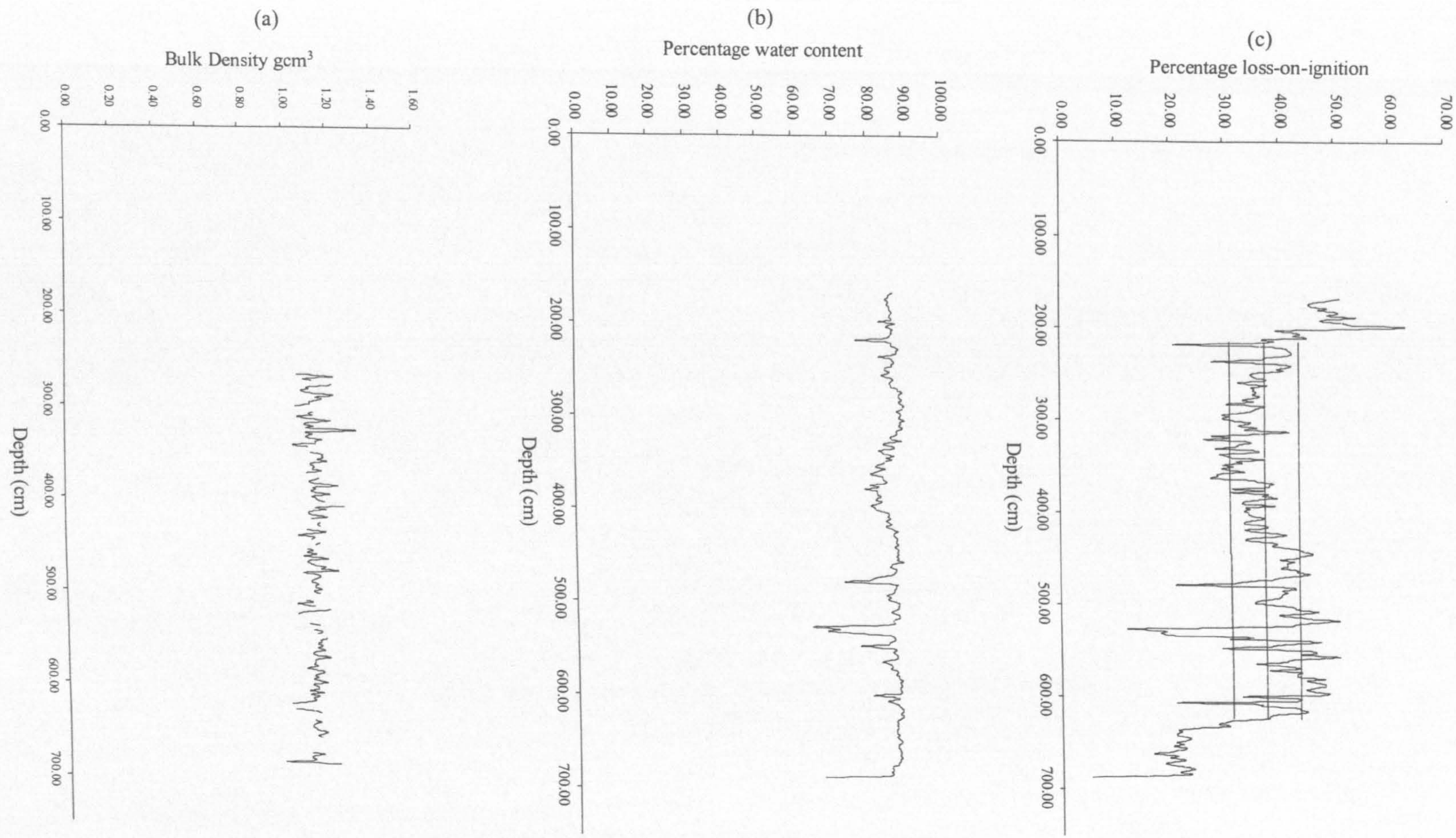


Figure 6.34 (a, b & c) Core LCVII bulk density (a) percentage water content (b) and percentage loss-on-ignition (c) (see text) plotted against depth.

Depth (cm)	Number of analyses	Sediment description
625-612	7	Period of increasing organic content with a significant ↑ in loss-on-ignition between 616-614 cm
612-604	4	Decreasing organic content with a significant ↓ in loss-on-ignition between 606-604 cm. Stratigraphically this is a well defined sand unit and a minor reduction in percentage water content.
604-578	13	Increasing organic content with a minor reduction at 599cm, a prolonged period of a significant ↑ in loss-on-ignition between 598-578 cm.
578-563	7	A minor decrease in organic content.
563-548	8	Increasing organic content with a significant ↑ in loss-on-ignition between 558-548 cm.
548-546	1	Abrupt significant ↓ in loss-on-ignition. Stratigraphically this is a well defined sand unit and a notable reduction in percentage water content.
546-538	7	Increasing organic content with a significant ↑ in loss-on-ignition between 542-538 cm.
538-523	8	A period of decreasing organic content with a significant ↓ in loss-on-ignition between 533-523 cm. Stratigraphically this is a well defined sand unit and a very notable reduction in percentage water content
523-505	9	After an initial rapid increase on organic content there is a significant ↑ in loss-on-ignition between 519-505cm with a minor reduction between 510-512 cm.
505-477	14	A gradual decline in organic content with a significant ↓ between 479-477 cm. Stratigraphically this is a well defined sand unit and a notable reduction in percentage water content
477-441	18	After an initial rapid increase on organic content organic content is higher with a significant ↑ in loss-on-ignition between 469-465cm and 447-441cm
441-217	112	A prolonged period of reduced organic content, with variable loss-on-ignition values. Several significant ↓ in loss-on-ignition occur in particular between 361-315 cm with a series of 7 often single sample significant reductions, occurring during a period of reduced water content. Other significant ↓ occur at 299-297 cm, 295-293 cm, and 283-281cm. Significant ↓ at 237-234 cm and 219-217 cm are associated with a reduction in water content.
217-214	2	Abrupt increase in loss-on-ignition and organic content.

