A MULTI-PROXY STUDY OF LATE HOLOCENE ENVIRONMENTAL CHANGE IN THE PROKLETIJE MOUNTAINS, MONTENEGRO AND ALBANIA

A thesis submitted to The University of Manchester for the degree of Doctor of Philosophy in the Faculty of Humanities

2011

Rose Wilkinson

School of Environment & Development

CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

The University of Manchester, Rose Wilkinson, Ph.D., May 2011.
A multi-proxy study of late Holocene environmental change in the Prokletije Mountains, Montenegro and Albania.

Palaeoenvironmental investigations from the Lake Plav catchment of the Prokletije Mountains in Montenegro and Albania, allowed primarily climatic change and anthropogenic influences during the Late Holocene and particularly the Little Ice Age (LIA) to be identified. Three sediment cores were analysed, two from Lake Plav (904 m a.s.l., cores LPCA and LPCB) and one from the upper catchment site of Lake C in Buni i Jezerces (1754 m a.s.l., core BJC1). These sediments were analysed for a variety of proxies including pollen, ostracoda, organic content, magnetic susceptibility and particle size. Chronologies for each sediment core were constructed using AMS radiocarbon, ^{210}Pb and ^{137}Cs dating techniques. The lower sites provided a record of past flood events, anthropogenic influences, lake development and infilling that have occurred since ***c.*** AD 500. Core BJC1 provided longer-term data since ***c.*** 2720 BC, providing complementary records of *Pediastrum* and thermophilous arboreal types, identified following a catchment vegetation survey. Glacial geomorphological mapping of the Maja e Koljaet glacier in Buni i Jezerces, Albania, enabled a catchment specific palaeotemperature record to be constructed from AD 1859 to the present. Glacial features were dated using lichenometry before degree-day modelling enabled temperature reconstruction. The palaeotemperature reconstruction for the Albanian Little Ice Age glacial maximum (LIA_{GM}) suggests that temperatures were 0.9^{o}C below the 1980 – 2008 annual temperature mean. This work also provided the first record of glacial extent during the LIA in Albania, indicating that the Albania LIA_{GM} occurred ***c.*** AD 1859, around a decade after the European LIA_{GM} and two decades before that of Montenegro.

Anthropogenic indicators were used to reconstruct human activity in the catchment, which suggested that arable farming was pursued throughout the Medieval Warm Period (MWP; ***c.*** AD 800 – 1090) and continued during a period of transition to the LIA, between ***c.*** AD 1090 and AD 1300. The LIA (***c.*** AD 1300 - 1860) was characterised by an abrupt *Alnus* decline, thought to be the result of anthropogenic clearance of the floodplain and reduction of both arable and thermophilous types. During the LIA sedimentation rates were up to 1.41 $\pm$ 0.17 cm yr^{-1} at Lake Plav causing lake infilling and shallowing allowing wetland expansion ***c.*** AD 1570. The result of lake infilling is highlighted during the early 20th century, when the lake extent fell by around 42% as a result of climatic amelioration post-LIA causing lake levels to fall and wetland indicators to decline.

The inferred past climatic changes from the Lake Plav catchment are compared to data from around the Mediterranean and Southern Europe. This allows identification of the climatic influences affecting the site during the Late Holocene. Catchment records have provided evidence of cooler and wetter conditions coeval to the occurrence of solar minima such as the Wolf, Spörer and Maunder minima. Overall, the records suggest that continental atmospheric circulation patterns such as the North Sea – Caspian Pattern (NCP) and East Atlantic – West Russia pattern (EA-WR), dominated the site until the late 1800s, when records become more synchronous with the NAO index and Mediterranean/Southern European data.

Declaration: No portion of the work referred to in this thesis has been submitted in support of an application for another degree or qualification of this or any other university or other institute of learning.

COPYRIGHT STATEMENT

ACKNOWLEDGEMENTS

Many people have enabled the completion of this thesis over the past 5 years by providing me with support and encouragement. I would firstly like to thank my supervisors: Dr Jeff Blackford, Prof. Jamie Woodward and Dr Phil Hughes who have enabled many of the field expeditions to the site and provided both discussion and comments on my work. Grants were vital to cover the various costs incurred and the following are thanked; Postgraduate Research Fund (RGS - IBG), The Zochonis and The H.E. David travel awards (The University of Manchester) and the New Research Workers Award (QRA). Finally, without the award onto The Graduate Assistants Scheme from the School of Environment and Development this PhD could not have been completed.

The Montenegrin Geological Service coring team and Dr Bob Whallon are thanked for help extracting core LPCB. I pay particular thanks to Kathryn Adamson who was a willing (?) field assistant, who helped double-bag samples into the wee hours. Various contacts must be thanked for their inputs to this project including; Robert Aleksić (The United Nations Development Programme, Montenegro), Goran Petrusic (Montenegro Red Cross), Ahmet Rekovic, Planinarskog Drustva Prokletije Plav, the many locals who offered food and 'kafu' during fieldwork and Skender Šarkinović. Two invaluable contacts have been Aleksandar Petrovic and Kamil Weinberg whose knowledge, resources and discussions have been greatly appreciated!

I pay particular thanks to Dr Jackie Pates at The University of Lancaster who enabled Pb^{210} and Cs^{137} dating of various materials and Dr David Horne at Queen Mary's London for assistance in Ostracoda identification. Staff at The University of Manchester must also be thanked for their continued encouragement, provision of resources and employment (John!) throughout my research; John Moore (safety first, safety second), Dr Pete Ryan, Dr Tim Allott, Morag Robson, Karl Hennermann, Dr Martin Hess, Dr Sarah Lindley and Jonathan Yarwood. Banter and useful bits of information from many postgrads at the university has made this an enjoyable experience and in particular I thank, Ross, Kathryn, Gen and Jen.

My friends and family have been the most constant source of support and encouragement, especially when the going got tough. Jenny, Laura, Laurice, Sakina and Sarah have been the best of friends and I am indebted to them all. A special thank you to Lamy, who managed to provide much of the encouragement and perspective needed, even when I got a bit (?) ratty – a tough job! Finally, my Mum, Dad and TB are thanked for putting up with me, providing a roof over my head, their continual support and many biscuits! Everyone has been wicked, so I owe you all a massive thank you and I think we can safely say this thesis is my family and friends, as much as it is mine.

LIST OF ABBREVIATIONS USED

LIA – Little Ice Age

MWP – Medieval Warm Period

NAO – North Atlantic oscillation

WeMO – Western Mediterranean Oscillation

NCP – North Sea – Caspian Pattern

LMM – Late Maunder Minimum

TAP – Total arboreal pollen

ELA – Equilibrium line altitude

PS – Particle size

PSA – Particle size analysis

GISP2 – Greenland Ice Sheet Project 2

NATC – North Atlantic thermohaline circulation

^{210}Pb – lead-210

^{137}Cs – caesium-137

LOI – loss-on-ignition

TLP – total land pollen

AP – arboreal pollen

DDM – degree-day model

MAT – mean annual temperature

NAdDW – North Adriatic Dense Water

LIW – Levantine Intermediate Water

EA–WR - East Atlantic – West Russia pattern

LIA_{GM} – Little Ice Age glacial maximum

1.1 INTRODUCTION

'Palaeoclimatic studies use changes in climatically sensitive indicators to infer past changes in global climate on time scales ranging from decades to millions of years' (IPCC, 2007).

By using indicators sensitive to both climatic and environmental change, this study aims to create a new late Holocene palaeoenvironmental record for the Eastern Mediterranean, inferring changes specific to the Little Ice Age (LIA; AD 1350 - 1850) from the Lake Plav catchment, Prokletije Mountains in Montenegro and Albania. The research not only aims to uncover the climatically driven changes that occurred during the late Holocene and LIA, but also the anthropogenically-driven variations within the catchment by bringing together complementary datasets. A multi-proxy approach using lithostratigraphic and biostratigraphic data from lacustrine sediments allows the reconstruction of climatic variations and land use. Glacial geomorphological mapping is also employed to provide a catchment specific palaeotemperature record.

1.2 WHY IS LATE HOLOCENE ENVIRONMENTAL CHANGE IMPORTANT?

The climate across the world is likely to change in the coming centuries as human activity increases greenhouse gases. These disrupt the world's energy balance, resulting in raised surface temperatures and variations in precipitation patterns (IPCC, 2007). Evidence of how environments were affected by past changes could aid in understanding how future climate change will affect such systems. While Holocene climate has been variable throughout, one of the most widespread and important perturbations was the LIA, a relatively recent (***c.*** AD 1350 – 1850) period of generally cooler and wetter conditions. Most reconstructions of the LIA have focused on changes in temperature, precipitation and anthropogenic manipulations of the landscape, concentrating on particular systems either hydrological or glacial (e.g. Pfister, 1980; Pons and Reille, 1988; Kundzewicz and Parry, 2001; Valero-Garcés *et al.*, 2006; Hughes, 2007; González Trueba, 2003/2004, 2005;

Morellón *et al.*, 2008; Ivy-Ochs *et al.*, 2009). Little has been published discussing how vegetation had been affected by LIA climatic instability and whether climatic or anthropogenic modifications could be differentiated in the environmental record. As the Bradley and Jones (1992) quote suggests, the way in which a system responds to natural forcing events must be understood, as future events may be a combination and possible exacerbation of both natural and anthropogenic forcing factors.

‘Whatever the anthropogenic climatic effects may be in the future, they will be superimposed on a climatic system which also responds to “natural” forcing factors’ (Bradley and Jones, 1992 p. 1).

Ecological studies suggest that high mountain environments provide an opportunity to uncover an amplified climate signal, as montane vegetation in these environments have been shown to be more responsive to climatic change than lowland vegetation (Beniston *et al.*, 1997; Birks and Ammann, 2000). Therefore, the impacts of late Holocene climatic variability may be amplified in mountain environments. Mountain areas of the Balkans, such as the Pindus Mountains, were less affected by extensive human modification of the landscape. Garden agriculture dominated these topographically isolated highland economies throughout the late Holocene (McNeill, 1992), compared to the more populated European Alps and Pyrenees (Miras *et al.*, 2010).

1.3 RECONSTRUCTING ENVIRONMENTAL CHANGE DURING THE HOLOCENE

Environmental change in large mountain catchments is an issue that has been relatively overlooked across the Eastern Mediterranean and Balkans. The 269 km^2 Lake Plav catchment is located in an environmentally and climatically diverse area. The endemism and biodiversity of the site are disputed by various authors. Montenegro and Albania are located within the broad Mediterranean biodiversity hotspot described by Gaston and David (1994) and Mittermeier *et al.* (1998), but lie outside the area defined as the ‘Mediterranean basin’ by authors such as Médail and Quézel (1997, 1999). However, Médail and Quézel (1997, 1999) and Blondel and Médail (2009) mention 10 biodiversity hotspots within the Mediterranean basin area, overlooking Montenegro. This was a result of the study aim; to prioritise conservation. Therefore, only small areas of particular biodiversity within the region were identified and focused upon to provide a more 'practical' division for means of conservation and management in the Mediterranean (Médail and Quézel, 1999; see Figure 1.1). Indication of Montenegrin and Albanian biodiversity has been further alluded to, as Quaternary glacial refugia have been suggested to maintain vegetation populations and biodiversity (e.g. Tzedakis, 1993; Tzedakis *et al.*, 2002; Petit *et al.*, 2003), with Montenegro presented as a site of such refugia by Médail and Diadema (2009) and Hughes *et al.* (2010). Biodiversity may then be sustained by the variety of animals that pass through the area due to the migration lines they use to cross from continental Europe and Africa (e.g. Gaston and David, 1994; Ljubisavljević *et al.*, 2007; Hanson *et al.*, 2008).

Climatically, the Lake Plav catchment is atypical of a Mediterranean site due to the climatic juxtaposition between a Mediterranean and Continental climate in the area. Quaternary records from the Eastern Mediterranean have shown the area to be highly sensitive to change and an important site for glacial refugia maintaining Mediterranean biodiversity (Figure 1.2**,** e.g. Tzedakis, 1993; Denèfle *et al.*, 2000; Tonkov *et al.*, 2002; Lawson *et al.*, 2005; Médail and Diadema, 2006; Tonkov *et al.*, 2006; Fouache *et al.*, 2001, 2010).

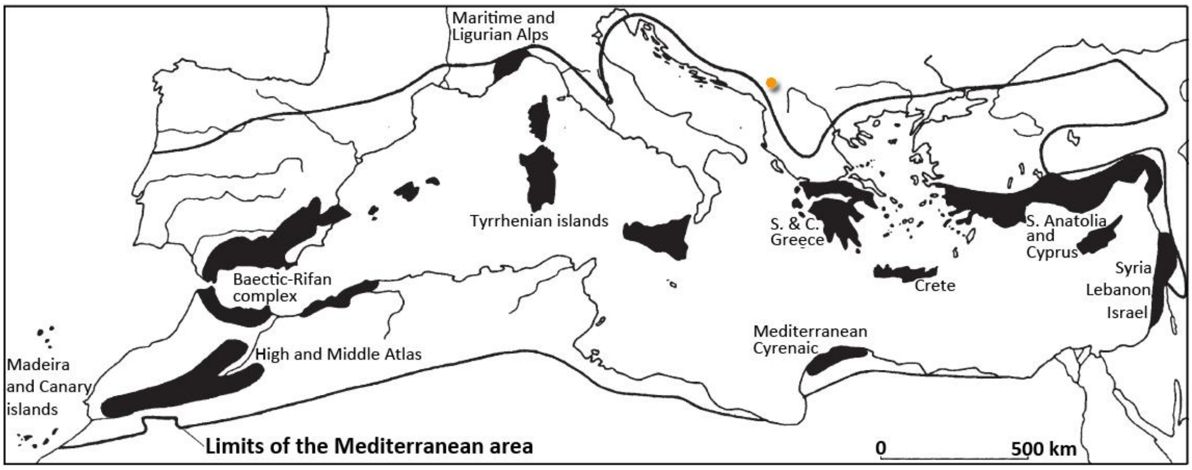


Figure 1.1: Top 10 hotspots in the Mediterranean basin following Médail and Quézel (1997), taken from Médail and Quézel (1999). Lake Plav catchment.

1.3.1 Mediterranean glacial histories

Glacial histories across the Mediterranean have confirmed the varied nature of the LIA and magnitude of glacial advance during this period. These have been concentrated in the European Alps and Pyrenees with additional studies in northern Spain, Italy and the Maritime Alps (Figure 1.2**,** Gellatly *et al.*, 1994; Grove and Gellatly, 1995; Grove, 1988; Alonso and González-Suárez, 1998; D'Orefice *et al.*, 2000; Pallás *et al.*, 2006). Recently, Hughes (2007, 2009a) produced a LIA glacial history for the Durmitor Massif (Figure 1.2). The presence of modern day glaciers in the Prokletije Mountains was noted by Hughes (2009b), but as of yet there are no LIA reconstructions for these glaciers. Figure 1.3 illustrates the 'gap' in spatial spread of palaeoenvironmental and LIA glacial reconstructions for the wider Mediterranean region. Albanian and particularly Montenegrin research has been sporadic, with glacial studies and archaeological excavations dominating. These studies have generally dealt with Pleistocene glaciations and the middle Pleistocene to middle Mesolithic periods (e.g. Cvijić, 1907, 1913, 1917; Morley *et al.*, 2006; Hughes, 2008, 2009a, b; Milivojević *et al.,* 2008; Bakovic *et al.*, 2009; Morley and Woodward, 2011).

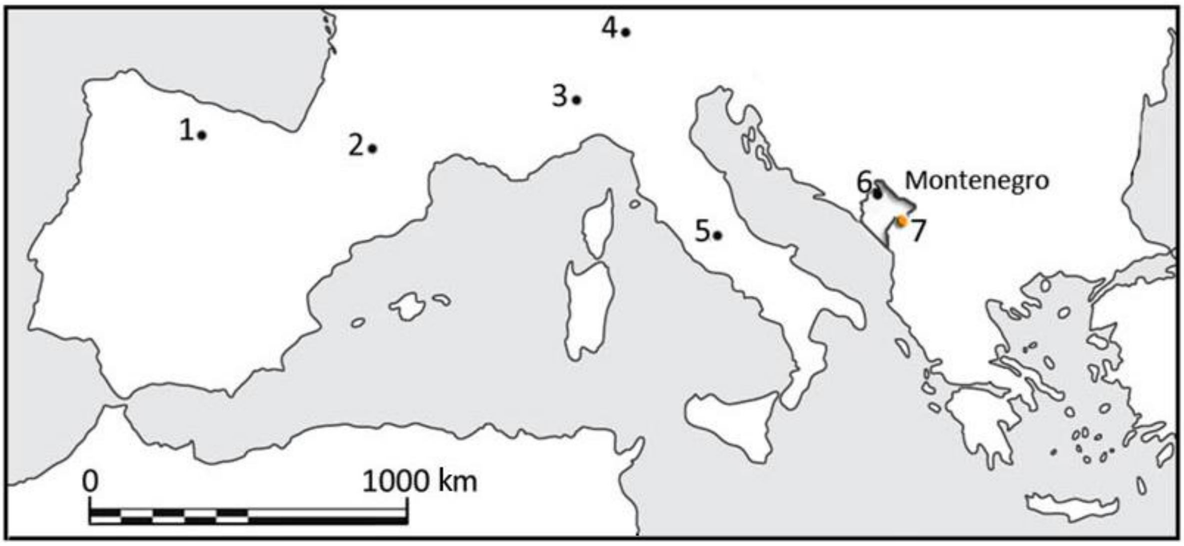


Figure 1.2: Sites of current LIA glacial records, (1) Picos de Europa, (2) Pyrenees, (3) Maritime Alps, (4) European Alps, (5) Gran Sasso Massif and (6) Durmitor Massif. ● Site of glacial records but with no LIA reconstruction (7) Prokletije Mountains (various sources).

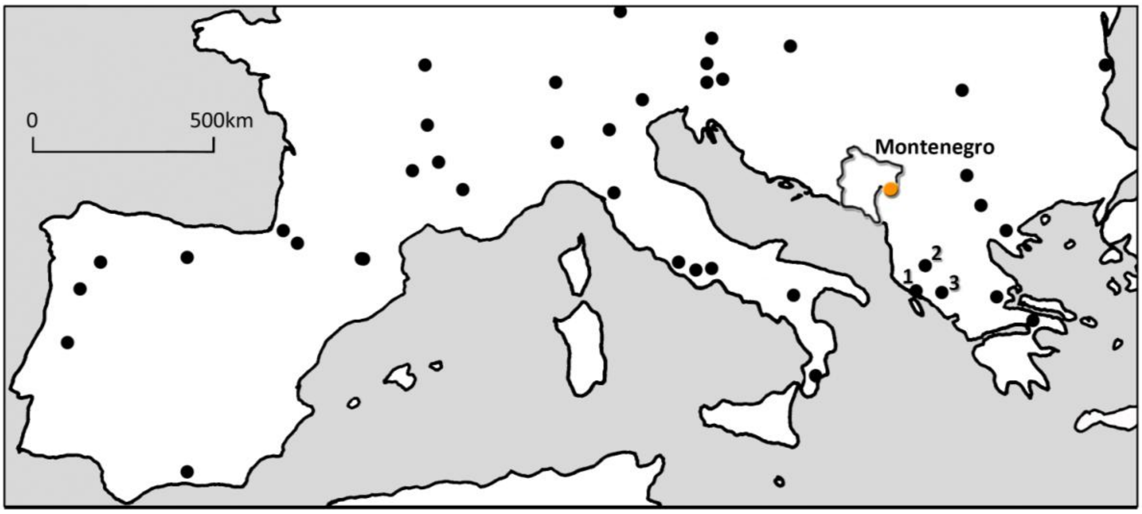


Figure 1.3: Selection of Quaternary and Holocene palaeoenvironmental records available across the Mediterranean (adapted from Bennett *et al.*, 1991), ● represents the Lake Plav catchment, (1) Lake Butrint, (2) Lake Maliq and (3) Ioannina basin.

1.3.2 The Mediterranean palaeoenvironment

As yet, a fully integrated palaeoenvironmental study, linking glacial records and proxy records from lacustrine sediments has yet to be completed in the Balkan/Eastern Mediterranean region. However, the study by Jalut *et al.* (2000) in the Western Mediterranean did bring these elements together for the Holocene period as a whole, with low resolution during the LIA. Contemporary vegetation cover and pollen rain data were associated with current Western Mediterranean climatic zones. Pollen ratios were then calculated along a transect to provide association with particular contemporary dominant vegetation, and thus climate. This enabled reconstruction of Holocene climate conditions using fossil pollen ratios. These data identified the presence of six periods of vegetational change as a result of climatic variation. Each period was assessed and correlated with data independent of anthropogenic activity, including alpine glacier and lake level fluctuations. A major vegetational shift was identified at AD 700 - 1250 relating to the Medieval Warm Period, although there was no direct reference to vegetational changes as a result of LIA onset. Correlation of the data created by Jalut *et al.* (2000) to lake level data of Magny (1999) identified 'a cooling phase at 300 cal. yrs BP', that corresponded to a major lake level rise in the Jura and French subalpine datasets (and corresponds to the glacial advance described in Grove, 1988). However, low sampling resolution did not allow a detailed assessment of pollen ratio changes throughout the LIA. Overall, the study highlights success in the comparison of fossil vegetation records with anthropogenically independent records, such as glacial records, to identify climatic deterioration/ameliorations and assess the impact upon vegetation. This emphasises the potential of integrating vegetational records with other types of records.

1.4 WHY MONTENEGRO/ALBANIA AND THE LAKE PLAV CATCHMENT?

Figure 1.3 highlights the lack of substantial palaeoenvironmental research in the Eastern Mediterranean/Balkan region north of Albania, and particularly Montenegro. Previous palaeoenvironmental records from the Mediterranean area have concentrated on long records, covering the Quaternary or Holocene periods. These do not discuss the impact of recent climatic variations on ecosystems. Although previous studies have suggested that pollen records are sensitive to climate change, and can provide an accurate record of how ecosystems alter during cooler/drier and warmer/wetter climate oscillations (Guiot *et al.*, 1989; Tzedakis, 1993; Tzedakis *et al.*, 1997; van der Wiel and Wijmstra, 1997; Lawson *et al.*, 2005). Investigations at Lake Maliq, Korcë basin, Albania (e.g. Denèfle *et al.*, 2000; Bordon *et al.*, 2009; Fouache *et al.* 2010) have concentrated on the Holocene and used pollen analysis as a proxy for vegetation, temperature and precipitation variation during the Holocene. Fouache *et al.* (2001, 2010) identified the influence of humans on the environment around Lake Maliq, by an increase in cereal types between 4255 ± 50 and 2420 ± 45 yrs BP. Research in the area has demonstrated regular climatic perturbations throughout the Holocene, and inference of past temperature and precipitation by Bordon *et al.* (2009) was stretched to indicate seasonal variations in lake level from the pollen data collected. Fouache *et al.* (2001, 2010) created a reconstruction of past lake extents at Lake Maliq between 14000 and 2000 yrs BP, described as 'palaeogeographical reconstructions'. This work relied on lithological changes to identify lake level variations and allowed GIS modelling of the past lake level. Research into natural and human-induced change at Lake Butrint on the Southern Albanian coast (Ariztegui *et al.*, 2010) sought to discuss more recent environmental change over the last 300 years. Ariztegui *et al.* (2010) used geochemical and lithological analyses to assess the natural and human induced environmental changes at Lake Butrint, suggesting that 'human occupation in the catchment only partially masked the climate record' of the sediments. The natural and human forcing factors were then said to be 'disentangled' with the use of stable isotopic and stratigraphic analyses. However, varve widths were used to determine the natural forcing factors, as the isotopic signal was influenced by anthropogenic activity and could not be used for climate assessment. The success of this method is questioned as correlation between lamina thickness and the NAO index was found to be poor and the annual variability of the NAO may make annual trends difficult to decipher.

1.4.1 Hydrological and climatic settings in the Mediterranean

River behaviour has been widely studied in the Mediterranean (e.g. Macklin *et al.*, 2006; Macklin and Woodward, 2009), but most sites have been steep gorge systems, such as the Ardena and Rapanas Gorges in Crete and Flumineddu Canyon in Central-East Sardinia (Maas and Macklin, 2002; Waele *et al.*, 2010). These high energy rivers are characterised by steep relief and coarse grained sediments. Flood histories for sites such as Ardena and Anapodaris Gorges in Crete, have provided both late Holocene and Quaternary long records for these steep environments (Maas and Macklin, 2002; Macklin *et al.*, 2010). Liquete *et al.* (2005) provided a record of water discharge for the coastal region of Andalusia in Spain indicating that the majority of rivers in the area were 'torrential', demonstrated by large mean axial gradients. Torrential precipitation events characterise eastern and central Andalusia producing continually variable water discharge and sediment transport patterns.

Unlike previously studied river systems, the River Ljuča that feeds Lake Plav (see Figure 1.4) has a low gradient and is dominated by overbank sedimentation. The Prokletije Mountains of Montenegro and Albania offer a unique study site as the high mountain region receives around 3000 mm yr^{-1} precipitation (Furlan, 1977; Ranković *et al.*, 1981; Bošković and Bajković, 2006), unlike other areas of Albania and Greece which receive between 500 mm and 2000 mm yr^{-1} (Liaccos and Mouloupoulos, 1967; Velo *et al.*, 2005). The Prokletije range experiences a "Continental Mediterranean climate" (Köppen, 1923, 1936 cited in Lohmann *et al.*, 1993; Gerstengarbe and Werner, 2009), with greater summer precipitation than other Mediterranean locations (Magaš, 2002; Tošić, 2004). The ample moisture availability helps to maintain a characteristic forest cover and has provided sufficient moisture to ensure the survival of glaciers to the present day (Garston and David, 1994; Hughes, 2009a).