Table 6.18 Core LCVII significant fluctuations in loss-on-ignition (Figure 6.34c), with notes on corresponding changes in percentage water content (Figure 6.34b) and sediment stratigraphy (Table 6.17) included.

Above 214 cm there is a continued rapid rise in organic content with a loss-on-ignition value of 63% at 198 cm. Within the sediment description a detrital peat is recorded between 155 and 208 cm, but loss-on-ignition data show that this peat is still minerogenic.

Many of the significant reductions in loss-on-ignition (Table 6.18) are single-sample events. However, the correspondence with either reductions in percentage water content or changes in sediment stratigraphy would suggest that these sediments do represent actual increases in minerogenic content. Significant reductions in loss-on-ignition recorded between 612-441 cm occur as abrupt reductions in loss-on-ignition, appearing to interrupt the deposition of highly organic sediment, with the rapid resumption of organic sedimentation after deposition of these more minerogenic sediments (see Section 7.7.1). Significant reductions in loss-on-ignition recorded between 441-219 cm occur during a period of sedimentation with a reduced organic content but these reductions also appear to be abrupt fluctuations in minerogenic content (see Section 7.7).

The well defined sand unit at 568-566 cm (Table 6.17) is recorded as a non-significant reduction in loss-on-ignition and is associated with a minor reduction in percentage water content. Owing to the well defined stratigraphic evidence this unit is considered as a significant increase in minerogenic content.

6.6.3 GRAIN SIZE ANALYSIS

Grain size analysis was carried out for sediments at contiguous 2.0 cm. Missing data are due to unreliable grain size measurements (Section 5.3).

Figure 6.35 summarises the grain size data. The diagram has been visually divided into zones defining periods of changing sediment grain size composition. These zones, LCVII (a-m), are described in Table 6.19 with coarse sediment depositional events labelled as LCVII (4-12). Several of the zones are described as prolonged coarsening or fining-up sequences. This suggests than during sedimentation within the basin at LCVII there have been sustained changes in the energy of deposition. The coarse grain size events (LCVII 4-12) represent abrupt changes in the energy of deposition and occur during these more prolonged periods of changing depositional energy. Event LCVII 7 consists of an initial coarsening-up sequence culminating in the deposition of medium to fine sand followed by a fining-up sequence. This suggests that the depositional event consisted of an initial increase in depositional energy followed by a decrease. Event LCVII 10 shows a similar coarsening-up sequence with a deposition of a medium to fine sand; however, the fining-up sequence is less well defined.

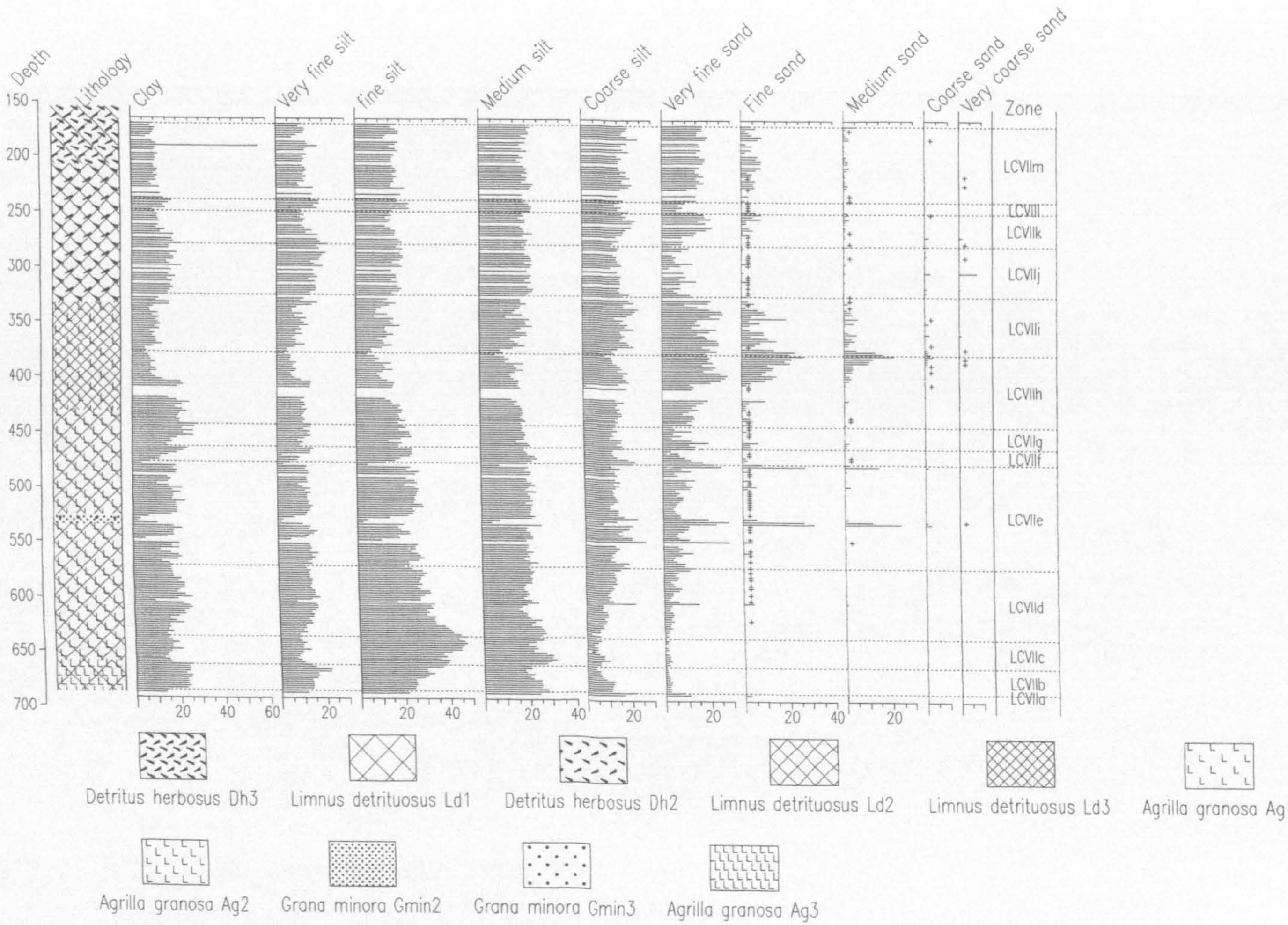


Figure 6.35 Core LCVII grain size plotted against depth.

Zone	Depth (cm)	Grain size description	Stratigraphic event
LCVIIa	687-685	Silty very fine sand, with 16-20% coarse silt and 7-10% very fine sand.	
LCVIIb	685-661	Fine silts. Gradual fining up sequence with ↑ clays, very fine silts and fine silt, with ↓ medium and coarse silts and <2% very fine sand.	
LCVIIc	661-635	Medium silts. Interruption to fining up sequence with ↑ medium and coarse silt content associated with ↓ clay and very fine silt. Although fine silt content continues to increase (46%)	
LCVIId	635-571	Fine to medium silts. Start of a gradual coarsening-up sequence with ↑ coarse silts (3-18%) and very fine sands (1-12%), also minor increases in clays and very fine silts, ↓ fine silt (46-20%). Abrupt deposition event of a coarser silty sand (22% coarse silt, 15% very fine sand 4% fine sand) at 606-604 cm.	LCVII(4) at 606-604 cm. LCVII (5) 568-566cm*
LCVIIe	571-476	Silty sands (15-22% medium silts, 12-15% coarse silts, 5-10% very fine sand). No overall grading within the unit. Abrupt deposition event of a coarser silty sand (27% coarse silt, 28% very fine sand, 4% fine sand) at 548-546 cm Well graded unit noted at 533-525 cm, an initial coarsening-up sequence with deposition of medium (24%) to fine (31%) sand at 531-529 cm followed by a fining up sequence.	LCVII (6) 548-546 cm. LCVII (7) 533-525 cm.
LCVIIf	476-462	Abrupt deposition of a fine (27%) to medium (14%) sand at 479-477cm, followed by a fining-up sequence with ↑ clay (4-24%), and fine silt (5-24%) and ↓ very fine sand (25-3%)..	LCVII (8) 479-477 cm.
LCVIIg	462-442	Abrupt deposition of a silty (16% coarse silt) very fine (16%) sand at 463-455 cm, followed by ↓ very fine sands.	LCVII (9) 463-455 cm
LCVIIh	442-377	A prolonged and gradual coarsening-up sequence with ↑ very fine (4-19%), fine (0-26%), medium (21%) and coarse sands (3%), with ↓ clays (29-5%) and fine silts (21-4%), with ↓ medium and coarse silts as sediment becomes notably coarser from 385 cm. Deposition of medium to fine sand at 379-377 cm	LCVII (10) 385-371 cm
LCVIIi	377-324	A prolonged and gradual fining-up sequence with ↓ fine (20-1%) and medium (16-<1%) sands., with 1% fine sand by 371 cm. Abrupt deposition of coarser very fine (26%) to fine (10%) sands at 341-335 cm.	LCVII (11) 341-335 cm
LCVIIj	324-278	Silty very fine sands, with 18-22% medium, 11-20% coarse silt, 3-12% very fine sand and <3% fine sand. Erratic sedimentation with units containing very coarse sands also deposited.	
LCVIIk	278-249	Coarsening-up sequence with ↑ coarse (11-22%) silts, very fine (3-21%) and fine (<1-8%) sands and ↓ clay (22-10%), very fine (22-10%) and fine (19-12%) silts. A very fine to fine sand deposited at 255-249 cm.	LCVII (12) 255-249 cm
LCVIII	249-239	Fining-up sequence with ↑ clay (9-13%), very fine (10-15%) and fine (16-23%) silts, with ↓ coarse (20-13%) silt, very fine (15-6%) and fine (5-,1%) sand.	
LCVIIm	239-170	Very fine (15-20%) to fine (5-10%) sand. Above 239 cm sediment remains coarser with ↑ amounts of coarse silts, very fine and fine sand.	