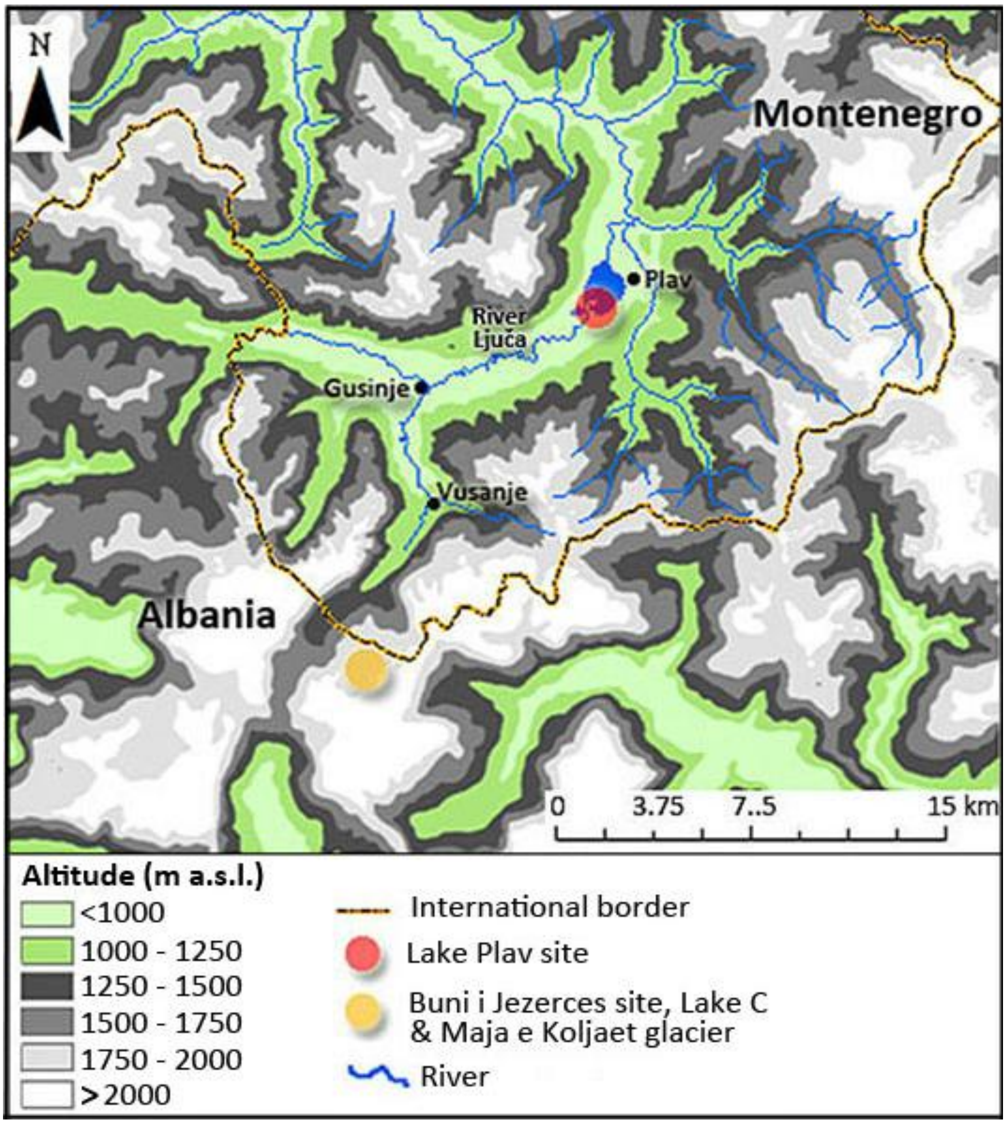


Figure 1.4: Site locations at the Lake Plav catchment.

The Prokletije Mountains are also known as the “The damned/cursed mountains”, Albanian Alps or Bjeshkët e Nemura (Milivojevic *et al.*, 2008). They lie between the Dinaric Alps and Pindus Mountains, straddling the borders of three countries; Montenegro, Albania and Kosovo. The highest peak Maja Jezerce (2694 m a.s.l.) is situated within Albania and the landscape history of the Prokletije has been one of glacial modification with large U-shaped valleys dominating the area, a remnant of Pleistocene glaciations (Auer, 2007; Milivojevic *et al.*, 2008). A series of limestone ridges makes up the Prokletije mountains (McNeill, 1992), which have a karst topography including glaciokarst and corrosion terraces on ‘*roches moutonees’* (Kunaver, 1991). The karst topography of the area has been profoundly affected by the high annual precipitation (Magaš, 2002), with vertical dissection of the relief producing elevation differences of up to 1800 m (Milivojevic *et al.*, 2008). Within the Lake Plav catchment, high elevation archives and lower lying lake deposits are found in a relatively short horizontal distance of 18 km, with around 1500 m vertical difference.

1.5 RESEARCH AIMS

This project aims to reconstruct late Holocene environmental changes in a large mountain catchment in the Northern Prokletije Mountains of Montenegro and Albania, to investigate the effects of climatic and land use changes, with a particular focus on the LIA.

1.6 RESEARCH OBJECTIVES

The research objectives are to reconstruct previous environmental and climatic fluctuations by:

1) Reconstructing vegetation and environmental changes from sediment cores, using a variety of proxies to capture local and catchment-scale changes.

2) Determining LIA glacial extent and subsequent glacial fluctuations to provide a catchment specific temperature record through to the present.

3) Examining the palaeoenvironmental changes at a high altitude site where climatic forcing is likely to be amplified and anthropogenic factors less influential.

4) Synthesising the data collected to provide an integrated climate, land use and sedimentary history for the Lake Plav catchment.

5) Analysing this integrated record in the wider context of LIA environmental change.

Following preliminary investigations within the Prokletije Mountains, the Lake Plav catchment was identified as a site that combined a large valley bottom lake (Lake Plav, 904m a.s.l.), with a high altitude lake at or above the treeline (within Buni I Jezerces ~1750m a.s.l), and glacial records from small contemporary glaciers around the Maja e Koljaet (2490 m a.s.l) and Maja e Jezerces (2694 m a.s.l.) peaks (see Figure 1.4). As such an independent climate record could be obtained and used to compare and assess the palaeoenvironmental records of the last 1400 years. In the catchment clear altitudinal zonation of vegetation was apparent and agriculture was spatially and altitudinally confined to valley bottoms and the lower hillsides (below 1000 m a.s.l.). The Lake Plav catchment therefore fulfilled several of the requirements for the completion of this study following the aims and objectives outlined above.

1.7 THESIS STRUCTURE

This thesis has been divided into seven chapters. Chapter one outlines the research project alongside the main aims, objectives and approaches used throughout the study. Chapter two is sub-divided into five sections, with the aim of putting the Little Ice Age in context. Each section provides a review of different records available within the Mediterranean and Southern European region and, where necessary, other areas relating to the period since the LIA. Chapter three aims to provide a geographical and regional context and sense of place for this study. The methodologies used during the research and study sites are discussed in chapter four and chapter five presents the results for the Lake Plav catchment in eight sections. Chapter six brings together the data collected and discusses the results as applied to previous studies within the Mediterranean region and the project aims and objectives. Finally, chapter seven will draw the main conclusions found during the research project and offer suggestions of further possible study at the site.

The mid to late Holocene was a period of significant climatic variability with a number of climatic deteriorations and ameliorations (Grove, 1979; Bell and Walker, 1992; Alley and Ágústsdóttir, 2005). Superimposed upon this, is the spread of anthropogenic activity, in particular agriculture endeavours, which were registered in Balkan palaeoenvironmental records from ***c.*** 6200 yrs BP (Willis and Bennett, 1994). This chapter summarises evidence for these palaeoenvironmental changes in the Mediterranean region and beyond.

2.1 THE 'LITTLE ICE AGE' IN CONTEXT

Since the beginning of the Holocene period there have been many climatic shifts, including the 8.2kyr BP event (Alley and Ágústsdóttir, 2005), Holocene Climatic Optimum and Medieval Warm Period (***c.*** AD 700 - 1300) (Grove, 1979; Bell and Walker, 1992). The climatic amelioration known as the Medieval Warm Period has in general been recorded as a period of increased temperature and aridity (Lamb, 1965; Campbell, 1998; Cronin *et al.*, 2003; Kremenetski *et al.*, 2004). The climatic deterioration, termed the 'Little Ice Age,' after Matthes (1939) suggested that the period began in the early 1300s, but this date has been contested. Throughout the literature various timings for the LIA are provided. Porter (1986) suggested the period AD 1250 - 1920, whilst Lamb (1977) and Jones and Bradley (1995) proposed an initiation in AD 1550 and termination between AD 1700 - 1850. The longest period suggested was AD 1300 to around AD 1860 by Fagan (2000) and Le Roy Ladurie (1971 cited in Grove, 1988). World wide variations in the timing of the considered coldest periods within the LIA may have led to some of this contention, as Europe experienced these during the 1600s, in the subtropical North Atlantic and Andes the 1600s to 1700s were harsh and North America suffered in the 1800s (Mann, 2002). Furthermore, uncertainty may have been a consequence of researchers failing to constrain the LIA to 'LIA glacierisation' or 'LIA climate.' The LIA glacierisation is dated to AD 1300 - 1950 by Matthews and Briffa (2005) and said to be highly dependent on winter precipitation. During this phase the extension of glaciers in the European Alps was greater than that seen before or since. However, defining the 'LIA climate' is more complex, because unlike the preceding Medieval Warm Period which was characterised by continuous climatic amelioration, there is no evidence for a continued cold period that can be associated with the LIA. Therefore, Matthews and Briffa (2005) follow the work undertaken by Briffa *et al.* (2001) that used dendroclimatological techniques to uncover a period between AD 1570 - 1900, when summer temperatures (April to September) in the Northern Hemisphere were substantially below the AD 1961 - 1990 mean. This is a general trend and within the Northern Hemisphere there are still variations in the timings and severity of climatic cooling. The 'LIA climate' period (AD 1570 - 1900) encompassed two of the glacial maxima identified by Matthews and Briffa (2005), the first ***c.*** AD 1650 and the second ***c.*** AD 1850. Although these general trends have been determined, for the purpose of this research the LIA is defined as the period between AD 1300 - 1900 as this not only offers a period when climatic deterioration had occurred but also when glaciers were at their Late Holocene maximum stand. The scope of this period also includes the subsequent warming into the twentieth century.

The early conception of the LIA was one of homogeneous cool and wet conditions (Lamb 1965, 1977), but this has since been challenged (e.g. Jalut *et al.*, 2000; Cronin *et al.*, 2003; Trouet *et al.*, 2009). Climate variations during the LIA have been recorded in a variety of data sets including instrumental and proxy climate records such as geomorphological evidence for glacial and fluvial activity, charcoal, pollen, ostracoda and stalagmite records (e.g. Grove, 1979, 1988; Hodell *et al.*, 2005; Frisia *et al.*, 2005; Pederson *et al.,* 2005; Hughes, 2007). Proxy records suggest that the LIA was characterised by an increase in yearly precipitation and a thermal decline of ***c.*** 0.9^{o}C and as much as 2^{o}C in Scotland, contributing to glacial advances in all glaciated areas of the world between the 1500s and 1800s (Matthes, 1939; Grove, 1979, 1988; Lamb, 1995). Historical documentation available for the LIA period generally focuses on lowland areas. These data concentrated on the socio-economic impacts of climate change and its consequences, with evidence of European crop failures in the 1300s and 1700s coeval to widespread food crises (Fagan, 2000). In contrast, the high mountain environments of the Mediterranean region register

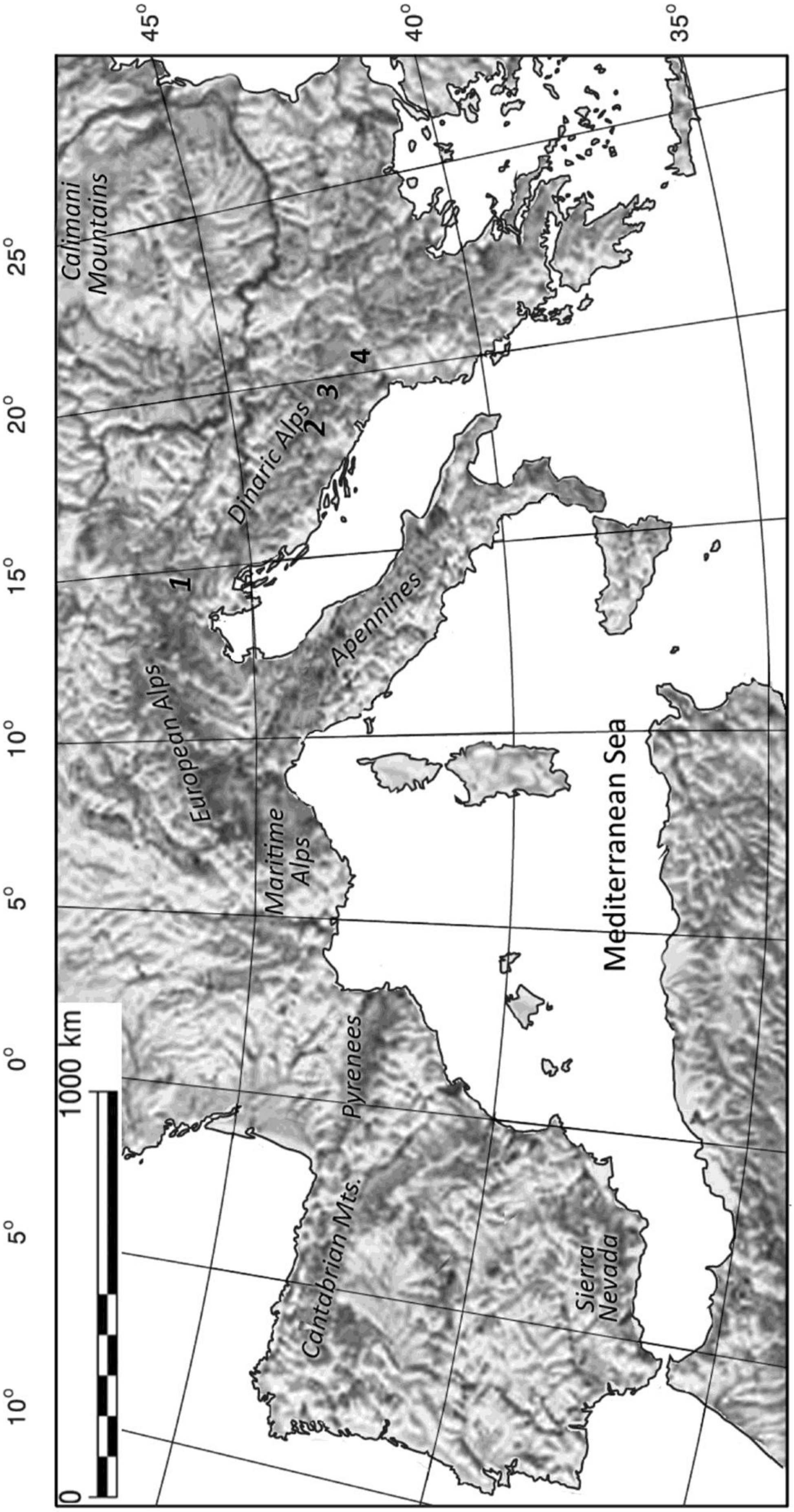


Figure 2.1: Map of Europe locating areas of late Holocene glacial geomorphological and dendrocliamtological study: (1) Skuta and Triglav glaciers, Slovenia, (2) Jahorina, Bosnia Herzegovina, (3) Durmitor Massif, Montenegro and (4) Lake Plav, Prokletije Mountains, Montenegro.

the development and growth of small high mountain glaciers as the most significant consequence of LIA environmental change and records of these transformations are particularly detailed in the Western Mediterranean (Figure 2.1, Grove, 1988; González-Trueba, 2008). Research in the Balkan region has focused on longer-term climatic variability, overlooking the LIA. Although evidence in other Mediterranean records support and identify the LIA and the short abrupt climate fluctuations that occurred throughout the period. The sites used for this research have been located in low level to high mountain areas and provide diverse records such as flood histories, geomorphological evidence, and biostratigraphic and lithostratigraphic sequences (e.g. Crowley and Lowery, 2000; Ghosh, 2006; Goosse *et al.*, 2006; Marquer *et al.*, 2008).

2.2 CLIMATE RECORDS FOR THE LITTLE ICE AGE PERIOD

Historical documentation and particularly climatic reconstructions have provided climate data since the beginning of the LIA and quantified the variability of this period with reconstructions of temperature, pressure and rainfall (Pfister, 1980; Kundzewicz and Parry, 2001). These data have challenged the early notion of a homogeneous cool and wet LIA (Lamb 1965, 1977), as evidence of wetter and cooler conditions, interspersed by warmer, drier periods throughout the period were detected (e.g. Jalut *et al.*, 2000; Cronin *et al.*, 2003; Trouet *et al.*, 2009).

2.2.1 Palaeoclimatic records for Europe

A variety of influences create variations in climate, but across Europe the North Atlantic Oscillation (NAO) is one of the primary modes of climatic variability in the region (e.g. Lamb and Peppler, 1987; Trigo *et al.*, 2002; Jones and Briffa, 2006). The NAO is a source of inter-annual variability within the atmospheric circulation of the North Atlantic, affecting the weather conditions (temperature, precipitation and wind) seen across Europe (Hurrell, 1995: Luterbacher *et al.*, 1999; Trigo *et al.*, 2002; García-Sellés *et al.*, 2010). Decadal trends have been identified within the NAO by Hurrell and van Loon (1997; Figure 2.2), with a dipole pattern across Europe. A positive NAO phase is associated with warmer conditions over Southern Europe and cooler conditions in Scandinavia (Hurrell, 1995; Hurrell and van Loon, 1997). This pattern has been documented in many studies, with Luterbacher *et al.* (2002) presenting the longest reconstruction back to AD 1500 using sea level pressure, temperature and precipitation data (Figure 2.2). Uncertainties in the Luterbacher *et al.* (2002) model decrease after AD 1820 following the addition of further datasets and this can be seen by the large standard errors included in Figure 2.2. Within the trend, strong decadal and inter-decadal patterns were identified, with positive NAO identified at the beginning of the 1500s, 1700s, ***c.*** AD 1850, 1900s and the end of the 1900s. Cook *et al.* (2002) provide a second reconstruction of winter NAO patterns since the 1500s, validated against independent estimates, such as dendrochronological records (Figure 2.2). Waveforms were identified in the NAO data by authors such as Cook *et al.* (1998), who suggest three terms at 24 years, 8 years and 2.1 years, with the latter two being weaker during the 1900s. Periods of reduced NAO index (< 0; Figure 2.2) can be associated with glacial advances recorded across Europe in the late 1800s, ***c.*** 1915s, 1930 to 1940s, 1960s to 1970s and 1980s. Across the Mediterranean, a positive NAO index is associated with dry periods, as anticyclones across Europe and/or the Eastern Mediterranean impede the movement of low-pressure systems (associated with increased precipitation) from the west (Xoplaki *et al.*, 2001; Luterbacher *et al.*, 2001). Xoplaki *et al.* (2001) conclude that the periods AD 1675 - 1715 and AD 1780 - 1830 in the Eastern Mediterranean were cooler, wetter and overall more variable than the period 1961 - 1990. This could be the result of high pressure in Northern Europe combined with lower pressure over the Eastern Mediterranean i.e. a negative NAO, which resulted in these cooler wetter conditions (Hurrell, 1995; Xoplaki *et al.*, 2001). A negative NAO index has been cited as an influence upon avalanche activity in the Pyrenees by García-Sellés *et al.* (2010) as it is associated with increased precipitation and stronger winds within the North Eastern Iberian Peninsula. However, in the European Alps Jomelli *et al.* (2007) suggested that avalanche frequency cannot be linked to this or any circulation pattern.

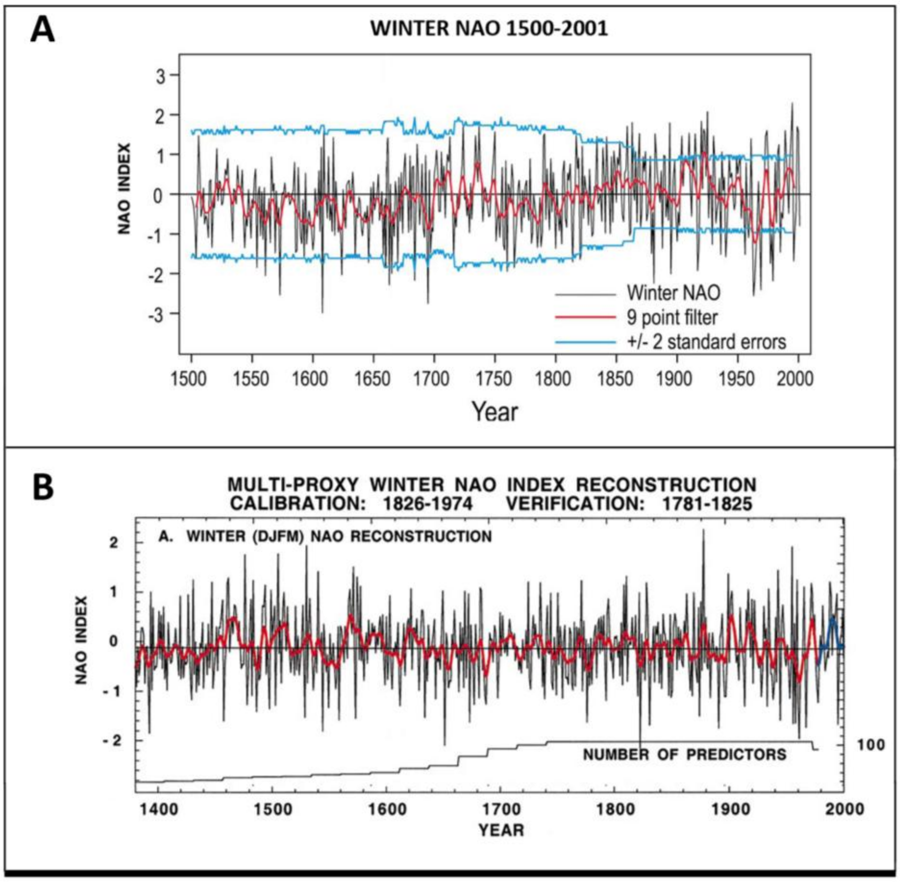


Figure 2.2: NAO index as reconstructed by (A) Luterbacher *et al.* (2002) since AD 1500 (monthly variations to AD 1659 and seasonal variations to AD 1500) and (B) Cook *et al.* (2002) since AD 1380.

The WeMO (Western Mediterranean Oscillation) is a second atmospheric oscillation to affect the Mediterranean, and is a 'low frequency' atmospheric pattern accountable for precipitation variations across the Iberian Peninsula (López-Bustins, 2007 cited in García-Sellés *et al.*, 2010). Kutiel *et al.* (2002) discuss a third atmospheric circulation that affects the Eastern Mediterranean, the North Sea – Caspian Pattern (NCP). A negative NCP was associated with above normal temperatures in the region as a 'south-westerly anomaly circulation' became more dominant, whilst a 'dominant north-easterly anomaly circulation' was associated with suppressed temperatures. The influence of the NCP upon rainfall was thought to be limited as many local factors contribute to localised precipitation types (Kutiel *et al.*, 1996), but precipitation was found to be significantly greater during periods of positive NCP (Kutiel *et al.*, 2002). The impact of the NCP has since been identified in Iran where a positive NCP is linked to increased precipitation and cloud cover resulting in reduced temperatures (Ghasemi and Khalili, 2008). Studies on its effect upon heat fluxes in the Mediterranean, Black and Caspian Seas have also been investigated (Gündüz and Özsoy, 2005). Overall, three atmospheric circulation patterns may affect the Eastern Mediterranean, but the longest records available are those of the NAO, providing information on positive and negative index years throughout the majority of the LIA.

Bond *et al.* (1997) reintroduced the idea of a 1500 year cycle of climatic change in the North Atlantic following the first suggestion of such cycles by Denton and Karlén (1973). Using the presence of hematite stained grains within ice core records, Bond *et al.* (1997) were able to identify a 1536 $\pm$ 563 year cyclic system, which corresponded to geochemical records of the Greenland Ice Sheet Project 2 (GISP2). These and other cyclic events, such as Dansgaard/Oeschger events ($\delta^{18}O$ shifts), were found to be forced by outside influences rather than internal instabilities within ice sheets. The findings of Bond *et al.* (1997) suggested that the LIA was part of the latest cold phase in the series of 1500 year cycles seen throughout the Holocene (Rind *et al.*, 1999). The forcing mechanism has been attributed to solar variations (e.g. Bond *et al.*, 2001), such as the Maunder Minimum (AD 1645 - 1715; Eddy, 1976; Lockwood, 2001; Xoplaki *et al.*, 2001). These minima were characterised by the disappearance of sunspots causing reduced solar irradiance (Eddy, 1976; Bard *et al.* 2000). The Maunder Minimum coincided with the height of the LIA (Rind *et al.*, 1999) and was registered across North-Western Europe (e.g. Eddy, 1976, Borisenkov, 1994; Stuiver *et al.*, 1998; van der Linden and van Geel, 2006). The period between AD 1645 - 1715 also corresponds to a phase of glacial advance across the Western Mediterranean (Russell, 1908; Alonso *et al.*, 1983 cited in González Trueba *et al.*, 2008; Grove and Gellatly, 1995) and was characterised by cool winters and greater summer precipitation in Western Europe (Pfister, 1994, 1999). By the end of the 1600s the climatic low had spread to the rest of Europe. Lockwood (2001) reviewed sudden and abrupt climatic fluctuations, considering the 1500 year cycle. This work highlighted the influence of the North Atlantic thermohaline circulation (NATC), illustrating that if the NATC slowed abrupt climatic changes would occur in Europe. Within the 1500 year cycle, Wanner *et al.* (2000 cited in Lockwood, 2001) identified a 200 to 400 cycle whereby conditions encouraged or discouraged glacial expansion. Research using proxy reconstructions from European peat accumulations have provided evidence for three further 'minima' during the LIA, namely the Wolf (***c.*** AD 1280 - 1342), Spörer (***c.*** AD 1460 - 1550) and Dalton (***c.*** AD 1790 - 1830) minima (Stuiver and Braziunas, 1995; Rigozo *et al.*, 2001; Mauquoy *et al.*, 2002; Wagner and Zorita, 2005). These were similar to the Maunder minimum as they were periods of low solar activity characterised by increased wetness and cooling in the Northern Hemisphere.

Overall, the solar anomalies and climatic oscillations discussed here go some way to describing the reasons for climatic variability during and since the LIA. The climatic fluctuations that occurred throughout the period are discussed further in section 2.4.

2.2.2 Dendroclimatological records from Europe

Tree rings are a source of easily accessible and widely distributed proxy climate data that provide a valuable method for reconstructing both present and past climate shifts (Fritts, 1976; Schweingruber, 1988; Sheppard, 2010). Dendrochronologies were first developed in areas of America and Europe as early as AD 1800, but these data aimed to quantify the damage caused by pollution around this time (Eckstein, 1972; Baillie, 1982). As the science progressed, the relationship between climate and growth became the focus and this was most successful using coniferous types (Fritts, 1971). Although the Mediterranean is rich in tree species, few have been shown to be suitable for dendrochronology. Trees are required to be climatically sensitive to a particular factor to ensure accurate reconstruction of e.g. temperature, which is not always easily defined in the region. Therefore, dendroclimatological data that traces the climatic variations experienced during the LIA are concentrated in Northern/North-Western European and Scandinavian sites, extending to 6,700 BC in Germany (e.g. Brehme, 1951; Schweingruber, 1988; Schweingruber *et al.*, 1991; Becker, 1993; Büntgen *et al.*, 2005). These data have been built upon to create pan-European networks of dendroclimatological data (Kelly *et*

European network		Alpine region	
cold	warm	cold	warm
			1070 - 1220
		1250 - 1320	
			1460 - 1570
1530			
		1571 - 1640	
1708/1709			
1730			
	1772	1780 - 1810	
1790			
1800 - 1840		1810 - 1830	
1850			
	1870 - 1890		
1892 - 1910			
	1910 - 1930		
1930		1920 - 1930	
1950			

Table 2.1: Periods of cold and warm conditions from Alpine and pan-European dendrochronologies (sourced: Briffa *et al.*, 1988; Schweingruber *et al.*, 1988, 1991; Luterbacher *et al.*, 2004).

al., 2002). The development of dendroclimatological records across Europe has been slower than that of America, due to the supply of 'old' living trees and the more complex climatic relationships across Europe. This makes it difficult to distinguish between the varying climatic factors that affect a trees growth (Eckstein, 1972; Tessier *et al.*, 1997; Friedrichs *et al.*, 2009). Generally, dendroclimatic reconstructions in Europe have utilised records from living trees spanning 200 - 300 years (Eckstein, 1982) and by using historical tree rings climatic reconstructions have been extended beyond the range of instrumental climate data (Wilson *et al.*, 2004). The variations in tree ring growth and chemical properties have provided data on a variety of variables including summer temperature/precipitation and annual precipitation (Guitérrez, 1989; Tessier *et al.*, 1997;

Briffa *et al.*, 1998; Büntgen *et al.*, 2005; Frank and Esper, 2005a; Touchan *et al.*, 2007; Linderholm *et al.*, 2010).