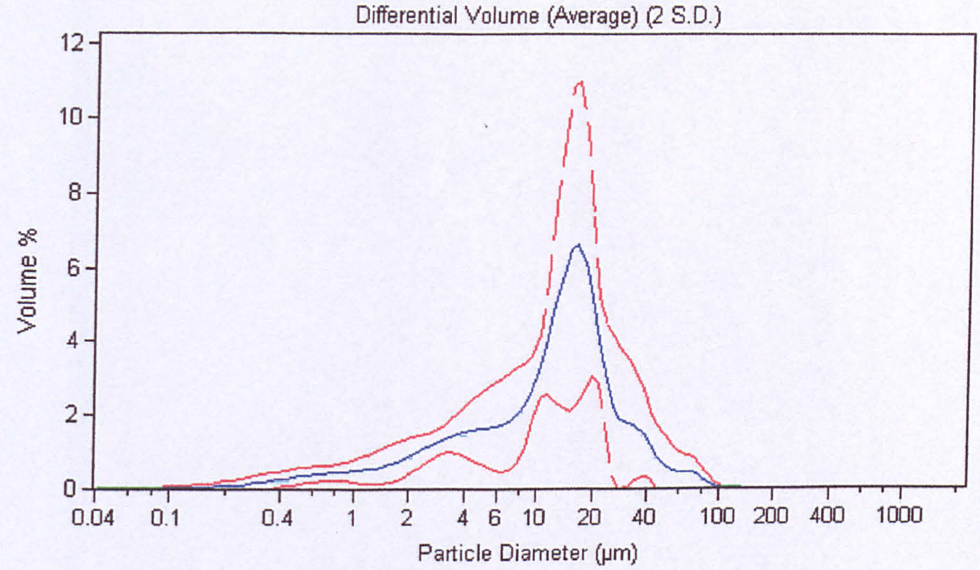
Table 6.19 Core LCVII summary of grain size data as presented in Figure 6.35, with stratigraphic events LCVII (4-12) described. * event LCVII (5) further discussed in the text.

Event LCVII 5 is recorded as a non-significant reduction in loss-on-ignition and percentage water content and as a stratigraphically well defined sand unit (Table 6.17). It is not recorded within the grain size analysis as an abrupt increase in sediment grain size. However this unit occurs during a period of increasing grain size and although potentially less significant than other events it is still considered as a depositional event.

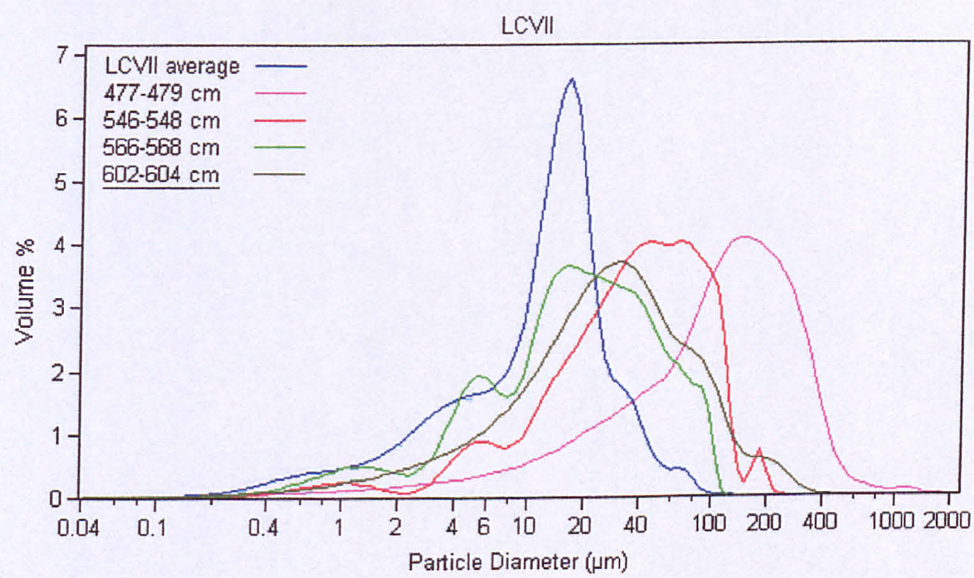
Events LCVII 4, 6, 7 and 8, are recorded as significant reductions in loss-on-ignition (Section 6.6.2; Table 6.18) with increasingly minerogenic sediments recorded as coarse grain size units. However, further significant reductions in loss-on-ignition as outlined in Table 6.18 are not recorded as increasingly coarse minerogenic units. Coarse grain size events LCVII (9-12) are not recorded as significant reductions in loss-on-ignition. The relationship between increased grain size and the loss-on-ignition record as a proxy for variation in minerogenic content is discussed in Section 7.7.2.

Further characterisation of the mode of deposition of sediment events (LCVII 4-12) was determined through the comparison of particle size distributions. Basal fine-grained organic-rich (23-33% loss-on-ignition) silts between 684-633 cm were considered representative of deposition where the primary sedimentation control were lacustrine sedimentary processes. Particle size distributions from these sediments were averaged within the COULTER LS230 software to produce a mean (n=26) particle size distribution curve with 2σ Standard Deviation Figure 6.36a, labelled as LCVII average.

Particle size distributions for sediment events LCVII 4-12 compared with the average distribution are shown in Figures 6.36b to 6.41. All the particle size distributions lie outside the 2σ average curve, suggesting a mode of deposition different to predominantly lacustrine sedimentation processes, such as fluvial or slope inwashing.



(a)



(b)

Figure 6.36 (a) and (b) Core LCVII percentage volume particle size distributions. (a) Average particle size distribution for minerogenic sediments typifying lake sedimentation within core LCVII (for definition see text) to a 2σ Standard Deviation. (b) Overlays of particle size distributions of Events LCVII 4, 5, 6 and 8 (Table 6.19) with the particle size distribution of typical minerogenic lake sediments (note the 2σ Standard Deviation is not shown).

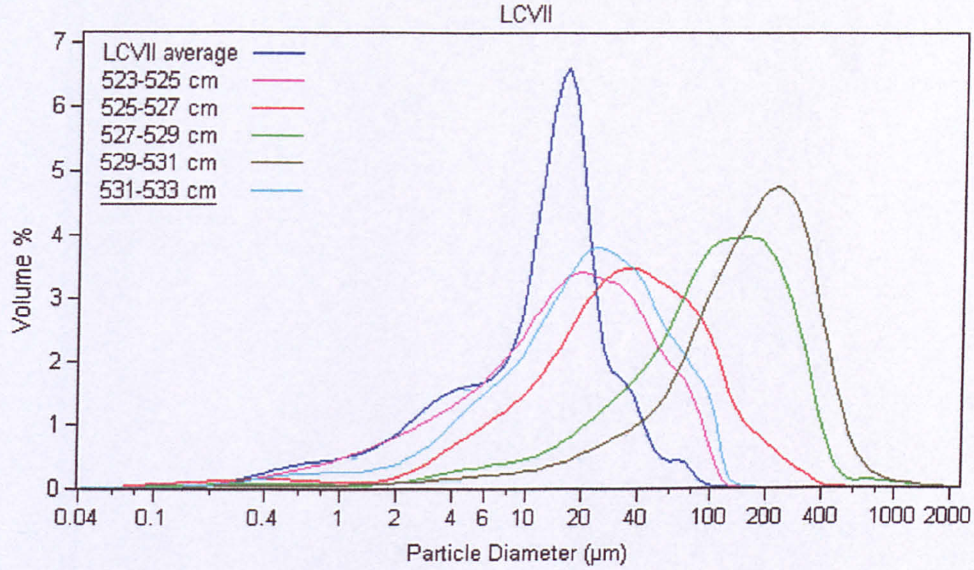


Figure 6.37 Core LCVII percentage volume particle size distributions. The graph overlays particle size distributions of the sequence of minerogenic sediments, Event LCVII 7 with the particle size distribution of typical minerogenic lake sediments (note the 2σ Standard Deviation is not shown, see Figure 6.36(a)).

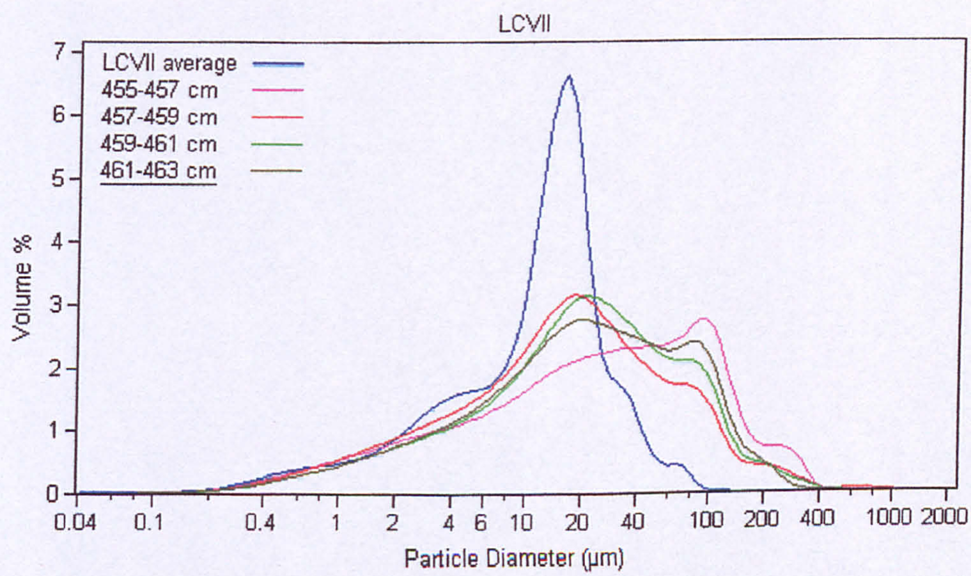


Figure 6.38 Core LCVII percentage volume particle size distributions. The graph overlays particle size distributions of the sequence of minerogenic sediments, Event LCVII 9 with the particle size distribution of typical minerogenic lake sediments (note the 2σ Standard Deviation is not shown, see Figure 6.36(a)).