One of the first researchers to construct a dendrochronological network across Western Europe was Schweingruber (1985), using 132 sites in 16 countries; however, this dataset only spanned the period AD 1881 - 1976. Records soon began to increase with Schweingruber *et al.* (1988) providing data predominantly from *Picea* for Scandinavia (AD 481 onwards) and the Alps (AD 982 onwards). These emphasised the warm and cool periods that occurred during the MWP to LIA transition (Table 2.1). Alpine dendroclimatological data suggested that the LIA began around AD 1350, with a general downturn in annual temperature (e.g. Lamb, 1965; Böhm *et al.*, 2001; Luterbacher *et al.*, 2004; Büntgen *et al.*, 2005; Frank and Esper 2005b). Further network data in Briffa *et al.* (1988) provided data for summer temperature patterns across Europe, using a network of 37 chronologies and trees including *Picea abies, Pinus sylvestris and Abies alba*, specifically selected for their potential temperature sensitivity (e.g. Briffa *et al.*, 1988; Briffa and Schweingruber, 1995; Frank and Esper, 2005a). One Eastern Mediterranean site was used by Briffa *et al.* (1988) and Schweingruber *et al.* (1991) to reconstruct summer temperatures, Jahorina (1700 m a.s.l.) in Bosnia Herzegovina. The Briffa *et al.* (1988) network used a 25 grid point method, whereby data were collated to provide general European temperature changes for the period AD 1750 – 1850, Schweingruber *et al.* (1991) then extended the record to 1975. Verification for the Jahorina reconstructed dendrochronology was less than 20%, as only a single site was used in the area (Briffa *et al.*, 1988). Periods highlighted as particularly warm or cold are detailed in Table 2.1. Cooler summers dominated the period AD 1812 - 1816 and the coldest occurred around AD 1930. Luterbacher *et al.* (2004) provided information on European seasonal temperature trends since AD 1500, showing that some of the harshest winters occurred in the late 1500s, late 1600s and end of the 1800s (Table 2.1). More recently *Picea abies* and *Larix decidua* (other temperature sensitive arboreal species) provided evidence of cooler summer temperatures between AD 1807 - 1814 in the Swiss and Austrian Alps (Büntgen *et al.*, 2005) consistent with the results of Schweingruber (1985) and Briffa *et al.* (1988). Data presented by Cía *et al.* (2005) for the central Pyrenees suggest that the period between AD 1814 - 1850 displayed no extreme temperature changes unlike that of Schweingruber (1985) and Briffa *et al.* (1988). This may be a function of the tree species used, as Büntgen *et al.* (2005) found warmth registered in the 1570s was more pronounced in *Pinus* records than *Larix*. Different de-trending techniques can be problematic for comparison of datasets, as seasonality or calibration techniques can lead to discrepancies between century long trends (Büntgen *et al.*, 2005). Cooler summer temperatures persisted in Northern Italy until around AD 1865 (Camuffo and Enzi, 1995), before cool conditions returned and became recurrent during the 1930s/1940s. During these cool periods conditions were conducive to glacial expansion, with a peak in glacier growth around the 1950s. A final minor cooling event occurred during the early to mid-1970s, again associated with glacial expansion, before a recent warming trend is registered throughout European records (e.g. Luterbacher *et al.*, 2004; Cía *et al.*, 2005).

The datasets produced in European and Alpine dendroclimatological studies were sampled to ensure high potential temperature sensitivity and thus regional climate forcing played an important role in the records retrieved. Within the European networks discussed, only one Balkan site was used; Jahorina, and as the majority of other sites where located in European or Scandinavian areas, the results are skewed towards regional forcing factors affecting these areas. Raw data for Jahorina was produced by Schweingruber using exclusively *Picea abies* samples, for the period AD 1736 - 1981 (Schweingruber cited in NOAA, 2008). Much of these data were then made available via NOAA, but a record of maximum latewood density was not available from the NOAA website. Many studies have found a positive correlation between temperatures and maximum latewood density (e.g. Briffa *et al.*, 1988; Schweingruber *et al.*, 1988; D'arrigo *et al.*, 2000). Only maximum ring density data by Schweingruber were available via NOAA (2008), but as the majority of ring growth occurs as latewood the information provided is likely to be skewed towards summer temperature variations (taken as April to September; Schweingruber *et al.*, 1991). Standardised maximum ring density data were not used as comparison to chronologies using different tree species was not required. These maximum ring density data were then transformed using the method detailed in Appendix I to produce Figure 2.3 A (Schweingruber cited in NOAA, 2008). Figure 2.3 B provides a record of Southern European temperature changes, which were produced using a network of data including that of Jahorina (Briffa *et al.*, 2002 cited in NOAA, 2008). The location of the tree ring density chronologies used in the Southern European record and method used to transform the data can be found in Appendix I.

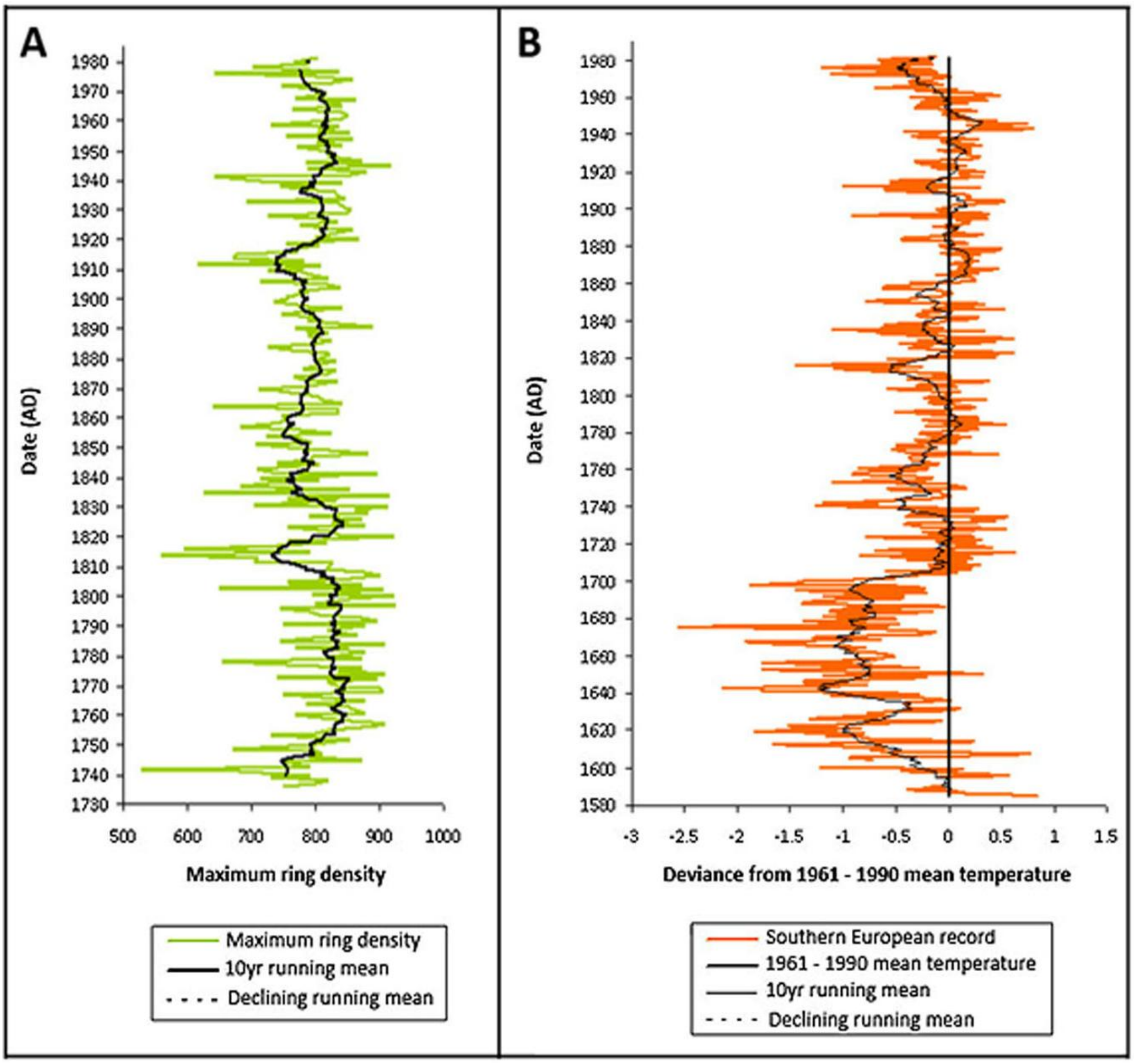


Figure 2.3: (A) de-trended maximum tree ring density for Jahorina, Bosnia-Herzegovina (Schweingruber, cited in NOAA, 2008) and (B) reconstruction of Southern European summer temperatures deviance from the 1961-1990 mean (Briffa *et al.*, 2002 cited in NOAA, 2008).

The Jahorina data provide a comparatively short record, beginning AD 1735 (see Figure 2.3 A). Maximum ring density values were lowest at AD 1742, indicating particularly cool summer temperatures. The 10 year running mean highlights rising maximum density values up to AD 1758. Values then plateau to ***c.*** AD 1810 suggesting climatic amelioration at Jahorina during the latter stages of the 19th century. This corresponds to the LIA glacial maximum (LIA $_{GM}$) in the Apennine Mountains and glacial advances in Europe (see section 2.3.1; Figure 2.7), suggesting that influences across Europe were not affecting Jahorina. A downturn in maximum ring density occurred between AD 1810 – 1820 indicating cooler summer temperatures, a trend also evident in the Calimani Mountains record (see Table 2.2). The AD 1815 Tambora eruption is the probable cause for falling temperatures, as it is thought to have affected worldwide temperatures, particularly in the following year (see section 2.3.1). Maximum ring density rose to some of the highest values during AD 1820 - 1830. Values soon decreased indicating cooling temperatures from AD 1840 - 1870, which coincided with the Western Mediterranean LIA $_{GM}$ (see Figure 2.7). Therefore, climate conditions affecting Europe were now seemingly affecting Jahorina in the Eastern Mediterranean. A second reduction occurs around AD 1910[1] coinciding with similar changes in both European and Calimani Mountain records (Popa and Kern, 2009). This highlights a period of particular cold that again affected a large area of Europe and the Mediterranean. The decline in the record from AD 1960 may be the product of divergence between tree ring density and summer temperatures, a problem identified in the late 20th century by Briffa *et al.* (1998).

Figure 2.3 B provides a Southern Europe temperature signal from AD 1590 and was created using data from Briffa *et al.* (2002 cited in NOAA, 2008). Summer temperatures during the mid- LIA were particularly low in Southern Europe, with the period between AD 1637 - 1702 at more than -0.5^{o}C deviance from the 1961 – 1990 mean summer temperature. This coincided with the first smaller LIA$_{GM}$ of Europe, as discussed in section 2.3.1. Temperatures rose around AD 1700, but remained just below the 1961-1990 mean until AD 1861 for all but two periods; AD 1787 - 1793 and during AD 1826. By AD 1870 temperatures had risen above the 1961 – 1990 mean when many glaciers across Europe were beginning to retreat from LIA$_{GM}$ positions (see section 2.3.1). A brief temperature

[1] This may relate to the AD 1912 Katmai eruption at the Alaskan Peninsula, which has been described as the largest 20th century volcanic eruption (see Hildreth and Fierstein, 2000).

depression between AD 1907 - 1917 corresponds to a cool period noted throughout climatological and glacial records in the Mediterranean and Europe (see sections 2.2.3 and 2.3.1). Following these declines, deviance from the temperature mean was generally positive, peaking at AD 1945 to + 0.75^{o}C. Reconstructed temperatures then decline from AD 1960 as a result of divergence, mentioned previously (Briffa *et al.*, 2001).

European network		Calimani Mts	
cold	warm	cold	warm
	1052 - 1250	1170	1160 - 1370
		1280 - 1290	
1250 - 1350			
		1370 - 1430	
1460s		1430	
	1470 - 1570s	↓	
			1490s
1530			
			1540s
1571 - 1640			
		1630	
			1630 - 1740
1708/1709			
1730			
		1770	
1780 - 1810	1772		
1790		↓	
*1800 - 1840			
1850		*1850	
	1870 - 1890		1850 - 1890
1892 - 1910			
	1910 - 1930	1910 - 1970s	
1930			
1950			
	~1975 to present		
			1980 to present

Table 2.2: Highlighted periods of contrasting and similar climatic data generated at European sites and in the Calimani Mountains (Sources: Briffa *et al.*, 1988; Schweingruber *et al.*, 1988, 1991; Luterbacher *et al.*, 2004; Büntgen *et al.*, 2005; Popa and Kern, 2009; *represents AD 1815, Tambora eruption).

A further dendroclimatic reconstruction from the Calimani Mountains, Eastern Carpathians in Romania (Figure 2.1), offers an alternative dendroclimatic reconstruction beginning AD 1160. Popa and Kern (2009) provided a record of summer temperature reconstructions using live and sub-fossil stone pine (*Pinus cembra* L.), that generally contrasted to Alpine reconstructions. An LIA 'fingerprint' between AD 1370 and AD 1630 in the Calimani Mountains was observed by Popa and Kern (2009) as the beginning of a climatic downturn. However, 'the longest cold epoch' occurred between AD 1430 – 1630, which is almost a century after that in the European network. There seems to be little synchrony between the European and Romanian records, with temperatures amelioration between AD 1630 – 1740. This is in stark contrast to European records where a climatic downturn occurred, thought to be the result of the Maunder minimum. Calimani records, thus suggest this period of low solar activity had little or no effect on the Romanian climate. At the beginning of the 1700s the European network records indicate cooler conditions, likely to be the result of the Maunder Minimum (AD 1645 - 1715; Eddy, 1976), but warmer conditions were reconstructed in the Calimani record. The Tambora eruption of AD 1815 (*Table 2.2; Stothers, 1984) was registered in both records, but the much discussed 'year without a summer' associated with this eruption was not felt in the Calimani record until AD 1818 (Popa and Kern, 2009). During the mid- to late 1800s both the European network and Calimani records show evidence of warmer conditions, but this parallel is lost by the 1900s when Europe registers warmer conditions and the Calimani record, cooler summer temperatures. Recent warming was not registered until the 1980s in the Calimani Mountains, around 5 - 10 years later than European records. Variations between the European network and Calimani record indicate that during periods of similarity strong large scale forcing influences both the European and Romanian record, but during contrasts periods the regional forcing factors dominate (Popa and Kern, 2009).

2.2.3 Historical climatic records

Two studies using documentary evidence are the estimated rainfall anomalies for Eastern Andalusia (Rodrigo *et al.*, 1995) and an annual reconstructed rainfall series from Toledo in Spain (Rodrigo and Barriendos, 2008), presented in Figure 2.4. The Eastern Andalusia data presents the rainfall 'deviance (mm)' from the mean precipitation between AD 1941 – 1985 (Rodrigo *et al.*, 1995). A short 200 year period is presented, but it represents the central period of the LIA, when cold conditions have been identified across Europe and the Mediterranean in dendrochronological records (e.g. Briffa *et al.*, 1988; Schweingruber *et al.*, 1991; Popa and Kern, 2009). The wettest periods are identified as AD 1510 - 1525, AD 1545 - 1570 and AD 1605 - 1651, all of which correspond to periods of suggested warmth in the European dendroclimatological network (e.g. Briffa *et al.*, 1988; Schweingruber *et al.*, 1988, 1991; Luterbacher *et al.*, 2004) and periods of cold in the Calimani Mountains (Popa and Kern, 2009). The precipitation curve from Toledo (Iberian Peninsula) was created using a variety of documentary data from Toledo before calibration was undertaken (AD 1961 - 1990 used as reference period; Rodrigo and Barriendos, 2008). The summary data created for Toledo in Figure 2.4, is interpreted as a record of high (+1; persistent intense rainfall or floods), normal (0) or low (-1; drought) precipitation. Overall, lower annual precipitation occurred at Toledo during the late 1500s, early 1600s and throughout the 1700s (see Figure 2.4). Periods of drought seem more intense at Toledo reaching an index of nearly -1.5 around AD 1610 - 1620, but this may be a product of the geographical position of Toledo in central Spain. Rather than consistent periods of drought or low precipitation at Toledo, the climate seems variable with periods of drought soon followed by periods of above average precipitation (Rodrigo and Barriendos, 2008). This continual shift between wet and dry conditions is indicative of a variable LIA climate, as previously discussed (e.g. Jalut *et al.*, 2000; Cronin *et al.*, 2003; Trouet *et al.*, 2009).

Annual temperature curves in Figure 2.5 provide a record of change since AD 1000 (Pfister, 1995; Moberg *et al.*, 2005). Moberg *et al.* (2005) used multi-proxy analysis to provide the long reconstruction (Figure 2.5 A) and continuous instrumental data was used for graph B, Figure 2.5, created by Jones and Bradley (1995). Figure 2.5 A highlights the cooling trend around AD 1150 into the LIA, but during the late 1300s a rising trend is exposed. Cooling then continued throughout the 1400s and 1500s before peaking around AD 1600, when temperatures begin to rise. During the 1800s a plateau occurs, registered as cooling/expansion events in dendroclimatological and glacial records (discussed in section 2.3). A general trend of warming is then obvious from the 1900s to present. Figure 2.5 B relies upon instrumental data alone and as no continuous instrumental record

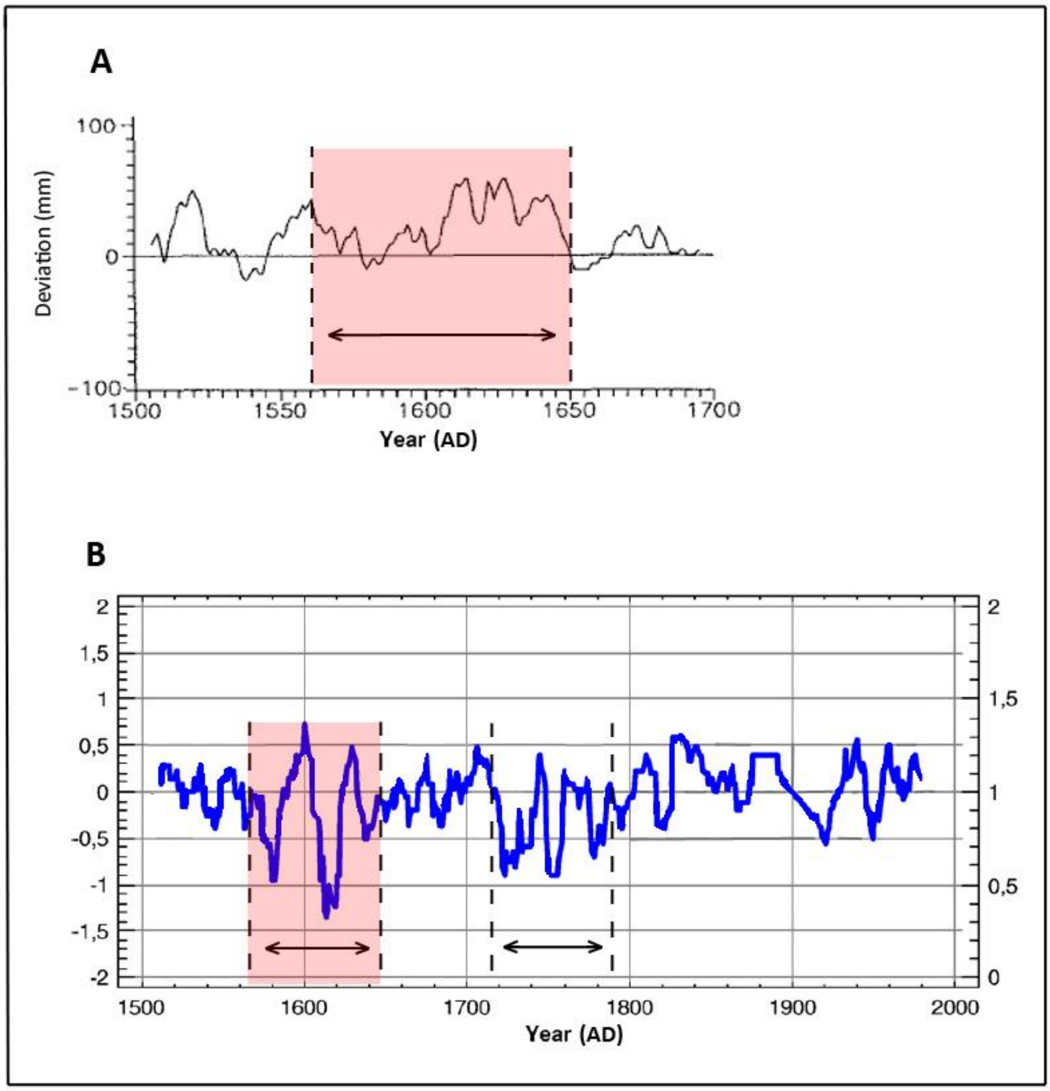


Figure 2.4: (A) Estimate of deviance from the mean AD 1941 - 1985 precipitation data for Eastern Andalusia (adapted from Rodrigo *et al.*, 1995). (B) Annual reconstructed rainfall series from Toledo, Spain (adapted from Rodrigo and Barriendos, 2008). The red shaded areas represent a comparable temporal range.

exists prior to the late 1800s, only the latter stages of the LIA were captured. The deviation from the AD 1951 - 1970 mean is nearly 0.5^{o}C throughout the latter stages of the LIA. Temperatures within the first red shaded area (Figure 2.5 B; AD 1890 - 1910) when compared to European glacial records (e.g. Grove, 1988; D'Orefice *et al.*, 2000; González Trueba, 2003/2004, 2005; González Trueba *et al.*, 2008) coincide with glacial expansion across the Mediterranean region and above average rainfall in Toledo, Spain

(Figure 2.4). Northern Hemisphere temperatures continually rise, until AD 1960 (second shaded area Figure 2.5 B) when temperatures plateau before declining up to AD 1980 and a steeper temperature rise occurs to AD 1990. More recent publications confirm that this rise in temperature has continued to the present day (e.g. Folland and Karl, 2001; Yeh and Kim, 2010).

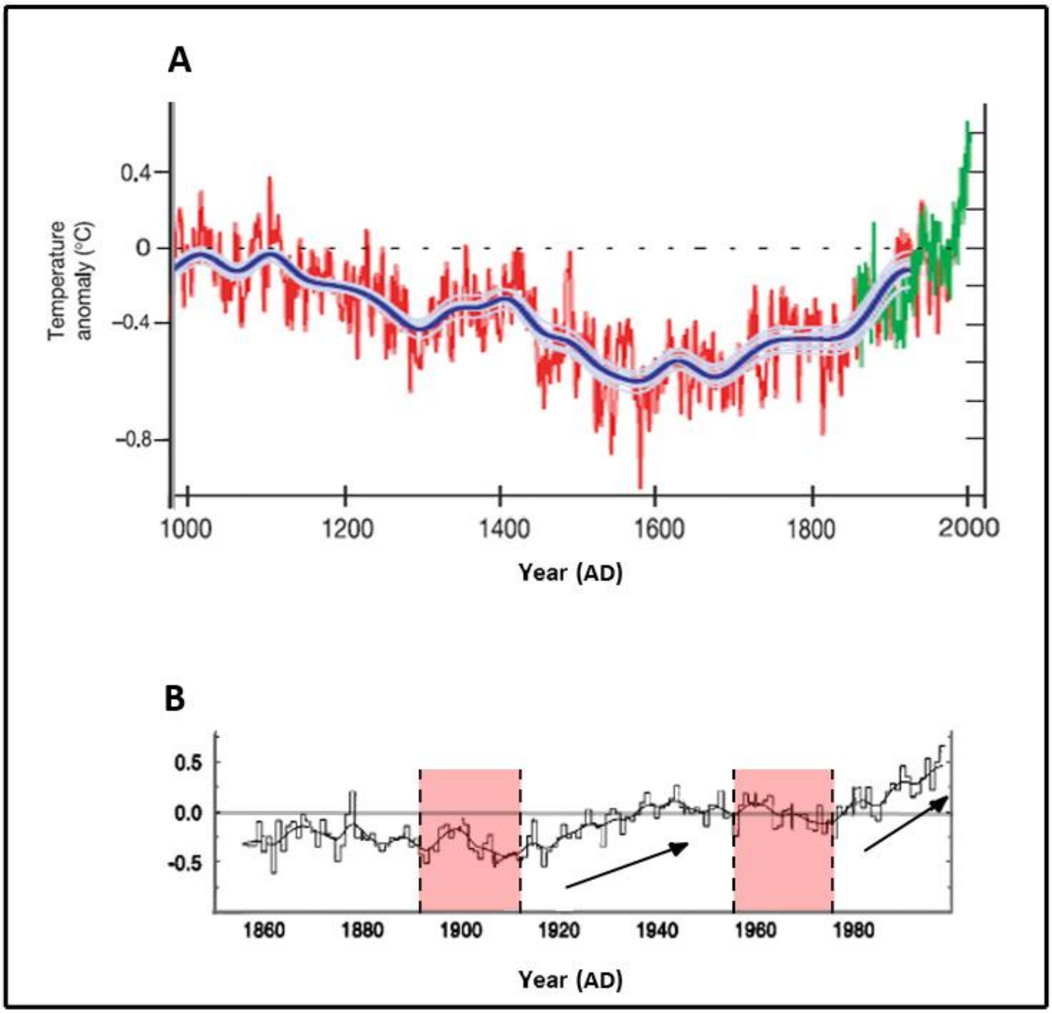


Figure 2.5: (A) annual temperature curve for the Northern Hemisphere, green curve represents continuous instrumental data (Moberg *et al.*, 2005). (B) Annual temperature curve for AD 1860 - 1990, temperatures expressed as deviation from the AD 1951 - 1970 mean (adapted from Jones and Bradley, 1995).