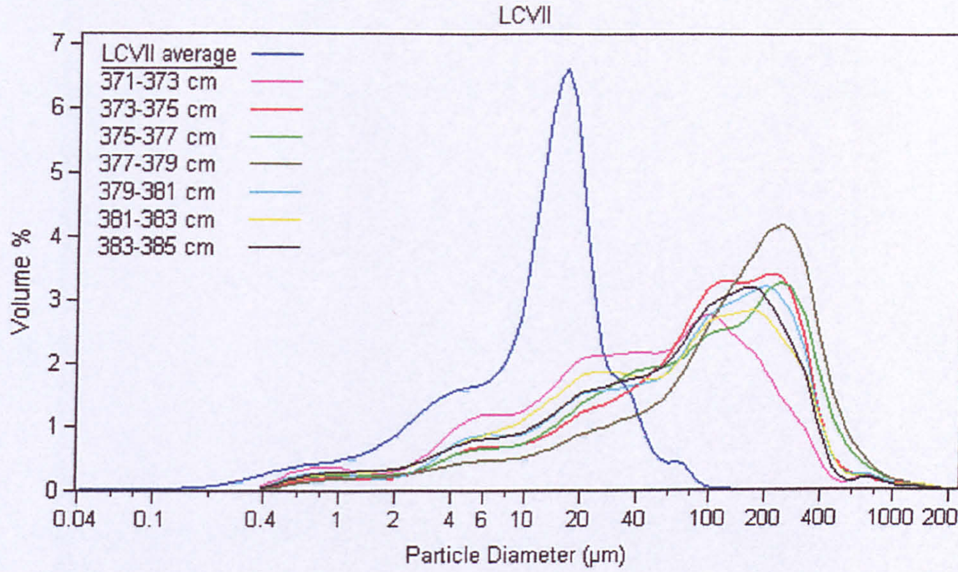


Figure 6.39 Core LCVII percentage volume particle size distributions. The graph overlays particle size distributions of minerogenic unit, Event LCVII 10 particle size distribution of typical minerogenic lake sediments (note the 2σ Standard Deviation is not shown, see Figure 6.36(a)). Note no PIDS particle size data was generated for sediment sample 343-345 cm (section 5.3).

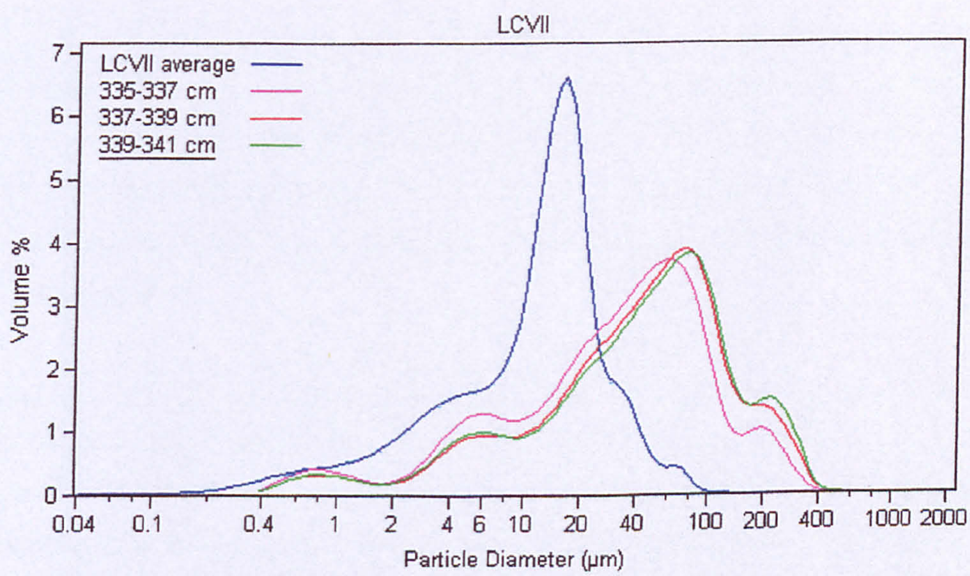


Figure 6.40 Core LCVII percentage volume particle size distributions. The graph overlays particle size distributions of the sequence of minerogenic sediments, Event LCVII 11 with the particle size distribution of typical minerogenic lake sediments (note the 2σ Standard Deviation is not shown, see Figure 6.36(a)).

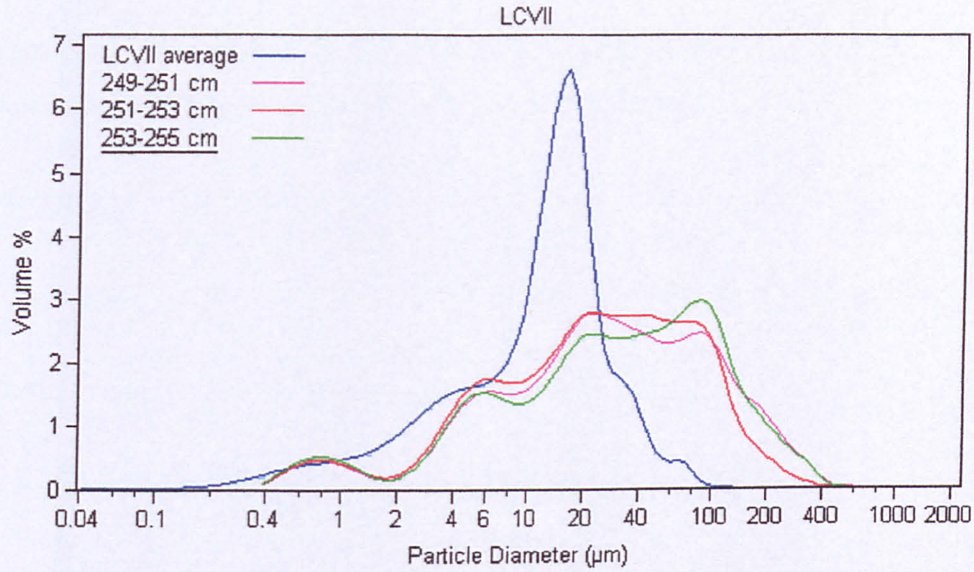


Figure 6.41 Core LCVII percentage volume particle size distributions. The graph overlays particle size distributions of the sequence of minerogenic sediments, Event LCVII 12 with the particle size distribution of typical minerogenic lake sediments (note the 2σ Standard Deviation is not shown, see Figure 6.36(a)).