The historical climatic data presented were generally created using historical documentary evidence to provide precipitation assessments and instrumental data were used to create temperature curves. Historical data does provide some insight into the variability of LIA climate and the effects and consequences of these climatic shifts on populations throughout historical time. As historical records were generally collected in lowland areas they are somewhat disconnected from mountain environments and the variations experienced in those areas. Reconstructions of precipitation using documentary data (see Rodrigo and Barriendos, 2008) could be seen as inherently unreliable as the validity of historical documents cannot be tested. Comparison of documentary or instrumental datasets to multi-proxy reconstructions (e.g. Moberg *et al.*, 2005) do however, provide a means of comparing and contrasting different data sets. The datasets produced within this section (see Figures 2.4, 2.5) offer some insight into the variability of the LIA, but other proxy records such as glacial reconstructions are useful for identifying climatic variations during the LIA period.

2.3 EVIDENCE OF LIA AND SUBSEQUENT CLIMATIC FLUCTUATIONS FROM GLACIAL GEOMORPHOLOGICAL RECORDS

Research into glacial activity across Europe has covered many aspects of the subject from the study of geomorphological remnants of Pleistocene glaciations to the current monitoring of glacial retreat (e.g. Bavec *et al.,* 2004; Hughes, 2007, 2008; Hughes *et al.*, 2006, Braithwaite and Raper, 2007; Delmas *et al.*, 2008). The majority of research has focused on the Alps and Pyrenees in Central and Southern Europe and in the Scandinavian mountains, probably because this is where the largest and most numerous glaciers are found. These studies have been useful in describing past trends in glacial expansion and retreat, but has generally overlooked the activity of small high mountain glaciers in the Balkans and Eastern Mediterranean (Kuhn, 1984; Hughes, 2007; González Trueba *et al.*, 2008). Research by Vincent *et al.* (2002) suggested that while contemporary glaciers in areas of the Alps are controlled by mean summer temperatures, in the past during periods such as the LIA they were more likely to be driven by regimes of winter precipitation.

2.3.1 Evidence from the Alps and the Western and central Mediterranean mountains

A majority of the literature discussing LIA glacial fluctuations is based on studies in mountains such as the European Alps and the Pyrenees. In other areas, such as the Cantabrian Mountains and Italian Apennines, fewer datasets are available (Figure 2.1). Both of these mountain ranges exist in an environment marginal for glaciation and where the small high mountain glaciers that do survive are highly responsive to climatic change (González Trueba *et al.*, 2008; Hughes, 2009b). Thus, areas such as these are ideal for assessing glacier response to climate fluctuations such as the LIA (e.g. Grudd, 1989; Kuhn, 1993; Kuhn *et al.*, 1997). Isolated glacial activity within these areas relies heavily upon local topoclimatic controls, such as glacier orientation (shading), avalanching and windblown snow (drifting) accumulation for their current survival (González Trueba, 2003/2004; González Trueba *et al.*, 2008; Milivojević *et al.*, 2008). Glaciers with a northerly orientation are often strongly shaded with reduced receipt of solar radiation, compared with other slope orientations. Ideal conditions for preservation of accumulated snow are created at northerly orientations, as daily ablation rates of strongly shaded areas are up to half that of those fully exposed (Kuhn, 1984; Kruss and Hastenrath, 1987; Chueca and Julian, 2004). It is these local conditions that have allowed glaciers within the Mediterranean region to exist throughout the LIA, and in many cases to the present day.

The European Alps

The LIA glacial history of the European Alps provides some of the most detailed glacial records, with historical documentation recording the areas of land overridden by ice during glacial expansion (Grove, 1988). Generally, most European glaciers saw their maximum LIA glacial extent ***c.*** AD 1850 – 1860. Glacial advances in the region have been frequent since 3,000 yrs BP, but glaciers were most extensive during the LIA (Grove, 1988; Ivy-Ochs *et al.*, 2009). The Mont Blanc Massif has particularly abundant historical documentation providing evidence for varying glacial extent prior to the LIA_{GM}.

Throughout the 1600s glacial extension throughout the Alps was capricious, with evidence of advances at ***c.*** AD 1640 and ***c.*** AD 1680, and withdrawals ***c.*** AD 1644 and ***c.*** AD 1690. This erratic behaviour continued throughout the 1700s, before large glacial advances took place between AD 1770 - 1780, AD 1818 - 1820 and AD 1850. The first of these advances coincides with LIA glacial maximum in the glaciers of the Italian Apennines following glacial retreat across the Alps (Stotter, 1846 cited in Grove, 1988) and the latter advance coincides with the LIA_{GM} of the Eastern Alps (Grove, 2001a; Holzhauser *et al.*, 2005; Joerin *et al.*, 2006; Ivy-Ochs *et al.*, 2009). During the 1900s geomorphological mapping of the European Alps was completed (Grove, 1988) with a number of expansions and retreats similar to those around the Mediterranean region identified in the area (Hoelzle *et al.*, 2003). By the late 1950s many glaciers in the Swiss Alps had begun to register a net mass balance loss (Ohmura, 2004), but this trend was reversed in the late 1970s - 1980s when glaciers in the Swiss Alps began to extend. Since the 1980s glacial recession has occurred throughout the European Alps (Hoelzle *et al.*, 2003; Dyurgerov and Meier, 2007).

The Pyrenees

The Pyrenees stretch from the Bay of Biscay to the Mediterranean Sea between 42^{o} 00' and 43^{o} 00' N (Figure 2.1). Of all the mountain ranges in the Mediterranean, the Pyrenees have the most extensive record of recent glacial activity (e.g. Alonso *et al.*, 1983 cited in González Trueba *et al.*, 2008; Grove, 1988; Serrano and Martínez de Pisón, 1994; Grove and Gellatly, 1995; Pallás *et al.,* 2006). The two largest Massifs in the Pyrenees, the Vignemale and Maladeta, have numerous LIA glacial remnants with 25 cirques and some contemporary isolated glaciers, the largest and most well preserved being in the Maladeta Massif (Cía *et al.*, 2005; González Trueba *et al.*, 2008). Permanent snowlines in the Pyrenees currently range from 2900 m a.s.l. in the west to 3100 m a.s.l. in the east (Grove and Gellatly, 1995) with geomorphological relics of LIA glacier advances found as low as 2200 m a.s.l. (Kuhn, 1984; González Trueba *et al.*, 2008).

Historical sketches, documents and photographs provide the earliest record of glacial activity and permanent snow in the area (Carbonnières, 1789 cited in Gonzalez Trueba *et al.*, 2008; Pasumot, 1797 cited in Grove 1988; Haeberli and Hoelzle, 1993) The record of LIA glacial history begins with glacial expansion in the 1700s (Russell, 1908; Alonso *et al.*, 1983 cited González Trueba *et al.*, 2008; Grove and Gellatly, 1995) with evidence of a 3 km glacial extension at Glacier d'Ossoue around this time (Russell, 1908; Grove and Gellatly, 1995). Carbonnières' (1789) contemporary accounts indicate glaciers in advanced positions ***c.*** mid-1700s with protection from late-lying snow, but by the end of the 1700s a sporadic retreat had begun with some minor glacier re-advances (Carbonnières, 1789 cited in González Trueba *et al.*, 2008; Pasumot, 1797 cited in Grove 1988). Glacial behaviour by the 1800s varied across the mountain range, with lichenometric dating of push moraines suggesting short periods of glacial expansion around the 1820s (Julián and Chueca, 1998; Cía *et al.*, 2005). The period between AD 1850 - 1870 is less easily defined with data indicating glacial expansion and retreat throughout this time (e.g. Trutat, 1875; Michelier, 1887; Bonaparte, 1891; Eydoux and Maury, 1907; Russell, 1908; Schrader, 1936: all cited Grove, 1988; Grove and Gellatly, 1995). The late 1800s were then characterised by predominantly stationary glaciers (Schrader, 1894; Vallot, 1887; Beraldi, 1898: all cited Grove, 1988). Geomorphological mapping of the Glacier du Pays Bache has since suggested that the re-advances at this time were limited, while reconstruction of equilibrium line altitudes (ELA) suggested that LIA temperatures were 0.7^{o}C to 0.9^{o}C cooler than present (González Trueba *et al.*, 2008). Glacial retreat continued into the 1900s across the Mediterranean region, but unlike ranges such as the Cantabrian Mountains, glaciers of the Pyrenees did not experience complete deglaciation (Grove and Gellatly, 1995; González Trueba, 2003/2004, 2005; González Trueba *et al.*, 2008). Twentieth century Pyrenean glacial retreat appeared to be interrupted twice, during the 1920s - 1930s and 1960s - late 1970s. These periods were characterised by below average temperature and above average winter precipitation (Grove, 1988, 2001a; see section 2.3.3), causing some glacier systems to transform into snowfields e.g. Frondella and Brècha Latour. Data from the Spanish Pyrenees for the World Glacier Survey (Serrat and Ventura, 1993) and Ministerioa de Obras Publicas y Urbanismo (M.P.O.U.; Martinez de Pisón and Parra, 1988), suggest that many glaciers remained stationary and even expanded during the late 1970s and early 1980s as a result of positive mass balances. Since the 1980s a steady glacial retreat across the Pyrenees has resulted from summer warming and reduced winter precipitation. The surviving 22 Pyrenean glaciers have endured due to favourable local conditions with predominantly north-south orientations and input of snow from avalanching (Kruss and Hastenrath, 1987; Kuhn, 1984; Chueca and Julian, 2004).

The Maritime Alps and Apennine Mountains

Located in North-West Italy (Clapier and Maledia-Gelas Massifs) and South-East France (Les Grandes Rousses range), between 44^{o} 8' - 44^{o} 5' N, data regarding the LIA glacial history in the Maritime Alps is sparse. The available data indicates glaciers in the French area of the range reached their maximum extent ***c.*** AD 1850 prior to Pyrenean retreat of the 1860s and 1870s (Michelier, 1887; Grove, 1988; González Trueba *et al.*, 2008). Recent retreat of small glaciers at the Les Grandes Rousses range followed a period of stability in the 1950s to 1980s (Vivien, 1975; Kruss and Hastenrath, 1987; Grove, 1988) as seen in the previously mentioned ranges. Favourable local conditions then enabled the survival of six glaciers into the 1990s (Kuhn, 1984).

The recent glacial history of the Apennine Mountains in Italy is comparable to that of the Cantabrian Mountains of Spain (see below), as are the summit heights at 2500 - 2900 m a.s.l. The Ghiacciaio del Calderone glacier, Gran Sasso Massif, Central Italy, was identified as the last remaining glacier in the Apennines and most southerly glacier in the Mediterranean region at 42^{o} 28' 15" N (Grove, 1988; Gellatly *et al.*, 1994; D'Orefice *et al.*, 2000; Giraudi, 2005). Modelling of this glacier implied a maximum 3.0^{o}C LIA decrease in Apennine winter temperatures, causing glacial extension to 2500 - 2700 m a.s.l. (Giraudi, 2005). D'Orefice *et al.* (2000) suggested progressive ice loss at Ghiacciaio del Calderone glacier of around 30,000 m^2 in a 100 year period, AD 1880 - 1980 and successive periods of retreat and expansion, some coinciding with those elsewhere. Within this general retreat glacial expansion has been recorded at AD 1790 - 1819 (Apennine LIA$_{GM}$) and AD

1855 - 1856. By the late 1800s a retreat occurred coinciding with cool Alpine temperatures (Briffa *et al.*, 1988; Schweingruber *et al.*, 1988, 1991; Luterbacher *et al.*, 2004). A further Apennine expansion occurred AD 1930 - 1950, with subsequent retreat interrupted by a minor expansion in the 1970s (Wood, 1988; D'Orefice *et al.*, 2000).

Cantabrian Mountains and Sierra Nevada

The Cantabrian Mountains of North-West Spain (Figure 2.1) are probably one of the least studied areas in the Mediterranean, second only to the Balkan area. Evidence of LIA glaciation has only been reported for the Picos de Europa Massif, although this is based on retrospective field observations with little historical documentation available (e.g. Saint Saud, 1893 and Penck, 1897 cited in González Trueba *et al.*, 2008). Studies in the 1990s were the first to reconstruct LIA glacial advance, attaining that glacial processes were confined to northerly orientated niches beneath high summits (González-Suárez and Alonso, 1994, 1996; Alonso and González-Suárez, 1998 see González Trueba *et al.*, 2008), but no glaciers currently survive in the area (Clark, 1981).

The brief LIA glaciation of Picos de Europa was characterised by six north facing glaciers sustained by snow drift accumulation due to dominant westerly to northerly airflow (González Trueba *et al.*, 2008). The topoclimatic setting of these glaciers was strongly reliant upon snow drift and avalanching, similar to modern Pyrenean glaciers (Kruss and Hastenrath, 1987; Kuhn, 1984; Chueca and Julian, 2004). Between AD 1550 - 1750 the mean equilibrium line altitude (MELA) in the Picos de Europa was situated at ***c.*** 2600 m a.s.l., (González Trueba, 2003/2004, 2005; González Trueba *et al.*, 2008), but this has risen continuously to 2750 - 2850 m a.s.l. (González Trueba, 2003/2004). Using the limited historical data available it can be assumed that at ***c.*** AD 1850 glaciers retreated and by the 1900s glacial activity had ceased following substantial loss of surface, length and volume, as noted by commentators of the day (Pidal and Zabala, 1918 cited in González-Trueba *et al.*, 2008; Delgado Úbeda *et al.*, 1932 cited in Serrano and González-Trueba, 2005). This limited data corresponds to variations seen in the Pyrenees and the de-glaciation of the area was the result of a rising equilibrium line altitude to >2800 m a.s.l. (above the highest peak in the area, Torre de Cerrado at ~2600 m) due to an increase in air temperatures of around 0.9^{o}C (González Trueba, 2003/2004, 2005; González Trueba *et al.*, 2008).

The Corral del Veleta glacier (37^{o}N, 3^{o}W) in the Sierra Nevada represented the LIA 'southernmost glacier on the European continent' (Messerli, 1980; Schulte, 2002; González-Trueba *et al.*, 2008), and was situated in a north-east facing cirque below the Veleta peak (3398 m a.s.l.). The LIA_{GM} at the site is thought to have occurred at ***c.*** AD 1876 (Schulte, 2002) with continual glacial recession in the area throughout the 1900s with only minor periods of equilibrium/re-advance slowing the rate of retreat. By the 1990s only ice patches remained and these have since disappeared leaving glacial remnants in the form of debris-covered ice bodies and permafrost (Gómez *et al.*, 2001; Tanarro *et al.*, 2001).

Overall, the pattern of recession and advance mirrors that of other Western Mediterranean glaciers. Although, only the glaciers of the Cantabrian Mountains have become extinct in a similar way to the Corral del Veleta as other small high mountain glaciers still survive in areas such as the Pyrenees (e.g. Serrano and Martínez de Pisón, 1994; Grove and Gellatly, 1995; Cía *et al.*, 2005; Pallás *et al.,* 2006).

2.3.2 Evidence from the Balkans and South Eastern Mediterranean

Literature discussing the evidence of LIA activity in eastern areas of the Mediterranean is rare, with greater emphasis placed on research into Pleistocene glaciations (Hughes *et al.*, 2006; Milivojević *et al.,* 2008). Recent attempts have been made to discuss LIA glacial histories in the Eastern Mediterranean (Gabrovec, 1998; Hughes, 2007, 2008) with possibly the most southerly glacierets in the Mediterranean region being recorded in the Pirin Mountains, Bulgaria (Grunewald *et al.*, 1999, 2006; Hughes, 2007). Overall, the LIA glacial record is sparse from eastern areas of the Mediterranean, but some data is available within the Durmitor Massif, Montenegro.

The most recent monitoring of glacial recession in the Balkans/Eastern Mediterranean has been completed in Slovenia. Pavšek (2004, 2007) discussed the Skuta Glacier and Gabrovec (1998) produced data monitoring the Triglav Glacier (Figure 2.1) over a thirteen year period (AD 1986 - 1998). Monitoring at the Triglav site has highlighted its steady retreat and an absence of perennial snow cover during the period of research. Prior to Gabrovec's monitoring of the glacier there is some reference to past glacial extent. Sifrer (1963 cited in Hughes, 2010) stated that during the 1800s the glacier covered an area of 40 ha, and Gabrovec states, "Its surface area [Triglav Glacier], which was fifteen hectares in 1946, had shrunk to less than three hectares by 1998" (Gabrovec, 1998 p. 90). These observations give some limited insight into the history of glacial fluctuations at the Triglav glacier during the latter stages of the LIA. Glacial retreat in Slovenia has continued at both Triglav and Skuta glaciers since the 1980s following a similar trend to the glaciers of the Durmitor Massif, Montenegro (Section 2.3.2; e.g. Grove, 1988; D'Orefice *et al.*, 2000; Pavšek, 2007; González Trueba *et al.*, 2008).

Montenegro

The LIA glacial history in Montenegro is similarly sparse to those of the Maritime Alps and Cantabrian Mountains. Progress is being made in the Durmitor Massif (Figure 2.1) focusing on the Debeli Namet Glacier (42^{o} to 44^{o} N), currently one of the largest glaciers in the Northern Hemisphere. Debeli Namet is situated at a relatively low altitude of 2050 - 2300 m a.s.l., considering permanent snowlines in the Eastern Pyrenees have been estimated at around 3100 m a.s.l. (Grove and Gellatly, 1995). Lichenometric results produced by Hughes (2007) suggested that moraines close to the glacier snout originated from late 1800s glacial expansion, corresponding to cooler and wetter summers identified in the European Alps AD 1875 - 1925 (Hughes, 2007). Furthermore, the lichenometry data led to the suggestion that the maximum LIA glacial extent and coolest period in the Montenegrin Mountains occurred at ***c.*** AD 1878 - 1890, compared to ***c.*** AD 1850 in Europe. Little is known regarding the state and behaviour of the Debeli Namet glacier during the 20th century. Most recently, the activity of Debeli Namet has been monitored by Hughes between AD 2003 and AD 2007, recording glacial retreat of 5.2 m following one of the hottest summers in 2007. A new frontal moraine was formed between AD 2004 and AD 2006 following a reduction in mean summer temperatures and increase in winter precipitation (Hughes, 2008). This shows the sensitivity and rapid response of small glaciers like the Debeli Namet to climatic stimuli (cf. Bahr *et al.*, 1998 cited in Hughes, 2008). Hughes (2010) described karstic topographic forms including dolines and suggested that these were favourable locations for the build up of snow that could result in permanent ice. Today permanent snow fields and ice patches still survive in some dolines (Figure 2.6) and also in caves (Ledina Pecina ice cave; Hughes, 2010). The identification of these landforms alongside the presence of current glaciers at low altitudes and low latitude lead to the suggestion by Hughes (2010) that karstic terrain is important in glacier inception and later development.

Figure 2.6: The Šljeme ice doline at Durmitor Massif (taken from Hughes, 2010).

The Prokletije range (excluding Kosovo)

Unlike the Durmitor massif of Montenegro, no research has yet been undertaken regards recent and LIA glacial variations in the Prokletije Mountains (Figure 2.1). The Northern

Albania Prokletije range has been investigated by Milivojević *et al.* (2008), but research concentrated on Pleistocene glacial extents and geomorphological structures. Although they did note current active geomorphological processes around the Maja e Jezerces summit (2694 m a.s.l.) in Albania. Hughes alluded to the presence of LIA glacial remnants in this area; although his study was concerned with the extent of modern (21^{st} century) ice masses (Hughes, 2009a). Very little is known about the extent and timing of these former glaciers and whether they correspond with LIA glaciers nearby in Montenegro and elsewhere in the Mediterranean mountains.

2.3.3 Synthesis of glacial geomorphological evidence

Glacier research focusing on the LIA period has taken place across the Mediterranean region, with the most complete histories available for the Alps and Pyrenees. Data from around the Western Mediterranean and Europe suggests that during the late 1700s glaciers were at their LIA_{GM} in the French section of the Mont Blanc Massif and the Apennine Mountains (Italy). Glaciers in the Pyrenees and the remaining European Alps were at an extent similar to that of the LIA_{GM} of the area (AD 1845 - 1855). However, Eastern Mediterranean sites such as the Debeli Namet glacier in Montenegro seem to have reached maximum glacial extent around a century after the LIA_{GM} of the Apennines and Mont Blanc Massif and around four decades later than the LIA_{GM} of the European Alps and the Western Mediterranean mountains (Figure 2.6). This disparity may be due to small glaciers responding more sensitively to differing local factors (e.g. orientation, volume of snow accumulation via avalanching and snow drift) at the Western Mediterranean and Balkan sites. There may also be regional variability in climate, as seen in the differing dendrochronological records discussed previously (Kruss and Hastenrath, 1987; Grove, 1988; Kuhn, 1984).

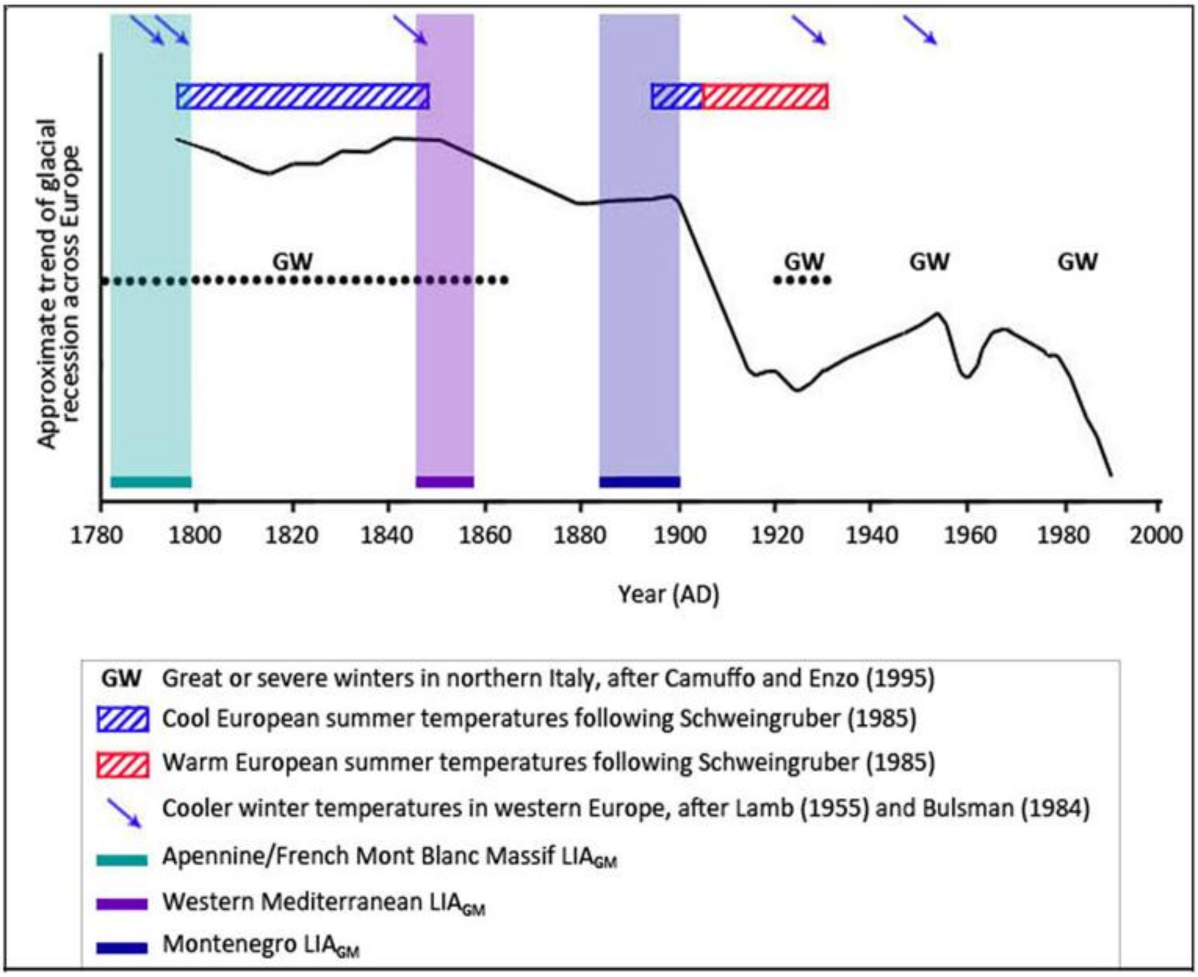


Figure 2.7: Combined illustrated synthesis of selected glacial and climate records of the Mediterranean and Southern Europe (adapted from D'Orefice *et al.*, 2000 and including various sources).

At the time of writing this chapter, there was not enough research discussing the dynamics of Eastern Mediterranean glaciers to establish firm conclusions for differences between the various sites. Although the timing of the LIA_{GM} differs across Europe and the Mediterranean (Figure 2.7), the periods of glacial retreat and expansion can be summarised as follows: glacial expansion during the late 1700s (Western Mediterranean LIA_{GM}), further expansion between AD 1820 – 1850, glacial retreat AD 1860 – 1870, further expansion or halt to recession during AD 1880 - 1900 (Montenegrin LIA_{GM}) and overall recession AD 1900 onwards. The latter period included minor fluctuations where recession paused and expansion occurred at approximately: AD 1910; 1920s; 1930s to

1950s; late 1960s to late 1970s. A final period of overall recession occurred from AD 1980 to present. The record shown in Figure 2.7 uses the curve of glacial recession estimated by D'Orefice *et al.* (2000) for the European region. Overlain on to this diagram are dendroclimatological and historical documentary climate data, along with the LIA glacial maxima for both the Western and Eastern Mediterranean.

Western Mediterranean		Eastern Mediterranean/Balkans	
expansion	retreat	expansion	retreat
**1780 - 1810			
	1810 - 1820	?	
*1845 - 1856			
	1860 - 1870		
		*1878 - 1890	
1880 - 1900	1900		1900
1910	↓		↓
1920		?	
1930s to 1950s		↓	
late 1960s to 1970s			
	mid 1980 s		?
		(2004 - 2006)	

Table 2.3: Periods of glacial retreat and expansion in Western and Eastern Mediterranean regions (various sources). * Indicates period of possible LIA_{GM}, ** relates to Apennine glacial and French Mont Blanc Massif LIA_{GM} and bracketed dates relate to the Debeli Namet glacier in Montenegro.

Documentary evidence from around Europe throughout the LIA corresponds well to the overall trends seen during European glacial recession, with cooler or more severe winter temperatures corresponding to periods of glacial equilibrium or expansion. Table 2.3 shows the differences between the timings of general glacial retreats and extensions in the Western Mediterranean and the reconstructed glacial history of the Eastern Mediterranean using data from the Debeli Namet glacier in Montenegro and the Triglav and Skuta glaciers in Slovenia. Within the Apennine record, the LIA_{GM} occurred at the end of the 1700s, and severe winters were experienced throughout Northern Italy (Camuffo and Enzo, 1995). A subsequent glacial expansion commenced in AD 1820 (Figure 2.7) which may have been influenced by the earlier AD 1815 winter which is recorded as a major climatic event in global climate (Popa and Kern, 2009). As glacial retreat across Europe began severe winter temperatures were documented, with Schweingruber (1985) reconstructing cooler summer periods from ***c.*** AD 1800 - 1850 (Figure 2.7). By the mid 1800s glaciers across Europe had began to expand reaching LIA_{GM} around AD 1850 and then from AD 1860s glacial recession began throughout Europe and the Western Mediterranean, as seen in Figure 2.7. Data prior to the LIA_{GM} in the Eastern Mediterranean is unavailable as the glacial expansion in the region would have overridden any earlier geomorphological structures that enable the mapping of past extents.