Events LCVII 4, 5 (Figure 6.36b), 9, (Figure 6.38) and 12 (Figure 6.41) are finer grained than the other events and represent a relative lower energy of deposition. These lower energy events also tend to be more poorly sorted, in particular events LCVII 9 and 12, possibly reflecting sediment mixing with finer grained lacustrine sediments during deposition.

Events LCVII 6 (Figure 6.36b), 7 (Figure 6.37), 8 (Figure 6.36b), 10 (Figure 6.39), and 11 (Figure 6.40) are coarser than other events and represent a relatively higher energy of deposition. These higher energy events are also well sorted possibly a result of (a) a greater degree sorting of these easily transported fine-medium sands or (b) sorting during sediment transport for example by fluvial processes prior to deposition. Grain size grading recorded with events LCVII 7 and 10 (Table 6.19) is also shown in the particle size distributions (Figures 6.37, 6.38), suggesting an initial increase in depositional energy followed by a waning.

6.6.4 CHRONOLOGY

Within core LCVII an AMS radiocarbon date was obtained on the start of organic sedimentation (Table 6.17 and Section 6.5.2, Figure 6.34c).

Laboratory reference	Depth (cm)	C content (% by wt)	Conventional Radiocarbon age (yrs BP ± 1σ)	Calibrated age range to 1σ (yrs. BP)	Calibrated age range to 2σ (yrs. BP).
AA34222	681	7.1	9010 ± 65	10222-10156	10242-9918

Table 6.20 Core LCVII AMS radiocarbon (BP) and calibrated (cal. BP) ages.

6.6.5 SUMMARY OF STRATIGRAPHIC EVENTS

Sediment analysis has defined several significant sedimentary events within core LCVII; these are summarised in Figure 6.42, outlined below and further discussed in Chapter 7.

- Event LCVII 1: the onset of minerogenic sedimentation with the deposition of laminated silts and sands, recorded between 703-687 cm but probably extending to a depth of 751 cm (see Section 7.2).
- Event LCVII 2: an abrupt rise (over 4 cm) in organic content with loss-on-ignition values increasing from <5% at 687 cm to >23% at 683 cm (Section 6.6.2). Radiocarbon assay suggests that onset of organic sedimentation occurred at 9010 ± BP (10 200 cal. BP) (see Section 7.3).
- Event LCVII 3: a sustained rise in organic content with loss-on-ignition values increasing from 25% at 633cm to >33% at 625cm (Section 6.6.2) (see Section 7.5).
- Event LCVII 4: 606-604 cm a thin (2cm) relatively well sorted silty very fine sand defined from sediment description (Section 6.6.1) and loss-on-ignition (Section 6.6.2). Coarser grain sizes as compared to sediments typifying lacustrine sedimentation would suggest a higher energy mode of deposition for this unit, such as fluvial processes (see Section 7.7).
- Event LCVII 5: 566-568 cm, a thin (2cm) relatively well sorted silty very fine sand defined from sediment description (Section 6.6.1) and loss-on-ignition (Section 6.5.2). Grain size analysis (Section 6.6.3) suggests that this unit has higher

depositional energy than lacustrine processes, such as fluvial processes (see Section 7.7).

- Event LCVII 6: 548-546 cm a thin (2 cm) relatively well sorted silty very fine sand as defined from sediment description (Section 6.6.1) and loss-on-ignition (Section 6.6.2). Grain size analysis suggests that this unit has a coarser grain size indicating a high energy of deposition than predominantly lacustrine processes and those sediments recorded in Events 4 and 5 (see Section 7.7).
- Event LCVII 7: 533-525 cm is well defined from sediment description (Table 6.17) and loss on-ignition (Section 6.6.2) as a coarse minerogenic unit. Grain size analysis (Section 6.6.3; Figure 6.37) suggests that the event consists of a coarsening-up sequence with well sorted coarse silts passing up into very well sorted fine to medium sands, with a subsequent fining-up sequence and the deposition of well sorted coarse silts. The grain size grading indicates an initial increase in the energy of deposition followed by a waning. The well sorted nature of all the sediments suggests a period of sediment transport prior to deposition, such as within a fluvial system (see Section 7.7).
- Event LCVII 8: 479-477 cm is defined from sediment description (Table 6.17) and loss-on-ignition (Section 6.6.2) as a thin (2 cm) minerogenic sand unit. Grain size analysis (Section 6.6.3; Figure 6.36b) indicates that the event is a well sorted fine sand, with sorting reflecting grain size, that is an easily transported fine sand, and a period of sediment transport prior to deposition such as within a fluvial system (see Section 7.7).
- Event LCVII 9: 463-455 cm is defined only through grain size analysis (Section 6.6.3; Figure 6.37) as a poorly sorted silty very fine sand with a higher depositional energy than lacustrine sedimentary processes such as fluvial processes (see Section 7.7.2).
- Event LCVII 10: 385-371 cm is defined only through grain size analysis (Section 6.6.3; Figure 6.40). Sediments within this event show a gradual coarsening-up sequence within the predominantly poorly sorted fine sands with the deposition of a well sorted medium to fine sand at 379-377 cm. Any subsequent fining-up sequence is difficult to define from these data but is described in Table 6.19. Similar to Event LCVII 7, the grain size grading in LCVII 10 suggests changing energy of deposition within the unit. However, the smaller grain size and less well

sorted nature of the sediments indicates lower energy of deposition and a shorter sediment transport time or sediment mixing on deposition (see Section 7.7.2).