The LIA_{GM} in the Eastern Mediterranean (AD 1878 - 1890) corresponds to a pronounced cold period identified within dendroclimatological records for the Calimani Mountains (Romania) and a period of slight glacial expansion within the European record (Figure 2.7). However, the Eastern Mediterranean LIA_{GM} lagged that of the Western Mediterranean (AD 1780 - 1810) by a century. Dendroclimatological records from the European network suggest that during this time warm conditions were experienced (Table 2.2), yet glaciers were expanding in the region (Table 2.3). By AD 1892 a cold period had commenced in the Western Mediterranean (see Table 2.2) with glaciers in the area slowly expanding (Table 2.3; Figure 2.7). Records for the Eastern Mediterranean are fragmented after the 1900s, with only a general trend of retreat discernable from documentary records. Western Mediterranean data mirrors that of European glacial recession, with rapid glacial retreat punctuated by periods of glacial equilibrium or expansion. Figure 2.7 identifies periods of warmer reconstructed summer temperatures and glacial retreat, whilst periods of severe winter temperatures coincide with glacial expansion. Unlike Western Mediterranean glaciers, a recent glacial expansion was recorded by Hughes (2008) between AD 2004 – 2006, the result of below average summer temperatures and increased winter precipitation. This expansion has not been recorded elsewhere in the Mediterranean region and is likely to be the product of favourable local climate conditions (previously described) between AD 2004 - 2006 (Hughes, 2008). This highlights the capricious behaviour of local mountain glaciers in the Mediterranean in response to short-term climate change, and this makes it difficult to compare very short-term (inter-annual) glacier fluctuations over wide areas. Nevertheless, as shown in this discussion, more general similarities in decadal variability in glacier behaviour can be indentified between sites across the Mediterranean mountains (Figure 2.7; Table 2.3).

2.4 PALAEOENVIRONMENTAL RECONSTRUCTION

Palaeoenvironmental work has been undertaken for many years, but shifted towards a quantitative, analytical science more recently. This has allowed the impacts of environmental, climatic and anthropogenic change to be detected within different systems. As a result improvements in the statistical analyses, modern calibration data sets and methods used during such studies has occurred and placed emphasis on quantitative multi-proxy reconstructions of the palaeoenvironment (Smol, 2002; Birks and Birks, 2006). Throughout the world there have been a variety of palaeoenvironmental studies focusing on the impact and identification of late Holocene climatic variations such as the LIA. The impact of both climatic changes and influence of human activity upon the environment has been distinguished (e.g. Le Roy Ladurie, 1971 cited in Grove, 1988; Parry and Carter, 1985; Carrión and Navarro, 2002; Driese *et al.*, 2004; Chen *et al.*, 2005; Cohen *et al.*, 2005; Hotchkiss *et al.*, 2007; Rabatel *et al.*, 2008; Rebolledo *et al.*, 2008; Axford *et al.*, 2009; Fey *et al.*, 2009). Within this review it is aimed to identify the broad trends seen during the LIA and the current limitations of the palaeoenvironmental records available.

2.4.1 Global LIA palaeoenvironmental records

Across Europe, research into climatic variations has been undertaken with the use of multi-proxy research, and the effects of these climatic changes on the environment have been explored in some areas. For North-Western Europe peat bogs have provided multi-proxy data used to create a record of wetness, identifying the LIA climatic deterioration by intermittent periods of increased wetness. Barber *et al.* (2000), Mauquoy *et al.* (2002) and Lamentowicz *et al.* (2008) found evidence of water pools and wet shifts in plant macrofossil and testate amoebae records and suggested that periods of reduced peat accumulation during the LIA were the result of suppressed temperatures. European lacustrine records have provided similar evidence of wet shifts, recorded using organic carbon as a proxy for palaeoproductivity. This summarises the response of aquatic plants to augmented precipitation, which causes influxes of nutrients and periods of enhanced in-wash (Battarbee *et al.*, 2001). Across the French Pre-Alps and Europe a relative rise in lake level was indicated around AD 1394 (e.g. Harrison *et al.*, 1993, 1996; Magny, 2004; Chapron *et al.*, 2007a). Overall, the majority of European work has been concerned with the identification of particular climatic changes during the LIA and the reconstruction of these variations using proxies, as opposed to the impacts of the deterioration on the local environment. However, the information collected does provide insight into the climatic factors that have influenced environments throughout the late Holocene to present.

Research in America provided greater insight into the impacts of LIA cooling upon a forest ecosystem. Campbell and McAndrews (1993) simulated forest succession during and following the LIA and found the dominance of particular forest species changed following the deterioration. An alteration from *Fagus*, to *Quercus* and finally *Pinus* was suggested. It was mentioned that succession was less likely to be registered if a successional species was not well represented within the pollen assemblage, or where frequent prehistoric fire regimes existed - which would destroy contemporary surface layers and trapped pollen. Similar work has been undertaken in the Mediterranean where the ecological transition from deciduous dominating types to evergreen types throughout the Holocene has been reconstructed along the French/Spanish coast (Jalut *et al.*, 2000). Again in the USA, during the onset of the LIA in the Lower Hudson Valley, evidence of climatic cooling and increased precipitation came from the expansion of species preferring cooler climes (*Fagus* and *Pinus*) and moisture tolerant forms such as *Tsuga* (Pedersen *et al.*, 2005). These examples highlight that relatively 'short' climatic changes can have a profound effect on the environment causing transformations of plant communities and ecologies, but that LIA climatic variations seen in Europe were experienced in the USA too.

2.4.2 Reconstructions from the Mediterranean region

There is a rich tradition of palaeoenvironmental research in the Mediterranean region, as the area is renowned for its sensitivity to climatic change alongside the presence and consequent impact of human occupation (Butzer, 2005). Much of this research is focused on landscape dynamics in the late Quaternary and Holocene (e.g. Willis and Bennett, 1994; Allen, 2003; Aguzzi *et al.*, 2007). Longer Quaternary multi-proxy records are available across the Mediterranean from Spain (e.g. González-Sampériz *et al.*, 2009; Martín-Puertas *et al.*, 2009), Italy (e.g. Pini *et al.*, 2009), North Africa (e.g. Cheddadi *et al.*, 1998) and the South Eastern Mediterranean (Tzedakis *et al.*, 2003; Tzedakis *et al.*, 2004a; Bordon *et al.*, 2009). These provide environmental and climatic reconstructions on decadal, centennial and millennial time scales (Rodrigo *et al.*, 2000).

Late Holocene records from Spain have centred on the reconstruction of anthropogenic influences upon the natural landscape (e.g. Valero-Garcés *et al*, 2006; Morellón *et al.*, 2008). The available hydrological and vegetational records show that these variables are more responsive to moisture availability than temperature fluctuations, skewing the climatic changes that can be inferred (Pons and Reille, 1988; Valero-Garcés *et al.*, 2006). Mediterranean sites located in Southern/Central Spain and Tunisia provide the best late Holocene records and go some way to indentify the impacts of climate change, and associated environmental changes during the LIA. Studies at La Mancha Plain in South-Central Spain combined palynological, geochemical and sediment analysis data at a roughly decadal resolution to detect climatic deteriorations restricted to the beginning of the LIA (García *et al.*, 2007). Within the palynological record the ratio of evergreen *Quercus* to *Pinus* and *Artemisia* were used as palaeotemperature indicators. The onset of the LIA was dated to AD 1400 after increased levels of *Pinus* and *Artemisia* were identified, as both are indicative of cooler temperatures. Geochemical and sediment records complement these data as higher fluctuating values of organic material and evidence of eutrophication episodes were used to identify climatic instability associated with the LIA (Lamb, 1977; García *et al.*, 2007). Similar data have been reported for Southern Spanish sites at Laguna Zoñar (Zoñar Lake) and Laguna de Archidona, where a combination of sedimentological features and diatom assemblages were used to identify wet and dry shifts. During the mid 1600s dry shifts were indicated at Laguna de Archidona and Laguna Zoñar by an increased lacustrine component and evidence of weak soil processes (Luque *et al.*, 2004). By the late 1600s a return to wetter conditions occurred, coinciding with the LMM across North-West Europe (Valero-Garcés *et al.*, 2006), but unlike European records, these wetter conditions persisted in Spain until around AD 1850. This caused lake levels to rise at both Laguna de Archidona and Laguna Zoñar, forming shallow lakes in the Doñana National Park, Southern Spain (Sousa and García-Murillo, 2003). Spanish records point to a reduction in clastic and fluvial inputs to lakes, supported by diatom records at Laguna Zoñar. These suggested that as the climate became drier (post -1850), greater volumes of organic material were deposited causing finely laminated facies to appear in lake sequences. Increasing aridity during the late 1800s and 1900s resulted in many of the shallow lakes at Doñana National Park disappearing (Sousa and García-Murillo, 2003). Climatic signals at Laguna Zoñar and Laguna de Archidona are limited by anthropogenic influences during recent centuries as humans manipulated the natural landscape. Cultivation indicators, such as Olive (*Olea*), were used to identify the expansion of cultivation and thus presence of human occupation and activity at the sites (Roberts *et al.*, 2004; Riera *et al.*, 2004; Valero-Garcés *et al.*, 2006). Continued anthropogenic pressures since the 1950s across Spain have led to lowered lake levels, obscuring the climatic signal, but providing good evidence of anthropogenic influences in the region (Valero-Garcés *et al.*, 2006).

Tunisian records are similarly influenced by anthropogenic activity which can obscure climate signals and make differentiation between climatic and anthropogenic forcing factors within the palaeo-record difficult. To try and alleviate some of these difficulties, climatic reconstructions at Sebkha Mhabeul were developed with the use of palynological and sedimentary data. The record demonstrated a slight difference to previous Mediterranean data, with heavy rainfall inferred between AD 1420 – 1645. During this period major flood events occurred simultaneously across Tunisia at Sebkha Mhabeul and in the mid-Medjerda basin (Faust *et al.*, 2004; Marquer *et al.*, 2008). Climatic trends thereafter paralleled those of Spain with drier conditions dominating towards the end of the LIA (Ladkhar *et al.*, 2006; Marquer *et al.*, 2008). The sedimentary records collected in Tunisia are limited by a resolution of around 40 years, which has resulted in a limited reconstruction of the LIA climate and more recent climatic amelioration. Studies within Tunisia focused on the use of palynological records to decipher anthropogenic influences in the region with pollen types indicative of cultivation (*Olea*; *Cerealia* type) visible since Roman times (Marquer *et al.*, 2008). As with previous sites certain pollen types indicative of cooler wetter periods (*Artemisia;* Lovvorn *et al.*, 2001) were used to identify climatic variations through time.

2.4.3 Reconstructions from the Balkans, South Eastern Mediterranean and Eastern Europe

Recent palynological studies using lacustrine sediments in the South Eastern Mediterranean have generally focused upon vegetation refugia, late-glacial environmental change and Holocene variations at a coarse resolution (Tzedakis, 1993; Tzedakis *et al.*, 1997, 2002; Carrión and Navarro, 2002; Goñi *et al.*, 2002; Tonkov *et al.*, 2006; Lippi *et al.*, 2007; Bordon *et al.,* 2009). Research undertaken on long sediment cores from lacustrine sites in the Mediterranean, used long mean sampling intervals focusing on broad Quaternary changes (Wijmstra, 1969; Wijmstra and Smit, 1976; Follieri *et al.*, 1988; Tzedakis, 1999). However, the Quaternary records available do imply that the Balkan/South Eastern Mediterranean region is an area of vegetational sensitivity, where refugia existed throughout the Quaternary (e.g. Médail and Diadema, 2006). Palynological records have provided estimates of temperature and precipitation changes at sites such as Lake Maliq in Albania (Bordon *et al.*, 2009), whilst Quaternary vegetation refugia were identified during research in Bulgaria, at Lake Skedmo Rilsko, Lake Kremensko and Lake Besbog (Bozilova and Tonkov, 2000; Atanassova and Stefanova, 2003; Stefanova *et al.*, 2006a,b). Pollen types previously used to infer climate and land-use changes are summarised in Table 2.4. Although high pollen producers such as *Secale cereale* and *Humulus lupulus* should be analysed with care, to ensure the signal produced is not over interpreted as they are likely to be overrepresented within a pollen assemblage. Other anemophilous cereal types, such *Hordeum* are more likely to be locally derived (Sergerstrom, 1991). Further data have been gleaned by ostracoda analysis, using species identification to reconstruct the environment and quantity of oxygen isotopes ($\delta^{18}O$) sequestered in their shells (Holmes and Chivas, 2002). The assemblages found at the Ioannina lake basin in Greece showed various lake transformations and lake level changes. Samples were dominated by *Candona* cf. *parvula* suggesting sub-littoral habitats, but those dominated by *IIyocypris decipiens* and *Prionocypris* sp. indicated even shallower water (Galanidou *et al.*, 2000).

The Balkan and South Eastern Mediterranean region does have some records that provide evidence of anthropogenic activity and recent environmental impacts resulting from climatic deteriorations and ameliorations to the present day. Willis and Bennett (1994) described the spread of agriculture across the Balkans, citing indicators such as *Juglans*, cereal types and reductions in woodlands as representative of this anthropogenic activity. Agriculture reached sites at various times with Greece being the first area to register cultivation at 5500 yrs BP and Slovenia the last at 1400 yrs BP (Willis and Bennett, 1994).

In Eastern Europe temperature reconstructions were completed by Johnston *et al.* (2010) using a guano deposit in North Western Romania. This enabled reconstruction of bat inhabitancy at the Măgurici cave. Accumulation rates of guano diminished during cooler periods, possibly as the bats migrated to warmer climates at the beginning of the LIA, but increased during AD 1647 suggesting a respite from a cooler climate and confirming the heterogeneous nature of LIA climate. Across the rest of Europe and the Mediterranean the late 1600s were recorded as a period of increased severity (e.g. Schweingruber, 1985; Briffa *et al.*, 1988; Büntgen *et al.*, 2005), but dendrochronological data from the Calimani Mountains actually suggests that in this area there was no significant temperature decline at the time, and the temperature actually increased between AD 1640 - 1740, peaking to + 2.31^{O}C above the AD 1961 - 1990 mean (Popa and Kern, 2009). By the end of the LIA the rate of bat guano accumulation rose simultaneously with temperature suggesting improved and warmer conditions to harbour a growing bat population (Johnston *et al.*, 2010).

Species/type	Environment indicated	References
Secale cereale; Hordeum-type; *Humulus lupulus;* Asteraceae; *Brassica; Olea; Castanea*	Cultivated fields and thus anthropogenic activity; if the increase in such pollen types coincides with a decline in TAP the trend is indicative of agricultural expansion into mountain foothills	Behre, 1981; van den brink and Janssen, 1985; Sergerstrom, 1991
Rumex; Plantago lanceolata; *Artemisia*; Chenopodiaceae; *Vicia hirsute*; *Rumex*; *Centaurea; Ranunculus; Calluna vulgaris*	Weed types indicative of cultivated, pasture/ meadow and disturbed land; scattered presence within a pollen assemblage indicativeof local grassland maintained by large herbivores	West, 1956; Behre, 1981; Turner and Hodgson, 1981; Buckland and Edwards, 1984; Bakels, 1991; Day, 1991; Allen, 2001; Subally and Quézel, 2002
Artemisia; Chenopodiaceae	Mountain steppe type vegetation	Bottema, 1992; Subally and Quézel, 2002
Melampyrum; *Pteridium aquilinum*	Lightly shaded woodland; grazed forest	Turner and Hodgson, 1981; Behre, 1981; Odgaard, 1994
Quercus; Tilia; Ulmus; Corylus; Alnus	Closed deciduous forest	Bradshaw and Hannon 1992
*Betula; Fraxinus**	Open forest canopy; *opening in the forest canopy	Bradshaw and Hannon 1992
Cyperaceae	Wetland environment; shallow ponds; lake shore	Hibbert 1978; Theocharopoulos *et al.*, 2006
Ericaceae; *Juniperus*	Open areas	van den Brink and Janssen, 1985
Abies; Picea; Pinus; Fraxinus; Sciadopitys; Campanula; Trifolium pignatti;	High-altitude montane indicator	Carni *et al.* 2009; Jimenez Moreno *et al.*, 2010
Cupressaceae; *Myrica; Alnus*	More humid conditions	Jimenez Moreno *et al.*, 2010

Table 2.4: Pollen types as environmental indicators (various sources).

Mediterranean and European palaeoenvironmental reconstructions demonstrated that these regions are sensitive to climatic variations and that changes can be reconstructed using a range of proxies. However, across the region there are few detailed accounts of LIA climatic instability and the impact that these shifts had upon the environment as a whole. Those palaeoenvironmental reconstructions available for the Mediterranean suggest that the beginning of the LIA was marked by generally cooler conditions from the 1300s to late 1800s, with a shift to warmer conditions around the early 1900s. Within this general trend shifts to warmer conditions are registered in the mid 1600s before returning to cooler conditions, at the height of the LIA glacial advance in the Mediterranean (Sousa and García-Murillo, 2003; Faust *et al.*, 2004; Marquer *et al.*, 2008). These trends not only provide information on the changes occurring within the environment, but illustrate how the short abrupt climatic changes registered in dendrochronologies and glacial records are also recorded in palaeoenvironmental records. Overall, the records from the Mediterranean currently provide some insight into the climatic and environmental instability of the LIA, but many concentrate on the influence of anthropogenic activity on changing vegetation, rather than climate. A high resolution record from a Eastern Mediterranean mountain catchment that is climatically sensitive and has limited anthropogenic influences, would offer the opportunity to understand how climatic shifts during the LIA affect both a lake environment, and the wider catchment area.

2.4.4 Fluvial activity and hydroclimatic change during the LIA and subsequent climatic fluctuations

Throughout Europe and the Mediterranean hydrological research focusing on the LIA, suggests this was a period of increased fluvial activity. Geomorphological changes to river systems occurred as a result of hydroclimatic change and were accompanied by a greater frequency of large flood events (Figure 2.8; Arnaud-Fassetta, 2003; Macklin and Woodward, 2009). The majority of research provides data on the effect that the climatic variability during this period caused on hydrological systems. Overall, the changes have varied, with authors describing increased erosion in some areas (Gob *et al.*, 2005) contrasted by periods of sedimentation in others (Macklin *et al.*, 1995; Devillers and Provansal, 2003). The variety of topographic positions and different vegetation in areas has caused heterogeneity in results. Authors such as Macklin *et al.* (1995) describing the transformation to a more humid climate in the Mediterranean having a positive effect on vegetation growth, resulting in decreased sediment supply as vegetation cover expanded. The climatic deterioration of the LIA caused river aggradation in many French rivers, which was defined in a variety of sedimentological (dated using lichenometry) and historical sources as falling temperatures and increased precipitation generated greater amounts of coarse sediment (Bravard, 1989; Gob *et al.*, 2005, 2008). Changes in the supply and generation of sediment caused rivers to widen or become braided. Rivers such as the Rhône evolved throughout the LIA and changes progressed further and further downstream as the LIA took hold (Bravard *et al.*, 1992). Dating of Holocene river terraces in South Eastern Spain attributed the formation of four river terraces to increases in fluvial activity ***c.*** AD 910, AD 1417-1611, early 1900s and AD 1940 (Schulte, 2002). Historical documentation and dated sedimentological evidence from the Loire and upper Ardèche, suggests that the flood events of the area corresponded to those seen across the Mediterranean (Gob *et al.*, 2005), with flood frequency and magnitude increasing throughout the LIA. The record from this area as described by Gob *et al.* (2008) highlighted periods of high activity between AD 1530 – 1700, AD 1750 - 1810 and AD 1840 - 1910, which were interspersed by drier phases. The later phases link well with those of glacial variation for the Mediterranean, with the late 1700s/early 1800s being cited as a period of glacial expansion and maximum in the west of the region, and the late 1800s as the LIA_{GM} in the Eastern Mediterranean and Sierra Nevada (Grove, 1988; D'Orefice *et al.*, 2000; González Trueba, 2003/2004, 2005; Hughes, 2007, 2009b).

The influence of increased river activity is also evident in lacustrine sediments. Revel-Rolland *et al.* (2005) stated that as flood frequency and river activity increased in the Rhône, so too did the volume of clastic material entering Lake Le Bourget, suggesting that increases in river activity can be detected in lacustrine sediments by the volume of coarse/clastic material. However, variations in anthropogenic activity within a drainage basin, vegetational modifications such as deforestation or cultivation will change the volume and provenance of material entering a lake basin (Revel-Rolland *et al.*, 2005). Magny *et al.* (2002) came to a similar conclusion and highlighted the continuous reactions of river systems to climatic changes during the Holocene. Returning to the records of hydrological divergence, studies of the Rhône Delta (a large hydrosystem which drains into the Mediterranean Sea) suggest different factors have controlled the Rhône throughout the Holocene. These controlling factors include climate, sediment supply and the effect of human actions. Fluctuations in the hydrology of the Rhône delta system have been controlled in part by sediment supply, which when it exceeds river transport capacity leads to channel in-filling (Figure 2.8; Bravard *et al.*, 1997; Arnaud-Fassetta, 2003). The LIA at the Rhône delta was characterised by lower river competence and stream powers in the area but higher sediment loads than present. Higher levels of runoff during the LIA were the product of a cooler and more humid climate, and this lead to an increase in the number of flooding events to around 22 months per decade (Arnaud-Fassetta, 2003). A flood dominated regime characterised the periods between AD 1560 - 1600, AD 1651 - 1720 and AD 1751 – 1860; the two latter periods corresponded to glacial expansion/LIA_{GM} and rising lake levels across Europe. Each period of flooding alternated with a period of low flood frequency, drier conditions and a decrease in the river's energy at AD 1600 - 1640, AD 1721 - 1750 and AD 1861 - 1995 (Arnaud-Fassetta and Provansal, 1999; Arnaud-Fassetta, 2003). This is in contrast to the data presented by Gob *et al.* (2008), who suggested that the 1700s and 1800s were periods of increased river activity, highlighting the temporal variations of activity seen in records from across Europe and the Mediterranean. Since the end of the LIA relative degradation of fluvial beds has occurred due to falling levels of runoff and human controls on the Rhône River (Bravard *et al.*, 1992). Flooding events during the 1900s have halved since the height of the LIA, to around 10 months per decade (Arnaud-Fassetta, 2003).

Flood events in the Spanish Mediterranean coastal area have provided some insight into links between climatic oscillations and hydrological response. Documentary evidence from nine sites between Ter/Onyar in the northeast and Seguru in the southeast of Spain, have provided data on the timing of flood events during the LIA. The flooding frequency during the LIA in the Iberian Peninsula increased in a similar fashion to those in the Rhône Delta. Catastrophic flood events have been dated to AD 1572 - 1630, AD 1751 - 1810, and finally AD 1841 - 1870. The timing and intensity of flooding in Spain slightly differs to that of Rhône Delta, with contrasting periods of high and low flood frequency between Spanish and French records, for example AD 1661 - 1690 was recorded as a period of high flood frequency exclusively in France (Figure 2.8; Barriendos-Vallve and Martin-Vide, 1998). Three oscillations of similar intensity and duration, characterised by intense precipitation and increased flood frequency are registered in the Spanish data. Each oscillation corresponds to periods of increased flooding (AD 1570 - 1630; AD 1760 - 1800; AD 1830 - 1870), and when compared to the Grand Rhône River data highlights the importance and resulting effects of local factors to flood events (Figure 2.8). However, the maxima of flooding oscillations in Spain generally correspond to those in the Rhône Delta, suggesting the most intense climatic events had a more widespread flooding effect.

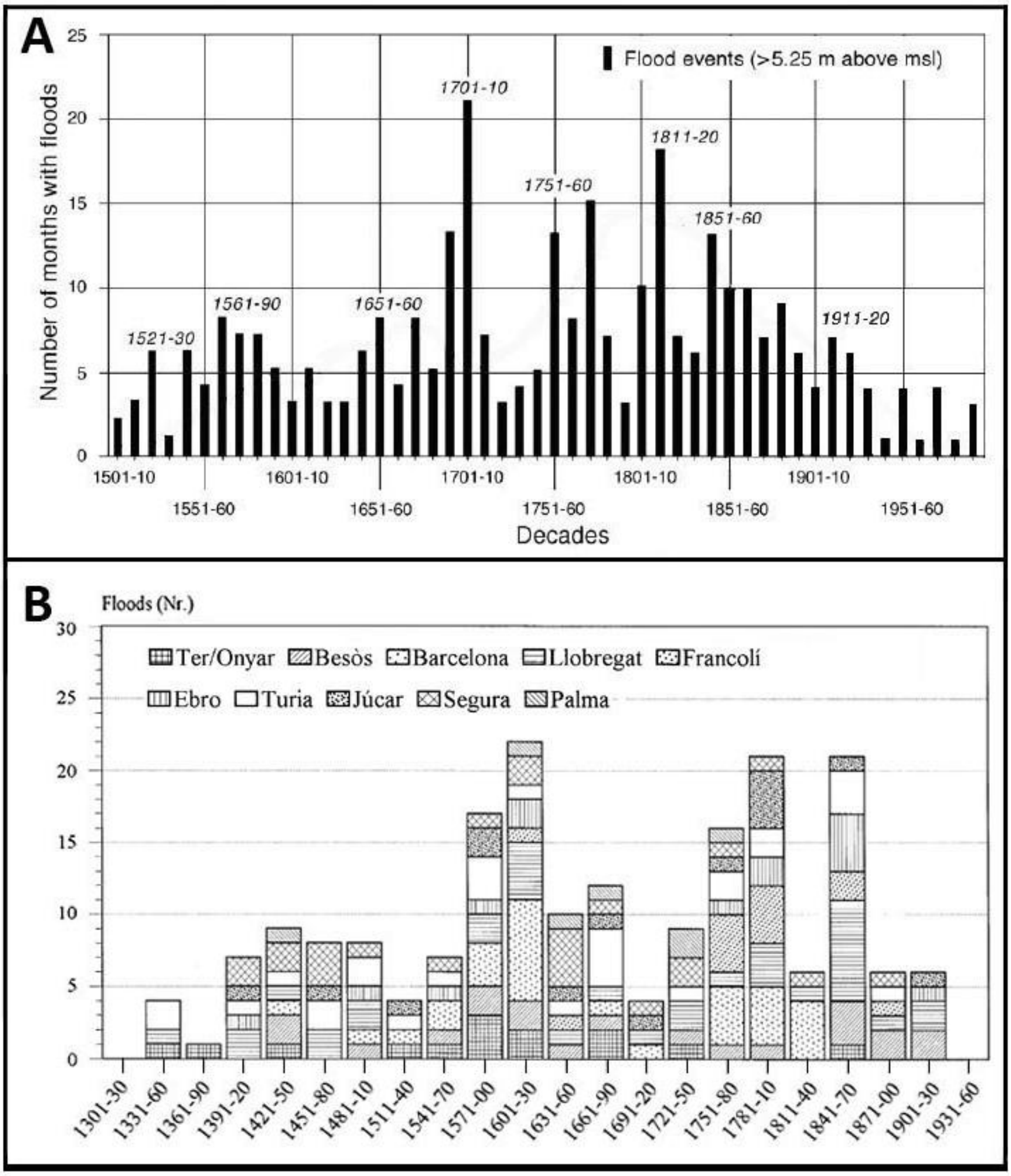


Figure 2.8: (**A**) Frequency of flooding episodes that have affected the Grand Rhone River at Arles since AD 1500 (from Pichard, 1995 cited in Provansal *et al.*, 2003). (**B**) Frequency of 'catastrophic' floods that have affected the Spanish Mediterranean coast, during 30 year periods since AD 1481 (from Barriendos-Vallve and Martin-Vide, 1998).

Maas and Macklin (2002) analysed the impact of climatic change on hydrological systems in the Ardena Gorge, South-West Crete in the Eastern Mediterranean region. The catchment is typical of Mediterranean fluvial systems, with a steep gradient and high sedimentation rates during winter months when intense storm events affect the region. Research at the Ardena Gorge provided data on twelve periods of increased flooding since AD 1850 which correspond to periods of intense precipitation caused by a declining or negative North Atlantic Oscillation (NAO). To determine the age of colluvial deposits indicative of flood units at Ardena Gorge lichenometry was used. This technique allowed periods of increased flood frequency and river activity at the end of the LIA to be dated to the 1850s, 1880s and 1900s to 1920s, similar to those in France and Spain. Recent reductions in the number of intense precipitation events in the Eastern Mediterranean correspond directly to the increasing winter NAO, diverting winter precipitation to the North Mediterranean. More recently the number of floods generated at Ardena Gorge has fallen as fewer intense rainfall events occur (Maas and Macklin, 2002).

2.5 OVERVIEW OF PALAEOENVIRONMENTAL RESEARCH FROM EUROPE AND THE MEDITERRANEAN REGION

Across Europe and the Mediterranean palaeoenvironmental records have provided varying amounts of information regarding changes during the LIA. The dendrochronological record is relatively well developed across Central Europe, but more information is required for the Eastern Mediterranean. This can also be said of the LIA glacial records, which require further investigation. Although, palynological records have been undertaken throughout the Mediterranean region, many, particularly in the Eastern Mediterranean, concentrate on Quaternary long records. These records have suggested that vegetation is sensitive to climatic variability, with species indicative of particular climatic conditions identified in records from across Europe and the Mediterranean. In contrast fluvial records for the Mediterranean region have provided detailed information regarding LIA variability, identifying hydrological trends in the region. The general trends within each of these records have been summarised previously, with some similarities/contrasts identified.

During the first stages of the LIA only dendrochronological and fluvial records are available. Although there are contrasts between dendroclimatological records from Europe and the Calimani Mountains, many periods identified as cold in both records are registered as phases of high fluvial activity in the Mediterranean. Fluvial records suggest that AD 1570 - 1630 was a period of high fluvial activity initiated from increased rainfall, whilst dendrochronological records from Europe and the Calimani Mountains experience cool conditions. Reduced fluvial activity is experienced until AD 1640, when cool conditions are indicated across the European and Mediterranean region. This suggests that the climate became drier before it became warmer. By the mid to late 1600s fluvial records from across the Mediterranean registered low phases of fluvial activity and dendroclimatological records from the Calimani Mountains indicate above average temperatures (Popa and Kern, 2009). Prior to the LIA_{GM} experienced across Europe (AD 1780 - 1810), dendroclimatological records and fluvial records indicate that the 1700s were predominantly cold and wet. Even within the Calimani Mountains, cooler conditions are experienced from the mid-1700s to 1820s. Fluvial activity rises as temperatures fall across Europe and the Mediterranean region during the early 1800s. However, during the brief glacial retreat (AD 1810 - 1820) no climatic warming or increase/decrease in fluvial activity was registered. This may suggest that the glacial retreat seen in Europe between AD 1810 - 1820 resulted from local climatic variability, not experienced across the whole European/Mediterranean region. European glaciers soon began to expand once more until AD 1850 and this was accompanied by an increase in fluvial activity across the region. By the late 1800s Western Mediterranean glaciers were extending and the LIA_{GM} of Eastern Mediterranean glaciers was reached at AD 1878 - 1890. This phase coincides with high fluvial activity, but contrasts with dendroclimatological records that suggest the Mediterranean and European regions were undergoing a phase of warm conditions during this period. Records of glacial recession during the 1900s are more complete from European and Western Mediterranean records and coincide with periods of sustained low fluvial activity. Interruption of glacial recession coincides with increases in fluvial activity, suggesting increased rainfall may have allowed glaciers in the region to reach equilibrium or extend. The glacial record of the Eastern Mediterranean record is less detailed, but dendroclimatological records for the Calimani Mountains do suggest that the climate was significantly cooler than average until AD 1970. Generally since the AD 1980s the climate amelioration is apparent across the European and Mediterranean region, with reduced precipitation indicated by lower fluvial activity. Although the importance of local climatic impacts on different records cannot be overlooked as the Debeli Namet glacier experienced glacial expansion between AD 2004 - 2006 due to below average temperatures and increased winter precipitation (Hughes, 2008).

Overall, these palaeoenvironmental records have provided a backdrop of substantial research, confirming that ecological, glacial geomorphological, sedimentological and hydrological systems are sensitive to abrupt climatic fluctuations and can provide a record of how a system will react and change to these fluctuations. Current research has yet to bring palaeoecological, historical climate records, glacial geomorphology and catchment sedimentation records together in the Balkans to provide a catchment-wide interpretation of changes that occurred during the LIA. Without research connecting these components, attempts at understanding how climate impacts upon both the vegetation and sedimentation of a catchment cannot be completed.

The Mediterranean region is an area of great environmental, climatological and geomorphological diversity. This chapter aims to describe the setting and landscape of the Lake Plav catchment to provide context for later chapters.

3.1 GEOGRAPHICAL SETTING

The Dinaric Alps run the length of the Balkan Peninsula, from Slovenia towards the Pindus Mountains of Greece and lie in a north-west to south-east direction. Ridges of typical karst topography build across the Dinaric Alps to the Adriatic Sea where high coastal mountains are situated (Polunin, 1980; Kranjc, 2009; Figure 3.1). Typical karst topography is dominated by limestone geology and karst processes such as, dissolution of limestones and dolomite rocks through the action of carbonic acid. Carbon dioxide produced by plant root respiration and decay of organic matter is dissolved in water to create carbonic acid, vital for limestone dissolution (Jennings, 1985; Hose and Pisarowicz, 1999). Processes that occur as a result of carbonic acid action include: piping and tunnelling, subsidence as a result of support removal beneath land and collapse such as rock falls or rock slides (Jennings, 1985). Corrosion of karst terrain also occurs, where water dissolves limestone bedrocks leading to fissures within the limestone creating underground drainage channels and caves (Ford and Williams, 1989; Matsuda, 2004; Figure 3.2). Many streams appear ephemeral, seasonal or non-existent throughout the Dinaric Alps as a result of running underground (Ford and Williams, 1989).

3.1.1 The Lake Plav catchment

On the southern reaches of the Dinaric Alps the Prokletije Mountains span Montenegro, Albania and Kosovo. The Lake Plav catchment is situated within the Montenegrin and Albania portion of the Prokletije Massif. The catchment area totals 269 km^2 (Figure 3.3) and is characterised by a typical Dinaric karst landscape. Seasonal rivers exist in the catchment with flowing water apparent in the River Crna Dolja (Grbaja valley) and the River Grnčar (Grnčar valley) during autumn visits to the area, but hydrological data supplied by the United Nations Development Programme does not recognise the full extent of these and other rivers, as seen in Figures 3.3 and 3.14.

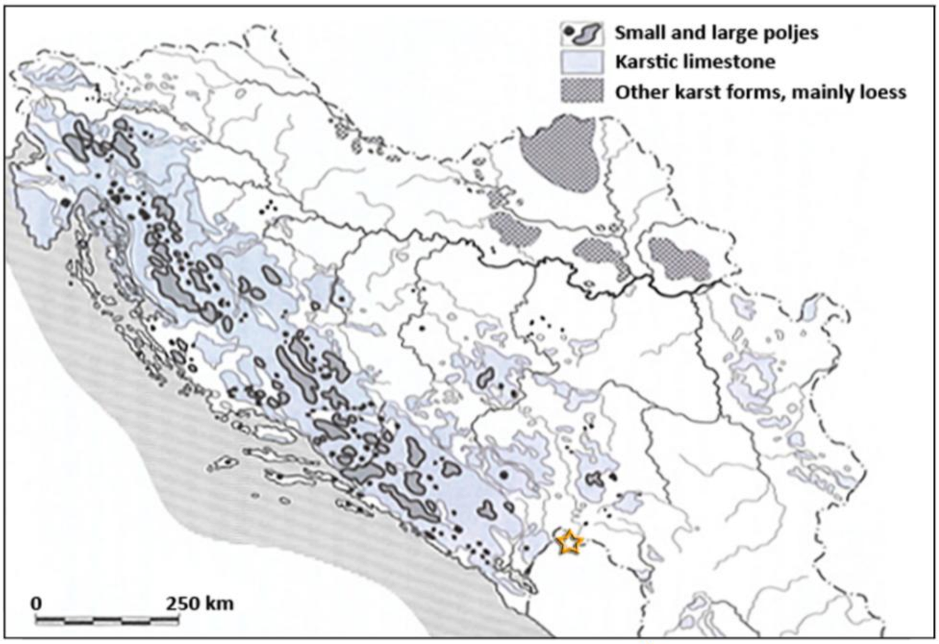


Figure 3.1: Map of Dinaric karst along the Adriatic coast ☆ represents location of Lake Plav catchment (adapted from Lewin and Woodward, 2009).

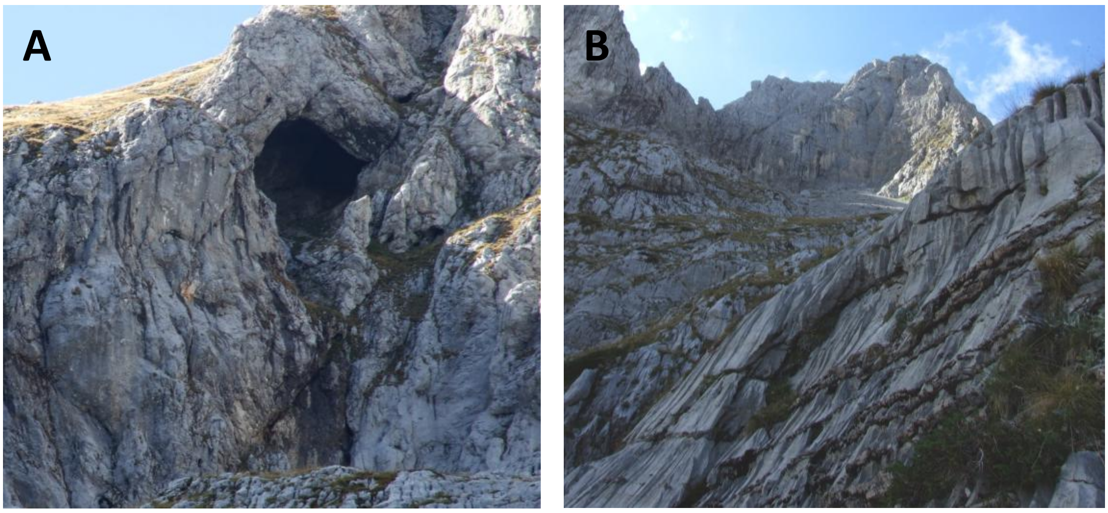


Figure 3.2: (A) Caves visible high above Buni i Jezerces and (B) smoothing and erosion of the limestone bedrock in the upper Lake Plav catchment.

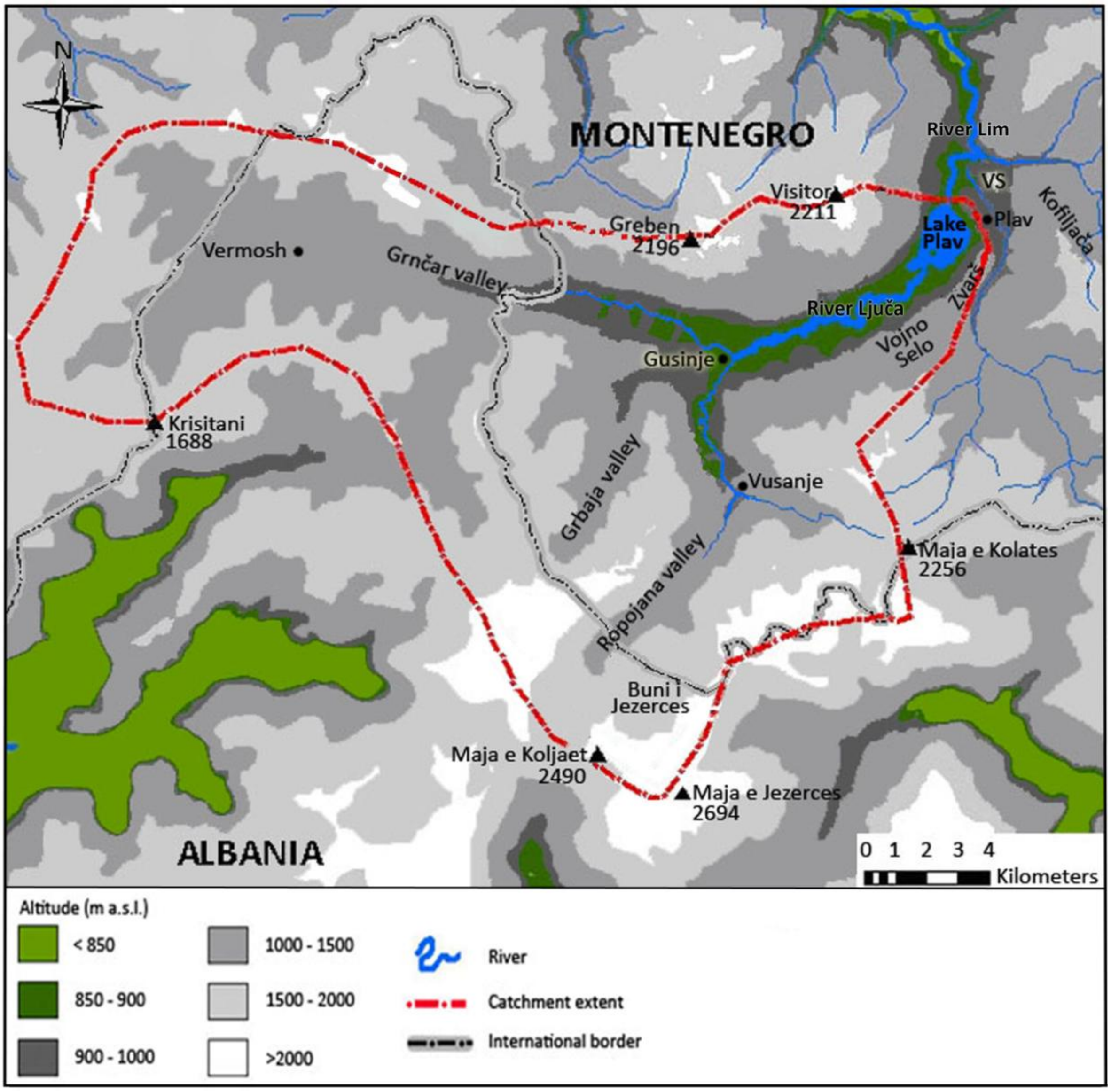


Figure 3.3: The Lake Plav catchment with altitudinal data provided for various features. VS = Veliki Skić. Data to create this map was supplied by Robert Aleksić (The United Nations Development Programme, Montenegro).

Glacial landforms, such as Pleistocene, Last Glacial Maximum (LGM) and Younger Dryas moraines overlay the karst environment (Cvijić, 1913; Milivojević *et al.*, 2008). In the upper extents of the catchment active geomorphological processes include frost heaving and shattering, creating large screes and widespread rock falls. Topographically, the most imposing feature of the lower catchment is Visitor Mountain which lies on the north-western edge of the lake, rising to 2211 m a.s.l. and the comparatively lower ridge of

Zvarš/Vojno Selo (1132 m a.s.l.) positioned south-east of the lake (Figure 3.3). Lake Plav then drains downstream towards Andrijevica and on to Berane as the River Lim. Across the catchment the altitudinal range is nearly 2000 m, rising from 904 m a.s.l. at Lake Plav, to 2694 m a.s.l. at Maja e Jezerces in the south of the catchment (see Figure 3.3). To the north of the catchment Greben (2196 m a.s.l.) and Visitor Mountain rise above the floodplain by around 1200 m. A variety of topographies and a broad altitudinal range exists at the Lake Plav catchment. This is evident from the steep slopes within the Ropojana and Grbaja valleys (Figure 3.4), which contrast to the relatively more vegetated and forested shallower slopes and wide floodplain of the lower Lake Plav catchment (Figure 3.5). The vertical dissection common within many Dinaric limestone catchments is less obvious in the Lake Plav catchment, compared to the extensive dissection of many high peaks, such as Severni Kranafil, further south into Albania. These characteristics highlight the unique nature of the Lake Plav catchment Mediterranean sedimentological studies, which have concentrated on steep sided gorges (e.g. Maas and Macklin, 2002; Macklin and Woodward, 2009). The impact of climate change on large shallow basins has not been a focus of previous studies. Therefore, this project may provide some insight into the effect climatic deteriorations in a catchment characterised by steep slopes in the upper areas and a shallow lower catchment.

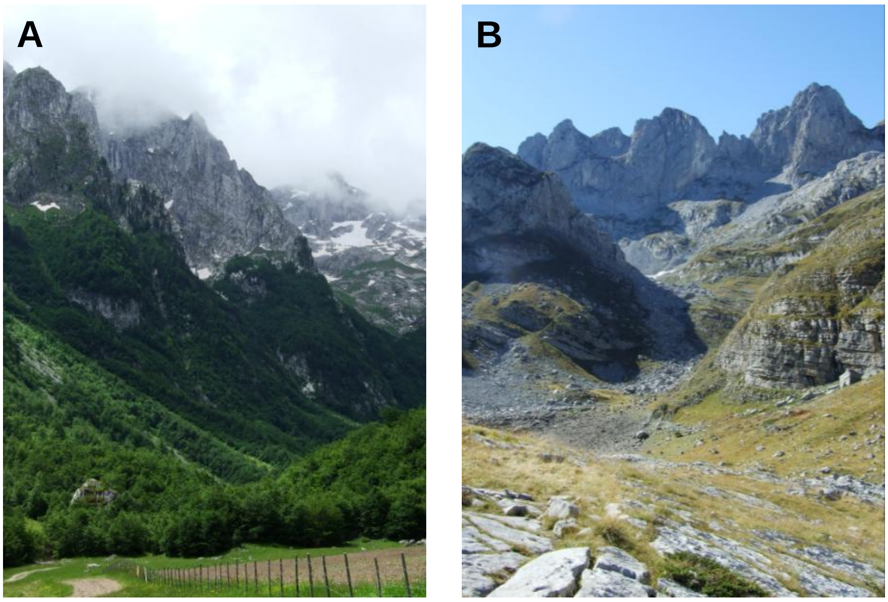


Figure 3.4: (A) steep valley sides of the Grbaja valley and (B) the ragged peak of Maja e Koljaet within the upper Lake Plav catchment above the Maja e Koljaet glacier.

Figure 3.5: The River Ljuča floodplain within the lower Lake Plav catchment.

3.1.2 Buni i Jezerces

Buni i Jezerces (Valley of Lakes) is located within the Albania area of the catchment, consisting of an upper and lower high mountain cirque beneath Maja e Koljaet peak (2490m a.s.l.). The site has been described by Milivojević *et al.* (2008) as the most complex cirque within the area, as the orientation of the cirque shifts from a north-western to north-eastern orientation as it leads into the lower cirque (Figure 3.4, B). A group of six glacial lakes are situated within the lower cirque, the largest being Liqueni Madhe, also known as Veliko Jezero (Abraham, 2007; Milivojević *et al.*, 2008). The only lake suitable for coring was Lake C (42^{o} 28’ 01.91”N, 19^{o} 48’ 43.0”E; 1754 m a.s.l.; Figures 3.6, 3.7), had limited clastic material on the surface of the lake bed and shallow water. Although there was evidence of drying and cracking on the lake bed, the presence of some water suggested pollen grains in the sediments may not have been desiccated in the past. The lake is also situated just above the tree line and provided a suitable site for extracting a primarily climatically driven palaeoenvironmental record.

Within the Buni i Jezerces site the Maja e Koljaet glacier exists beneath Maja e Koljaet peak. The altitudinal range from the peak to valley bottom (at Buni I Jezerces) is over 400 m, with Maja e Koljaet peak at 2162 m a.s.l. leading to the Maja e Koljaet glacier that reaches 1993 m a.s.l. and into the open valley floor that falls to 1754 m a.s.l. (Figure 3.4B, 3.6). Between Liqueni Madhe and the four un-named lakes (Figure 3.6) is a small lake which marks the site of a possibly Younger Dryas terminal moraine that exists between 1700 m and 1800 m a.s.l. (see Figure 3.6, 3.8). This moraine is not visible on any local maps of the area, but is a geomorphological feature that provides indication of past glaciations in the area.

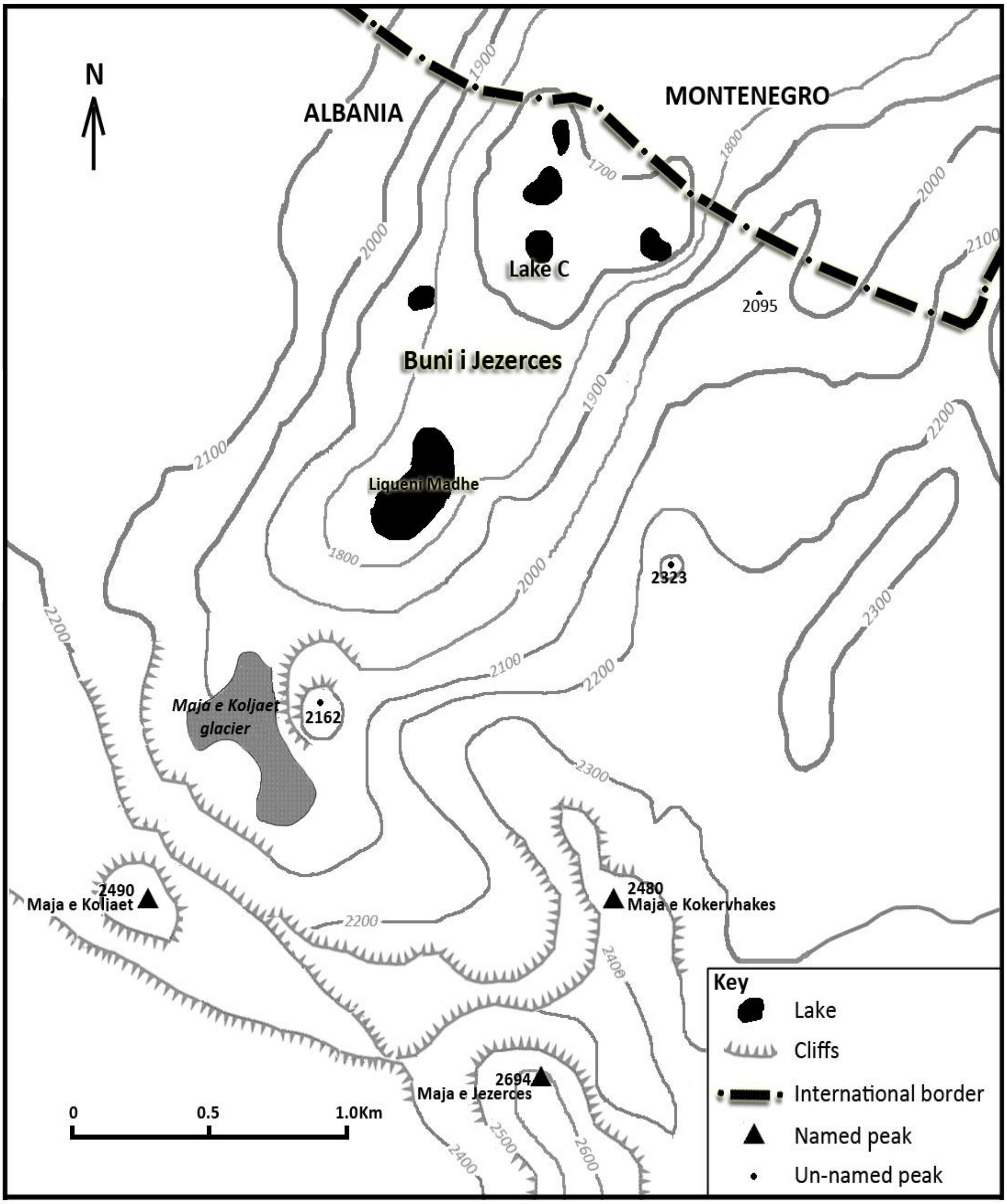


Figure 3.6: Map showing the Buni i Jezerces sites with Lake C and Maja e Koljaet glacier.

Figure 3.7: Photograph of Lake C in the Buni i Jezerces complex, the site for extraction of core BJC1.

Figure 3.8: Possible Younger Dryas terminal moraine in the middle ground, at Buni i Jezerces.

3.1.3 Maja e Koljaet glacier

The Maja e Koljaet glacier is located at Buni i Jezerces between 1993 – 2029 m a.s.l. (42^{o} 27’ 12.5”N, 19^{o} 47’ 52.8”; Figures 3.6, 3.10), and is located in a north-east facing niche providing ideal conditions for preservation of accumulated snow (Kruss and Hastenrath, 1987; Kuhn, 1989; Chueca and Julian, 2004). Maja e Kokervhakes and Maja e Jezerces lie to the east and south-east of the glacier providing further shade for the development and maintenance of a small high mountain glacier (Figures 3.6, 3.10). The recent extent of the glacier is being calculated using the metal measuring wire that is at least 3m above the current glacier (Figure 3.9). Kamil Weinberg (mountaineer) identified the measuring wire as an aid for monitoring the Maja e Koljaet glacier and it was attached to the boulder by Milovan Milivojević in AD 2005. Around Maja e Koljaet and the near-by peaks of Maja e Jezerces and Maja e Kokervhakes, steep sparsely vegetated cliffs characterise the typical karst topography with corrosion on surfaces across the area and some vertical dissection (Figures 3.8, 3.10).