- Event LCVII 11: 341-335 cm is defined only through grain size analysis (Section 6.6.3; Figure 6.40) and consists of relatively well sorted very fine to fine sands and shows a similar particle size distribution to Event LCVII 6, suggesting a common energy of deposition (see Section 7.7.2).
- Event LCVII 12, 255-249 cm is defined, only through grain size analysis (Section 6.6.3; Figure 6.41) as a poorly sorted very fine to fine sand. This event and event LCVII 9 show similar particle size distribution suggesting a similar mode of deposition. The poor sorting and finer grain size of both events could suggest sediment mixing with finer grain lacustrine sediments during these lower energy depositional events (see Section 7.7.2).
- Event LCVII 13 is the start of detrital peat accumulation, with a sustained rise in loss-on-ignition (Section 6.6.2) from <40% at 214 cm to >60% at 198 cm. Sediment description records the accumulation of fibrous detrital peat from 208 cm (Table 6.17). Loss-on-ignition suggests that this peat is still mineral rich (Figure 6.34c) (see Section 7.7.3).

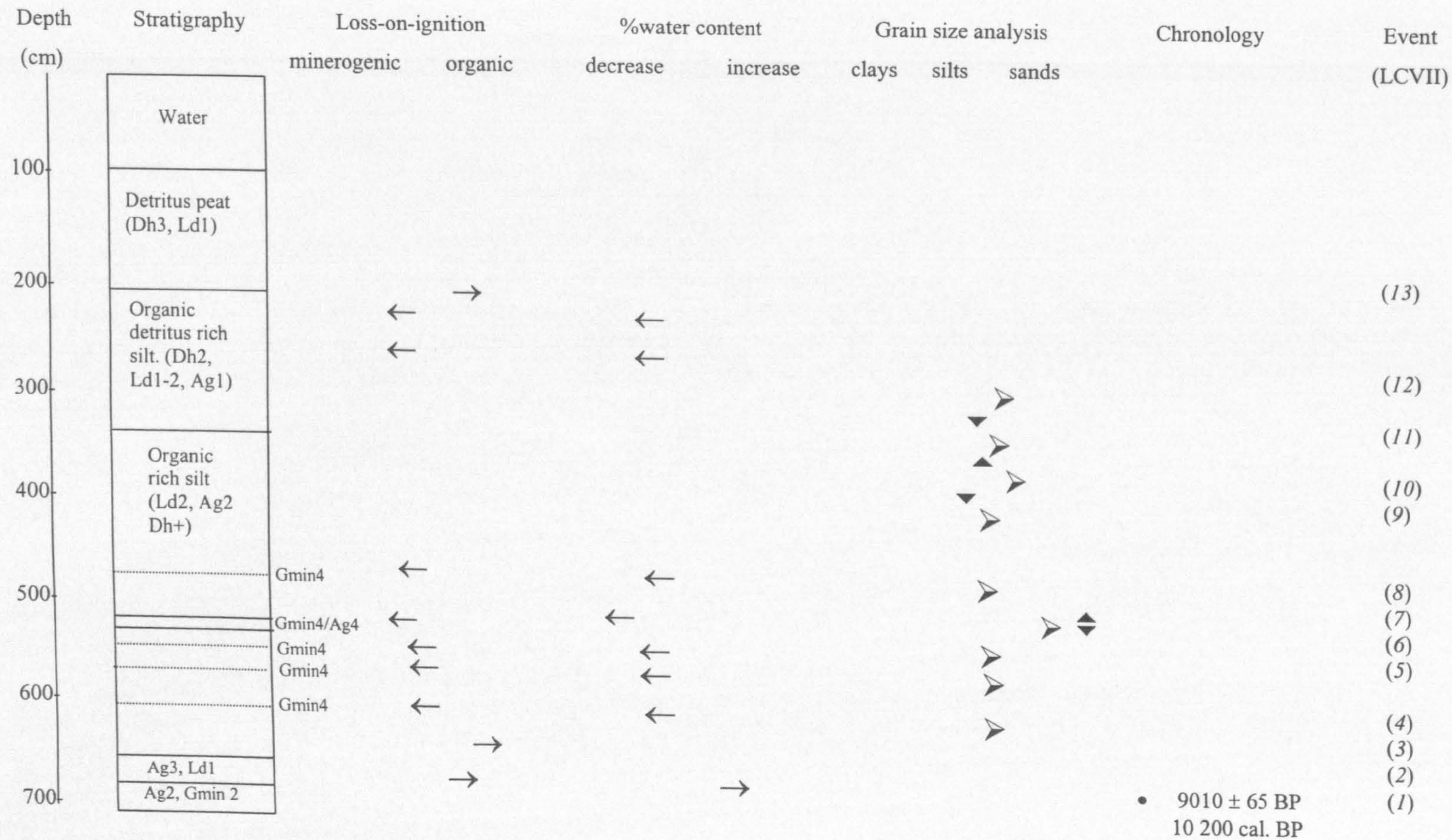


Figure 6.41 Core LCVII sediment analysis summary of stratigraphic events. ← & → denote decreasing and increasing values, ▲ denotes fining up sequence, ▼ denotes a coarsening up sequence, ➤ denotes increased grain size and ⬍ denotes a coarsening up and fining up sequence.