Figure 3.9: Measuring wire at the Maja e Koljaet glacier, the thin metal wire is attached at the red stripe.

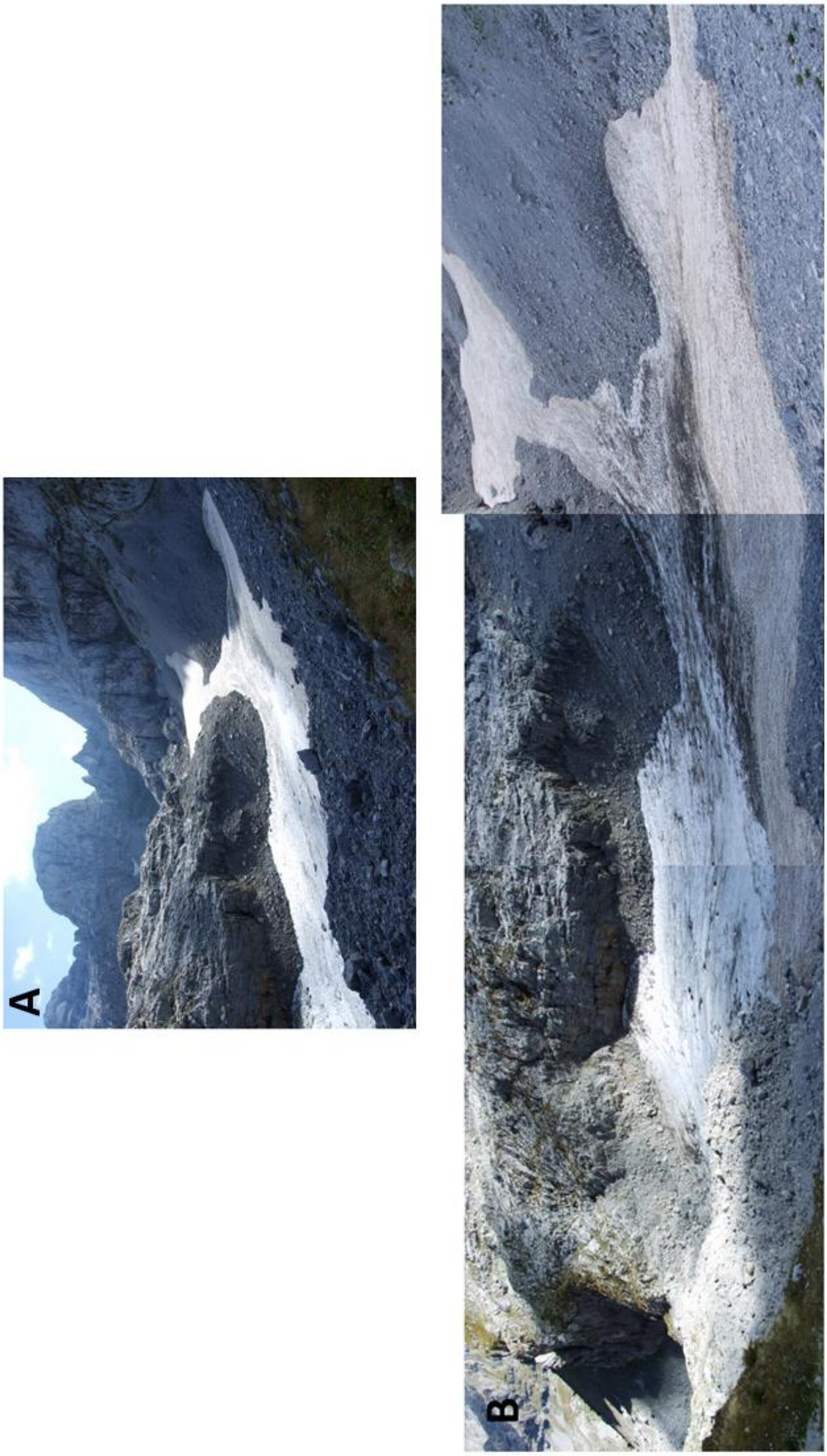


Figure 3.10: A, vertical dissection and scree slopes, and B, the Maja e Koljaet glacier.

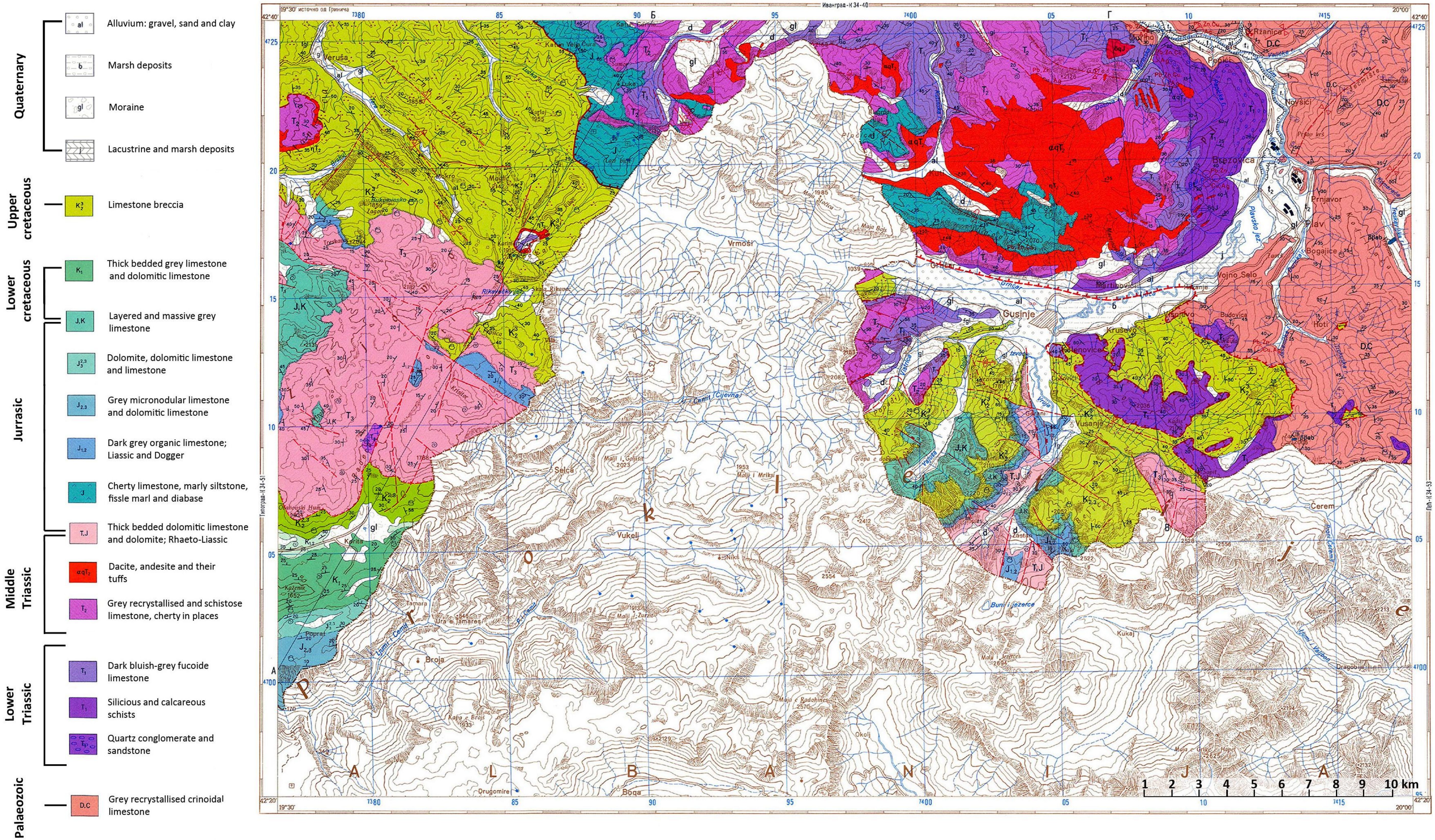


Figure 3.11: Geological map K34-52 (scale 1:100,000) for the Lake Plav catchment (adapted from The Institute for Geological Exploration, 1968). ***Plavsko jez.*** identifies Lake Plav.

3.2 GEOLOGICAL SETTING

As previously mentioned limestone dominates the Dinaric Alps geology and the Prokletije Mountains/Lake Plav catchment are no exception. The geology of the Lake Plav catchment is of sedimentary and magmatic rocks from the Palaeozoic, Triassic, Jurassic and Cretaceous periods. Figure 3.11 illustrates the geology of the Lake Plav catchment and the blank areas represent Albania territory for which geological maps cannot be accessed. Geological map K34-52 was created by the Institute for Geological Exploration, Titograd (Podgorica) between AD 1958 - 1968.

The geology of the Lake Plav catchment can be split in two, with the area covering Visitor Mountain and the Greben portion of Grnčar Valley differing somewhat to that of the southern catchment below the River Ljuča across Durika Rijeka and the Ropojana and Grbaja valleys. A mixture of diortic and diortic-porphyritic dikes in Lower Triassic sediments occur on the eastern slopes of Visitor Mountain, and Middle Triassic limestone and eruptive rocks that alternate with dacite. The Lower Triassic sediments of Visitor are folded and metamorphosed, containing sulphide ore emplacements. Middle Triassic rocks along Greben were formed at the end of the period, when volcanic activity had shifted from being a, 'subtle effusion of lava to gaseous volcanic activity forming chert masses'. The southern catchment is dominated by thick dolomitic limestone and dolomites in the Ropojana valley containing fauna such as *Triasina hantkeni* and *Palaeodasycladus mediterranea*, making the geology of this area similar to that of the Morača Valley (~ 40 km west of Lake Plav). In the area of Gusinje, limestone breccia dominates with some metamorphosed sub-flysch. In Grnčar Valley Quaternary formations are present including; fluvial formations, lake-marsh sediments, alluvial and deluvial (flood) deposits. Within the general geology of the area, there are small intrusions of rocks such as schist and sandstones with lenses of conglomerate and quartzite across Vojno Selo and Plav town, where a seismic station is positioned on sandstone and phillite soils (Institute for Geological Exploration, 1968; Geofizička mreža Crne Gore; 42.5950N, 19.9735E). Overall, the Lake Plav catchment is dominated by limestone and dolomitic limestones overlain by more recent Quaternary aged sediments.

The tectonic framework of the Lake Plav catchment is more complex with the area covered by map K34-52 being divided into three structural units: (i) the High Karst zone, (ii) the Durmitor over-thrust zone and (iii) the flysch zone. Fault lines can be seen across the catchment with the longest running along Grnčar Valley to between Gusinje and Plav before shifting southwards towards Visnjevo. Current seismological reports from the Geofizička mreža Crne Gore (Geophysical network of Montenegro, 2010) highlight the current seismic activity in the area with monthly reports available for Montenegrin sites, including Plav. These reports suggest that there is currently activity across the country, but the strongest most recent earthquake to affect Montenegro was that of AD 1979 which occurred along the Montenegrin coastline and measured 7.0 on the Richter scale, lasting for 10 seconds (Rovelli, 1984; Armenian *et al.*, 1992; Viti *et al.*, 2003).

3.3 CLIMATE

Montenegro is a climatically diverse country with a Mediterranean climate on the Adriatic coast, characterised by hot dry summers and cooler wet winters (McNeill, 1992; Lionello *et al.*, 2006). This climate can be described as a 'dry summer subtropical' climate using the Köppen classification (Köppen 1923; 1936 cited in Lohmann *et al.*, 1993; Gerstengarbe and Werner, 2009). Inland a Continental Mediterranean climate exists as a result of higher altitude and reduced influence of the Mediterranean Sea, producing winter precipitations almost equal to that of the summer months (Tošić, 2004). Continental climates are dominated by higher summer precipitation and more central/Eastern European countries such as Serbia are affected by such a climate (Unkašević and Radinović, 2000). Climate records for the Prokletije Mountains are sparse with only brief data available for the Lake Plav catchment provided by a meteorological station at Vermosh in Albania (see Figure 3.3). The incomplete data for Vermosh suggests that, for an unknown date, the annual temperature range was 18.6^{o}C, when the mean January temperature was -2.6^{o}C and mean July temperature equalled 16^{o}C (Palmentola *et al.*, 1995). Summer temperatures breach the 10^{o}C threshold connected to growth of trees described by Köppen indicting that any tree growth would be constrained to summer months in the area (Lohmann *et al.*, 1993). Precipitation can be ascertained using climatic models and current climate records from other close mountain sites, which suggest rainfall in the area to be between 2500 – 3000 mm yr^{-1} (Bošković and Bajković, 2006). Heavy rainfall events are relatively common in Montenegro, with Llasat and Rodriguez (1997) identifying 34 years, between AD 1923 - 1984, when more than 200 mm of precipitation fell within 24 hours. Only Italian records had a comparable record with 39 out of 67 years including high rainfall events. Both records indicated that the season most likely to have such rainfall events was identified as autumn (Llasat and Rodriguez, 1997). However, the Italian data was an average of data from 100 climate stations, but that of Montenegro was averaged from a single unnamed climate station. Climate records for towns such as Kolašin (954 m a.s.l.) 40 km north-west of the Lake Plav catchment are thought to provide comparable records due to similar situation and altitude. Figure 3.12 provides a summary of climate data available for Kolašin over a 35 year period (AD 1973 - 2008), using summer (Jun/Jul/Aug) and winter (Nov - Jan) averages. These suggest that winters in the area are relatively cool, with an average of 0.61^{o}C for the 35 year period, but not very wet, with an average of 113 mm each winter at the site (Tutiempo, 2009). Once more this data is incomplete with data absent from 68/420 months and no data available for AD 1984 (highlighted by the gaps in data Figure 3.12). Tošić (2004) provided further information for summer precipitation levels at Lake Plav, using analysis of winter and summer precipitation over former Yugoslavia from 30 meteorological stations including Kolašin. Figure 3.13 shows the modelled variability across Montenegro and the similar winter and summer precipitation around Lake Plav; winter (Dec/Jan/Feb) is near to 100 mm and during summer months (Jun/Jul/Aug) 70 – 80 mm rainfall is suggested. Winter precipitation is likely to be greater than summer precipitation within the catchment, but unlike typical Mediterranean climates in areas such as Southern Greece, there is no summer drought (Tutiempo, 2009). As the greatest precipitation does not fall during the summer months the climate cannot be described as Continental and must therefore be expressed as a Continental Mediterranean climate (Köppen, 1923; 1936 cited in Lohmann *et al.*, 1993; Gerstengarbe and Werner, 2009). Overall, the Lake Plav catchment has a high annual rainfall of >2500mm yr^{-1}, but with evidence of mean summer temperatures at 16°C and above 10°C the conditions are ideal for sustaining deciduous woodland.

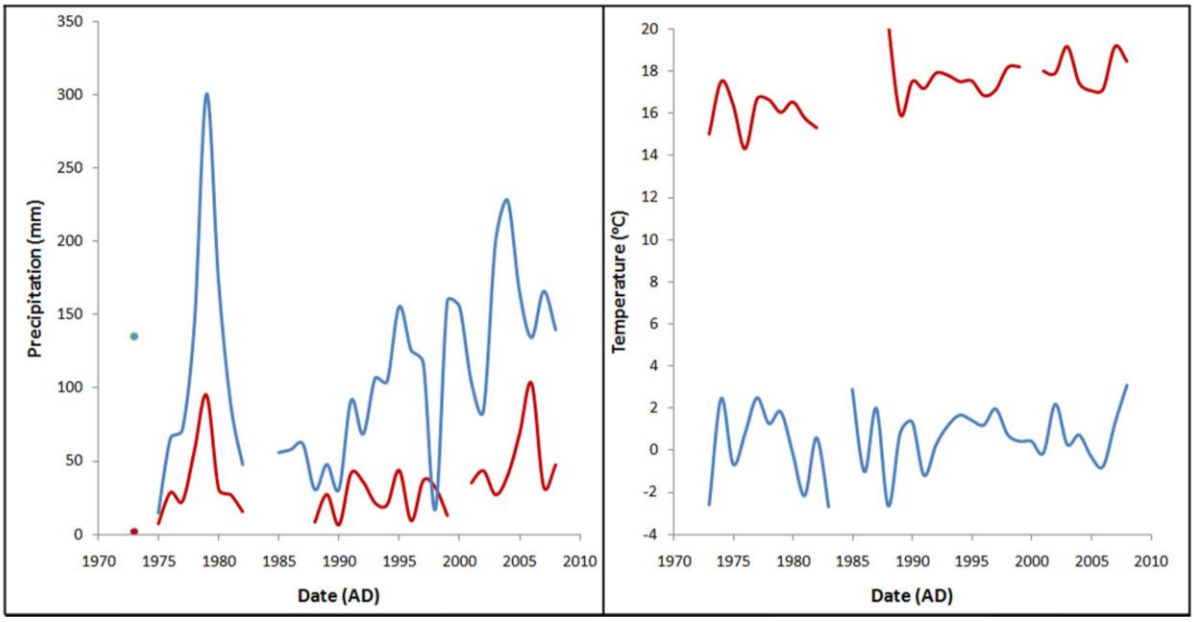


Figure 3.12: Average climate data for Kolašin (954 m a.s.l.), (A) precipitation data and (B) temperature data. Summer averages in red and winter averages in blue (data taken from Tutiempo, 2009).

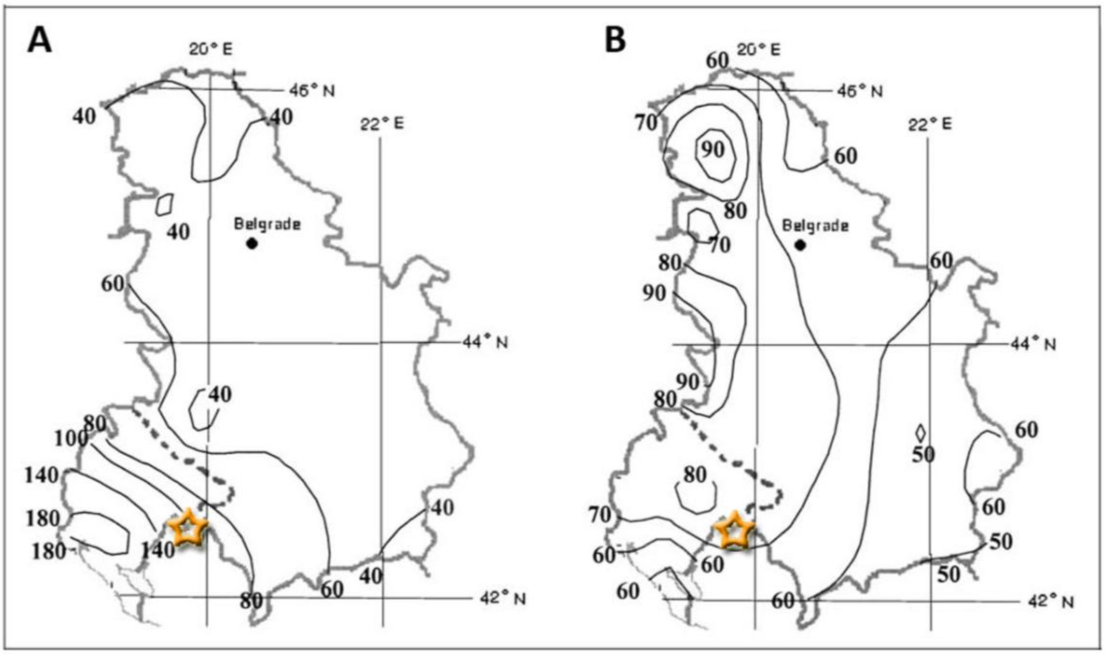


Figure 3.13: Modelled winter (A) and summer (B) precipitation over the period AD 1951 - 2000 (adapted from Tošić, 2004), ☆ represents Lake Plav catchment.

3.4 HYDROLOGICAL SETTING

Lake Plav is situated in the Montenegrin portion of the Prokletije Mountains (42^{O}53’47”N, 19^{O}55’48” E) at an altitude of 904 m a.s.l. (Figure 3.3) and is known locally as Plavsko Jezero (Blue Lake). The lake is dammed by a Pleistocene moraine (Cvijić, 1913; Kunaver, 1991; Milivojević *et al.*, 2008; see Appendix II) and lies in a U-shaped valley formed during previous glaciations of the area (Dobiński, 2005). Maps of Lake Plav created by Cvijić in the early 1900s show the area of the lake has shrunk by 42% over approximately 90 - 100 years (Figure 3.14). The current lake is around 2100 m in length, 1400 m wide and relatively shallow at a maximum of 9 m. During spring snowmelt, when deep circulation of karst ground-waters is typical, large quantities of eroded material are washed down valley to Lake Plav at the terminus of the catchment (Cerdà, 1998; Milivojević *et al.*, 2008). The River Ljuča feeds Lake Plav with three rivers converging to create the Ljuča: River Grnčar (Lumi I Vermoshit in Albania), River Crna Dolja and River Grlja, with the two former rivers appearing seasonal. Intermittent or seasonal streams are typical of karsified limestone terrains, due to the varying hydro-geological conditions in these areas as their discharge regime can alter both spatially and seasonally (Gordon *et al.*, 1992; Meyer and Meyer, 2000; Meyer *et al.*, 2003). A number of intermittent streams are apparent in Figure 3.14, with the river course seemingly disappearing. It is most likely that the rivers course diverts to an underground course at the point of disappearing, feeding the local groundwater. The effect of the converging rivers and inputs from the prograding River Ljuča are sediments transported and deposited by the river have caused Lake Plav to infill changing the size and proportions of the lake. The local population are also concerned about the current rate at which Lake Plav is infilling. A local environmental NGO (headed by Mensur Markisić) is currently concerned with the habitational and environmental impacts the infilling has for the future of the lake and surrounding populations.

The tributary rivers of the River Ljuča are wider and shallower with a coarse bed-load. Between Gusinje and Plav the river channel morphology is more contained with vegetated floodplains either side of the River Ljuča, but lacks the wide expanses of cobbles and boulders seen further upstream at the River Grnčar (Figure 3.15). Little more is known of the rivers feeding Lake Plav, but the outflow of Lake Plav, the River Lim, has been studied and revealed to have an average discharge within its lower course of 71 m^3 sec^{-1} (Bošković and Bajković, 2006). Hydrological studies have not been undertaken within the Lake Plav catchment, but comparison of maps drawn by Cvijić (1907; 1913; 1917, original presented in Appendix II) and Lokalna turistička organizacija Plav (2006) in Figure 3.14, provide some idea of how the catchment and in particular how Lake Plav has changed. Seasonal changes can also be observed at Lake Plav with photographs just after spring melt and during the autumn showing changes in the water levels (see Figure 3.16). Following spring melt in 2007 the water level reached the grassy meadow land in the centre of Lake Plav (Figure 3.16, A). In contrast the photograph taken in October 2009 suggests a lower lake level with a dark brown 'muddy shoreline' visible around the lake (Figure 3.16, B). Figure 3.12 suggests that the winter prior to the 2007 spring melt, was average for the area around Kolašin and if Plav underwent similar winter conditions then the lake levels recorded by Figure 3.16, A are likely to be representative of the seasonal variations at Lake Plav. Trees situated upon the meadow land are particularly useful in identifying the location and extent of the seasonal changes experienced at Lake Plav. The tree situated on the edge of meadow land stretching into Lake Plav offers a useful marker as its proximity to open water and the extent of a muddy shoreline changes between the May and October photographs (Figure 3.16).

Evidence of significant flooding throughout the Mediterranean region has been underlined by the work of Llasat and Rodriguez (1997). The most recent flood at Lake Plav occurred on 14th November 2010 and was captured by both Ahmet Rekovic of the Planinarskog Drustva Prokletije Plav and Goran Petrusic of Montenegro Red Cross (see Figures 3.17 - 3.19). Extensive overbank flooding occurred across the River Ljuča floodplain from Plav to Gusinje causing the Albanian refugee accommodation to be flooded (*pers. comm.* Ahmet Rekovic). The extent of the flooding can be seen within Figures 3.17 - 3.19, with roads and homes partially/fully submerged throughout the catchment. Figures 3.17 and 3.18 identify the devastation seen within Grbaja valley and the force such floods can inflict upon a catchment and local residents. Local records detailing flooding events are unavailable for the area, but during the field work of autumn

2006 and 2009 extensive flooding was not witnessed. Flooding to the extent seen in November 2010 may well be an 'extreme' event seen on a less regular timescale.

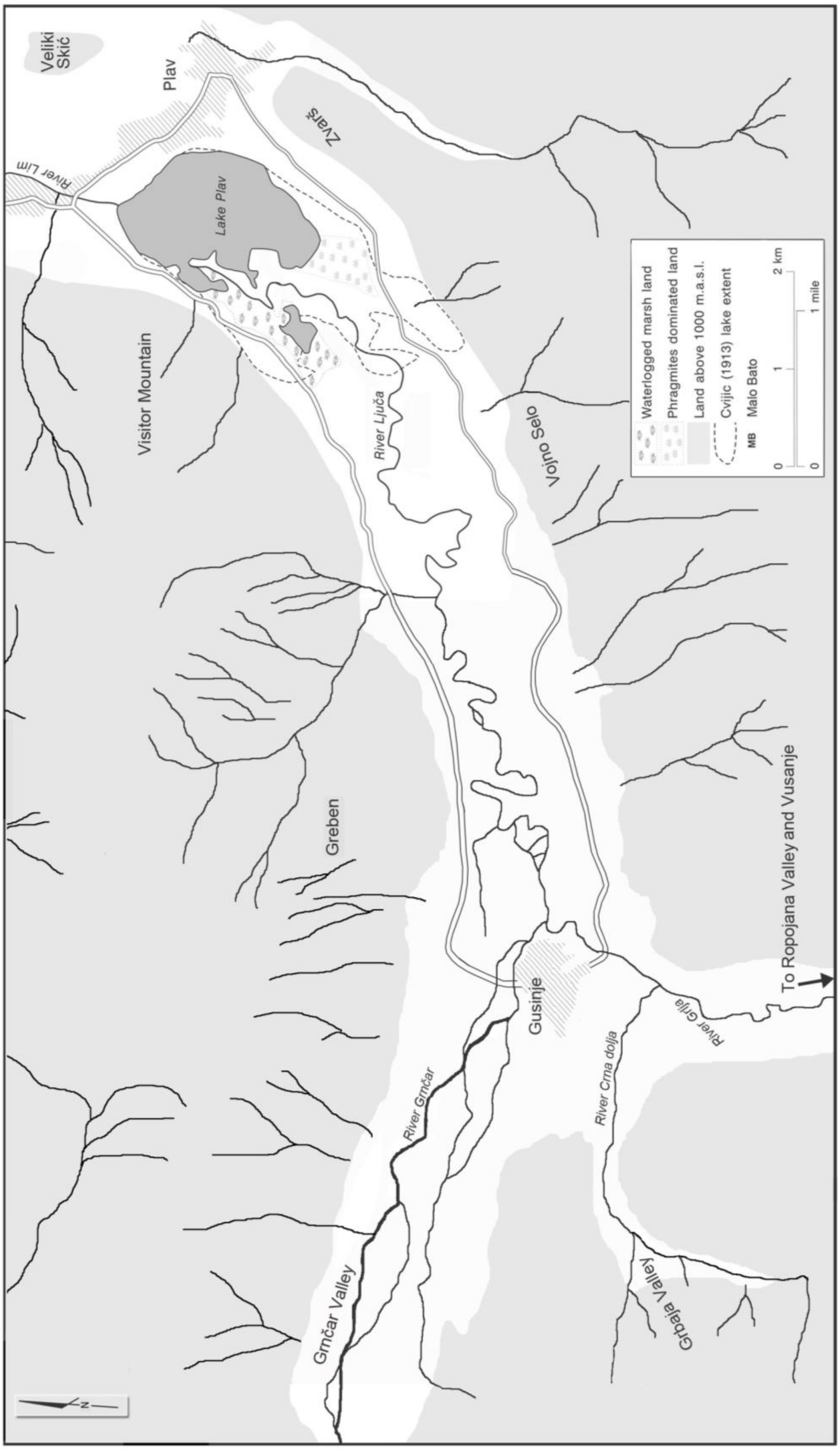


Figure 3.14: Comparison of the Lake extent mapped by Cvijić in 1913 to that of the present day using information from the recent Huber Verlag/Lokalna turistička organizacija Plav (2008).

Figure 3.15: The braided-type course of the tributary River Grnčar.

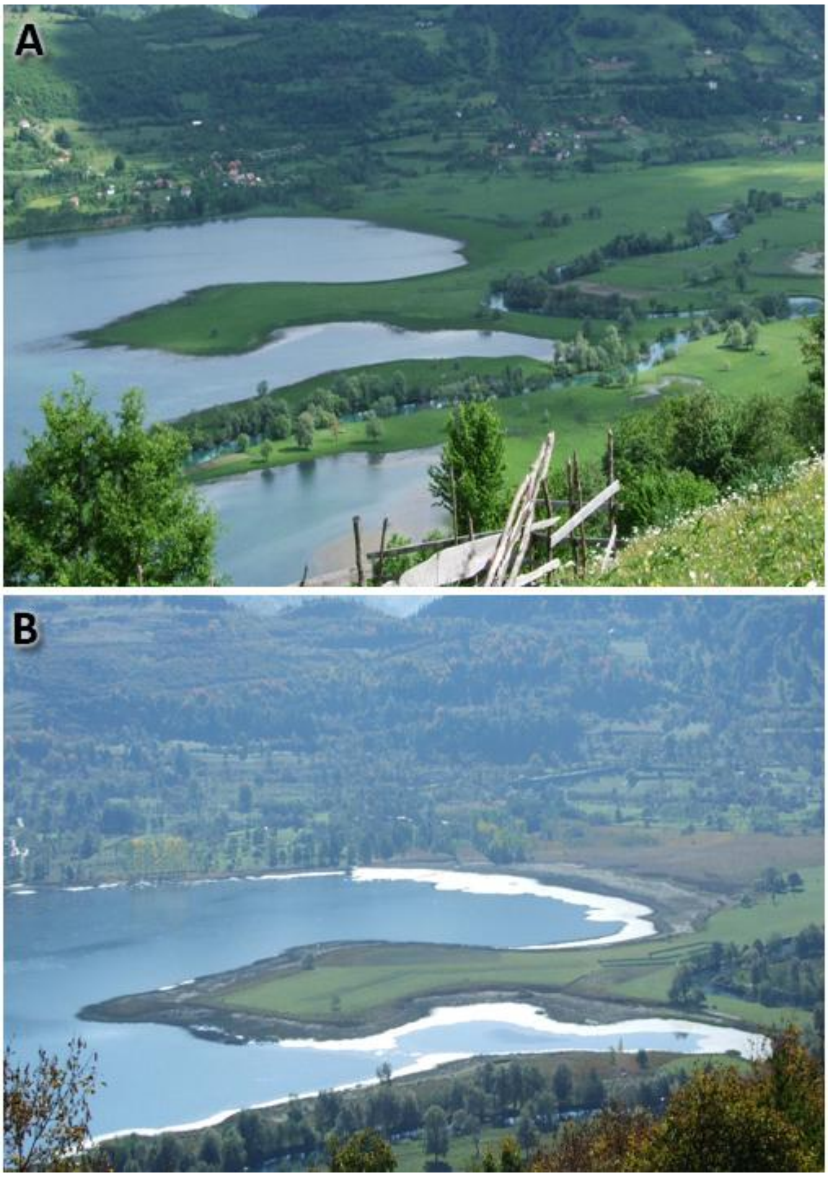


Figure 3.16: Varying lake levels at Lake Plav. A, taken in May 2007 following spring melt and B, taken in October 2009.

Figure 3.17: Top three images provide evidence of flood extent and power with roads swept away and the lower images identify the extent and rise in the Lake Plav water level (Photographs supplied by Ahmet Rekovic of the Planinarskog Drustva Prokletije Plav).

Figure 3.18: The Albanian refugee accommodation on the outskirts of Gusinje; (A) during 14th November 2010 flooding (Photograph supplied by Goran Petrusic, Montenegro Red Cross) and (B) in October 2009.

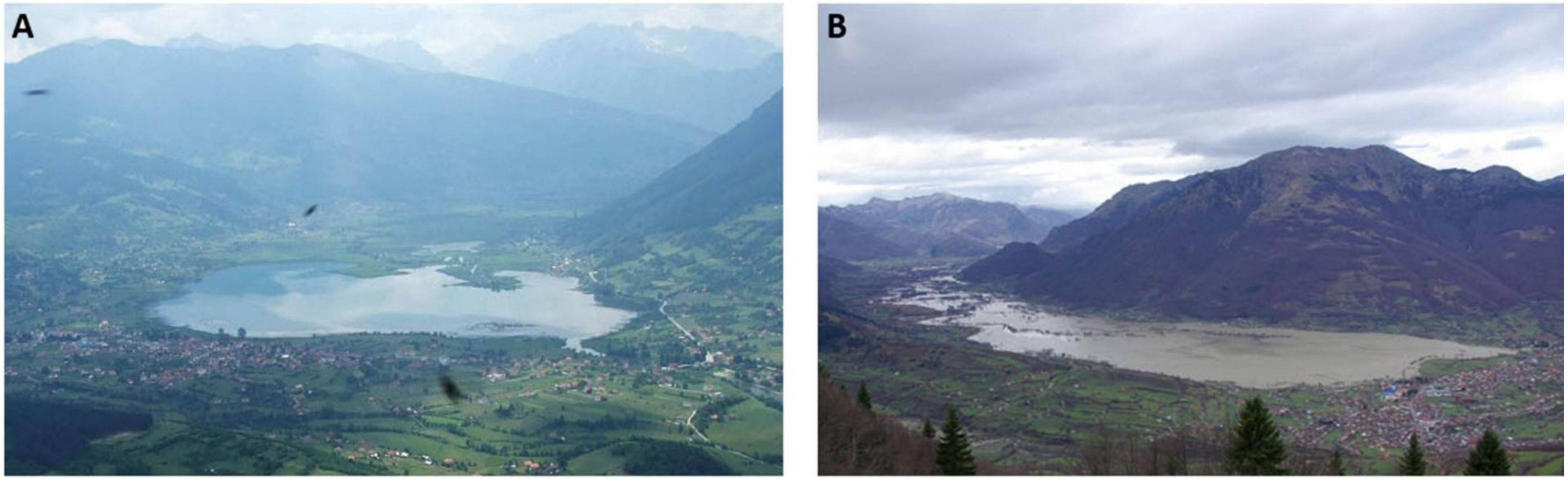


Figure 3.19: Two photographs of Lake Plav looking towards Gusinje. (A) Taken in May/June 2007 (photograph taken by Rose Wilkinson) and (B) Flooding in the Lake Plav catchment 14th November 2010 (photograph supplied by Ahmet Rekovic of Planinarskog Drustva Prokletije Plav).

3.5 CURRENT VEGETATION

Balkan plant-life is richer than any comparable area within Europe and forms part of a biodiversity hotspot with over 6530 species of native seed plants (Polunin, 1980; Gaston and Davis, 1994; Hanson *et al.*, 2009). Polunin (1980) stated four reasons for this biodiversity:

1) The presence of old flora containing species that have survived Quaternary Ice ages, as proved by reconstruction of vegetation refugia throughout Greece (e.g. Tzedakis, 1993; Tzedakis *et al.*, 1997; Bozilova and Tonkov, 2000; Taberlet and Cheddadi, 2002).
2) Islands and mountain habitats causing fragmentation and isolation of species, forcing species migration and allowing the formation of new habitats.
3) Migration from close proximity areas such as central Europe.
4) Anthropogenic influences that have modified habitats and vegetation cover, and introduced new species to the Balkan Peninsula.

Polunin (1980) also produced a map of Balkan and Greek vegetation, which highlighted the variety of flora habitats that exists through Montenegro/Albania and provided general vegetation zonation. Six types of vegetation are thought to dominate Montenegro/Albania and are summarised in Table 3.1. The coastal regions (< 700 m a.s.l.) of the countries are dominated by Mediterranean evergreen forests and maquis alongside transitional deciduous forests. Further inland between 700 – 1700 m a.s.l. montane beech and coniferous forests exist, with pockets of sub-alpine and alpine zone. The Lake Plav catchment lies between 904 – 2100 m a.s.l., falling within the Central European montane beech forest/coniferous forest and sub-alpine, alpine zones of Polunin (1980). Montane beech (*Fagus*) forests are widespread along the Dinaric Alps with Balkan sites often exclusively composed of *Fagus*. The quick regeneration of *Fagus* forests has ensured their continued existence and allowed their range to spread as coniferous types such as *Pinus nigra* (Black pine) are felled. *Pinus nigra*, alongside other coniferous types (presented in Table 3.1), is a species that forms part of the Balkan coniferous forests (Polunin, 1980). Throughout the Balkans deforestation has occurred since the Bronze

Age, particularly in coastal regions. During the early 1300s the Balkans became part of the Ottoman Empire, with only coastal regions remaining out of Turkish control. This had a direct impact on the vegetation of the Balkan interior (Dinaric karst area) as deforestation was carried out by both the Turkish army and civilians to destroy hiding places, make room for cavalry and prevent attack (Kranjc, 2009). Since the 1850s *Pinus nigra* has been used to complete reforestation of devastated areas (Kranjc, 2009). This is evident on the hill of Glaviće near Plav, where a strip of planted *Pinus nigra* forest is present (Figure 3.20). Finally, meadows in the sub-alpine and alpine zone are thought to be the result of anthropogenic activity following clearing of upper forest and shrub communities to provide grazing for livestock.

Figure 3.20: Strip of *Pinus nigra* forest visible upon Glaviće.

Overall, Polunin identified the area across the Dinaric Alps as one dominated by central European montane beech and coniferous forest, which Kranjc (2009) described as the natural vegetation of dinaric karst. Anthropogenic manipulation of the landscape since the Neolithic has caused the erroneous view of karst areas lacking dense vegetation. During this time clearing of forests began along the coast, but mountainous areas such as Prokletije were not penetrated (Velušček, 1999; Kranjc, 2009). Prokletije like many of the inland mountainous areas has remained relatively free from intensive agriculture with a concentration on 'garden' farming (McNeill, 1992). Therefore, the current vegetation in such mountainous areas is one of valley floors modified by small concentrated local populations with vegetational management (e.g. coppicing) of the surrounding forests on mountain sides. Within the Prokletije Massif deciduous and coniferous woodland dominate, with *Abies* and *Fagus* relics of the climatic forest that developed in the mountainous areas of the Dinaric Karst during the Atlantic period (Kranjc, 2009).

The vegetation of the Lake Plav catchment is further discussed within Chapter 5, Results, following vegetation surveying of the site during this research.

Altitudinal range (m a.s.l.)	Zone	Selection of species present within the zone
< 700	Mediterranean evergreen forest	*Pinus halepensis* (Aleppo Pine); *Quercus ilex* (Holm Oak); *Juniperus oxycedrus* (Prickly juniper); *Sarcopoterium spinosum* (Brushwood)
<700	Maquis	*Pinus halepensis* (Aleppo Pine); *Pistacia lentiscus*; *Quercus coccifera* (Kermes Oak); *Myrtus communis* (Common Myrtle); *Genista acanthoclada* (Hairy canary clover)
<700	Mediterranean transitional deciduous forest	*Fraxinus ornus* (Manna ash); *Quercus pubescens* (White Oak); *Carpinus betulus* (Hornbeam); *Corylus* (Hazel); *Celtis australis* (Hackberry); *Sorbus* (Whitebeam); *Acer campestre* (Field maple)
700 – 1700	Central European montane beech forest	*Fagus sylvatica*(Beech, dominant); *Abies alba* (Silver fir); *Picea abies* (Spruce); *Pinus sylvestris* (Scots pines); *Ulmus glabra* (Elm); *Corylus* (Hazel); *Salix caprea* (Willow); *Oxalis acetosella* (Common wood sorrel); *Euphorbia amygdaloides* (Wood spurge)
700 – 1700	Central European coniferous forests	*Pinus nigra* (Black pine); *Pinus sylvestris* (Scots pine); *Picea abies* (Spruce); *Ostyra* (Dogwood); *Acer* types (Maple); *Tilia* (Lime); *Fraxinus excelsior* (Ash)
> 1700	Sub-alpine, alpine	*Pinus mugo* (Mountain pine); *Juniperus communis* (common juniper); *Alnus virdis* (Slide alder); *Rhamnus alpinus* (Buckthorn); *Sideritis* (Mountain tea); *Thymus parnassicus* (Wild thyme)

Table 3.1: Vegetational zones present within Montenegro following Polunin (1980).

4.1 FIELD METHODOLOGY

Fieldwork was undertaken in the Lake Plav catchment between November 2006 and October 2009 during four separate field campaigns. These involved coring at Lake Plav itself, vegetation surveying, glacial geomorphological mapping, lichenometry and also further coring of lake basins in the upper catchment.

4.2 LOCALITY OF LAKE PLAV

The objectives at the Lake Plav site were to extract a number of boreholes to assess where the most suitable cores could be extracted for multi-proxy analysis. This then allowed sediment cores to be collected with manual and mechanical corers.

4.2.1 Sediment coring

Coring of sediments is the only efficient method of extraction to provide a complete and undisturbed sequence in a lake or lake edge location. The type of corer used is dependent on the site and aims of a project. Terrestrial extractions generally use chamber samplers such as the Russian and Hiller samplers to extract sediment instantaneously as the sampler is turned, ensuring limited sediment contamination (Aaby and Digerfeldt, 1986). Sampling from the centre of lakes can be undertaken using a Russian corer if the material is sufficiently unconsolidated, but Mackereth or Livingston corers are the more typical types used (Livingstone, 1955, 1967; Mackereth, 1958; Cushing and Wright, 1967).

For this research project sediment depth tests using a gauge corer around the periphery of the lake showed that on the eastern side of the River Ljuča depths were limited to around 1.5 m. Coarse grained material is thought to have limited coring here, but this fraction could not be sampled for reliable identification. Numerous homes and a main road from Plav to Gusinje resulted in limited access to the areas around Malo Bato and beneath the slopes of Visitor Mountain (see Figure 4.1). Extracting cores close to the slopes surrounding Lake Plav (Visitor Mountain and Zvarš; Figure 4.1) would have provided a palaeoenvironmental record related to changes associated with these slopes. This would not have fulfilled the aims of the project; to capture catchment wide changes. Therefore, a transect across the floodplain from the River Ljuča outlet to the eastern side of Lake Plav was more suitable. Following the map by Cvijić (1913; Figure 3.14 Chapter 3) it was hoped a transect along the periphery of the lake would provide a more lake-based record as these sites would have been submerged in AD 1913. Following preliminary investigations coring was constrained to the area shown in Figure 4.1 and the final sites for core extraction were established. All coring sites are terrestrial locations, but seasonally flooded during spring melt. Sediments were initially extracted using a Russian sampler, but due to the high clay content of the sediments this had to be substituted for a gauge corer. Gauge corers do not place sampled sediments in a chamber causing sediment smearing during extraction. The surface was therefore cleaned to allow accurate logging of the sediments and packaging for transport back to the UK.

During the November 2006 fieldwork core LPCA was extracted from the Lake Plav site, to a depth of 8.3m (42^{o} 35’ 24.4” N, 19^{o} 55’ 25.2” E; Figure 4.1). The site was located on hummocky marsh-like ground, dominated by *Phragmites* (common reed). It was selected due to its distance from the current and past location of the River Ljuča and disturbed slopes around Zvarš. A gauge corer (20 mm and 50 mm widths) was used to extract the core in 0.5 m sections and was synchronously logged to record any changes in sediment composition or hiatuses. Sections were wrapped in cling film and plastic sheeting to provide an airtight seal, ensuring no desiccation or contamination could occur, and then stored in cool (1 – 3^{o}C) conditions (following Lowe and Walker, 1984; Aaby and Digerfeldt, 1986).

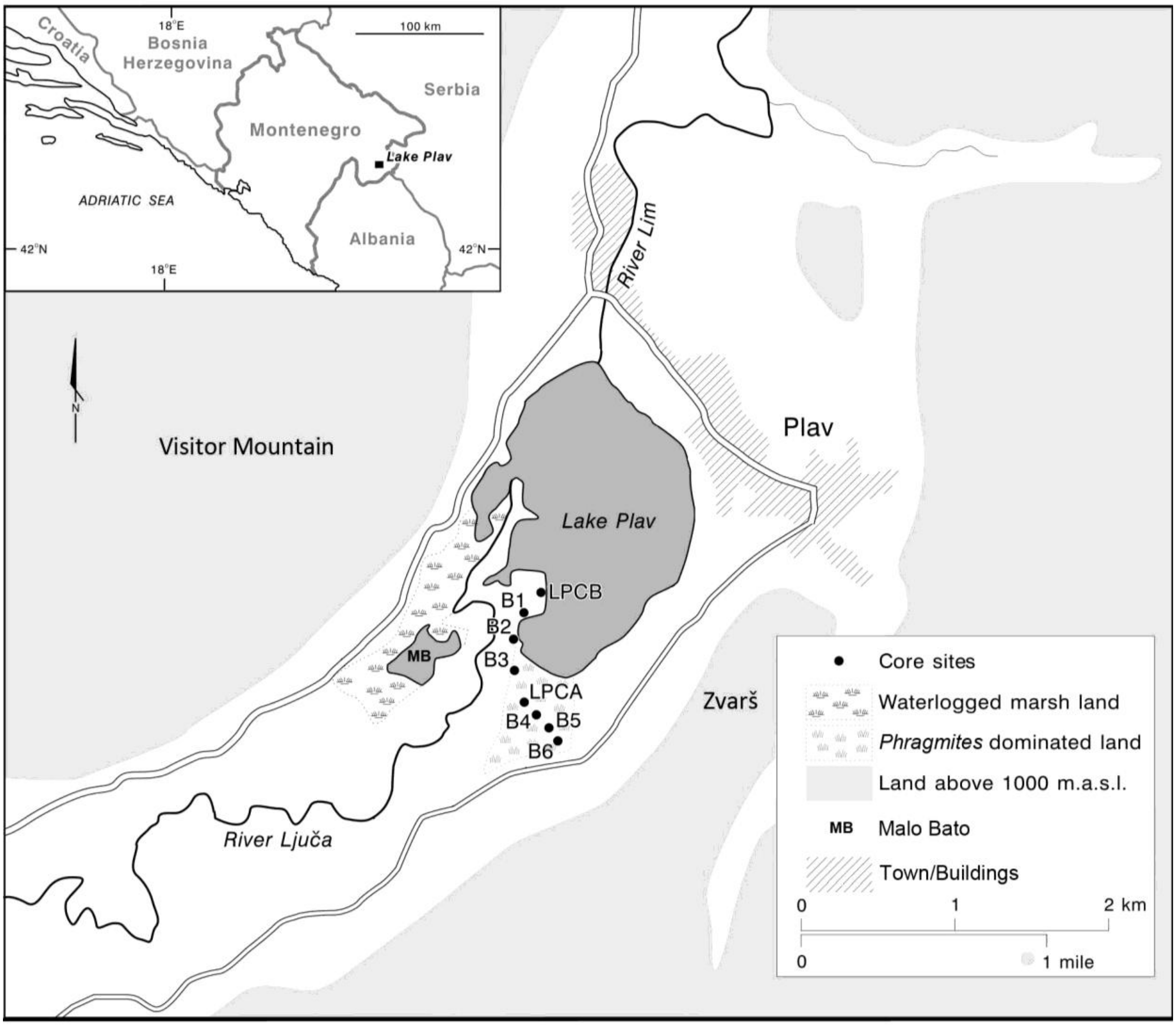


Figure 4.1: Sites of core extraction around Lake Plav.

The second core extracted from Lake Plav was core LPCB (42° 35' 22.84" N, 19° 55' 19.03 "E, Figure 4.1) in July 2008. Currently the site is situated on meadow/grazing land. Extracting this core aimed to produce a sediment record located closer to the centre of the lake than LPCA. The core was sampled by the Montenegro Geological service using a percussion corer with a large 8 m 'A' frame (Figure 4.2, A). The percussion corer was thought to allow greater volumes and depth of sediment extraction, but though a depth of 27.4 m was reached a large 18 m hiatus was encountered between 7.5 - 25.5 m. Prior to the hiatus the sediments at 5 - 7.5 m became very wet and to ensure successful removal of the core a heavy metal sleeve was used to line the extraction hole (Figure 4.2, B). On site sub-sampling of each 0.5 m to 1 m section drilled was completed at 30 mm to

60mm intervals, dependent upon any obvious lithological changes. Samples were then double bagged and placed in cool conditions.

Figure 4.2: Montenegrin Geological Service coring team extracting core LPCB (July, 2008). (A) The 8 metre 'A' frame. (B) Metal sleeve used to extract unconsolidated sediments.

4.2.2 Sediment logging

Sediment descriptions were based on the Tröels-Smith system (Tröels-Smith, 1955; see Appendix III), which assumes that sediments consist of various components and aims to characterise each deposit in a simple list of sediment types. The system used at the Lake Plav catchment was a modified Tröels-Smith produced by Aaby and Berglund (1986). This system reduced the classification types and provided a simplified version of the approach. Lawson *et al.* (2005) used the modified Tröels-Smith approach to log sediments for the Nisi Fen site in Greece. The colour or hue of sediments was described using a Munsell Colour Chart (Munsell soil chart, 2000; Sutherland *et al.*, 2000).

To identify the sediment stratigraphy across the site, a transect of six transect cores were systematically sampled across the Lake Plav basin (Figure 4.1). Each transect core was extracted using a 20 mm gauge auger in 500 mm sections to a depth of 3 – 7 m. Underlying sediments were too compacted and the suction too great to manually core beyond this. The sediment stratigraphy was logged on site using the Tröels-Smith method and a Munsell colour chart, whilst, wood or plant fragments found were sampled for possible dating.

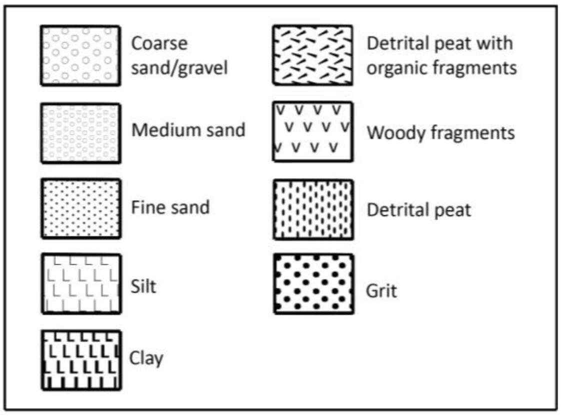


Figure 4.3: Edited Tröels-Smith key identifying the symbols used for different deposits (adapted from Tröels-Smith, 1955).

4.3 CURRENT VEGETATION IN THE LAKE PLAV CATCHMENT

One of the research aims for this project was to suggest past changes in vegetation within the catchment, using pollen analysis to describe the changes that may have occurred. However, there is currently no information describing the vegetation of this catchment or more broadly the Prokletije Mountains. Therefore information on the current vegetation would aid interpretation of the fossil pollen record by providing a modern analogue for the environments.

4.3.1 Vegetation surveying

Following Grieg-Smith (1957), quantitative analysis of vegetation can be split into three categories: 1) an estimation of vegetation composition within certain boundaries, 2) investigation into the vegetational variation within an area, or 3) correlating differences in vegetation with that of one or more habitat factors. To analyse the vegetation two methods of surveying are then available to the researcher; the relevé method and random quadrat method. The relevé method is an estimation of percentage cover of each vegetation type within the defined area (i.e. quadrat), but the random quadrat method provides a little more information. Lepš and Hadincová (1992) suggested that the relevé method is the most efficient surveying method available. Due to the limited time available in the catchment this method was used for the Lake Plav catchment survey. Varying quadrat sizes may introduce bias into results (Greig-Smith, 1957; Power and Cooper, 1995; Galatowitsch and van der Valk, 1996; Huang *et al.*, 2007), thus a 1 m^2 quadrat was used for all sites. The small quadrat size allowed five sites to be assessed within a particular vegetation zone following Galatowitsch and van der Valk (1996). This method aimed to determine the contemporary vegetation types and areas where they were concentrated and did not aim to produce a full vegetation survey of the area. As this would detract from the main aims of the research and transform the study into a modern day ecological assessment of the catchment. Thus, every plant is unlikely to have been accounted for due to: no identification if a plant was not in bloom, catchment size (269 km^2), existence of different microclimates and environments and time constraints.

Vegetation surveying took place in May/June 2007, when many plants are in bloom, to provide more reliable and successful identification of plant types (Polunin, 1980; Blamey and Grey-Wilson, 2004). A visual assessment of the site was undertaken first with vegetation zones identified by eye – later these were mapped using catchment maps and satellite imagery (Gaus-Krigerova, 1970; Huber Verlag/Lokalna turistička organizacija Plav, 2008; Google Earth). Each zone was then systematically surveyed using a quadrat and reference material by Polunin (1969; 1980) and Blamey and Grey-Wilson (2004). Zones were then sub-divided or merged where necessary, and the altitude at which zones changed recorded to enable GIS mapping of the data. Finally, any anthropogenic management of zones was noted.

4.4 BUNI I JEZERCES SITE

This ‘upper catchment’ site provided the necessary components to fulfil the remaining project objectives: (1) collect a high mountain lake core above or at the tree line to provide a predominantly climatically influenced palynological record, (2) map and date the geomorphological features of a high mountain glacier to enable a catchment specific palaeoclimatic reconstruction. The sites used at Buni i Jezerces included Lake C, which provided the site for lake sediment extraction and the Maja e Koljaet glacier which fulfilled the second objective described above (see Figure 3.6, Chapter 3).

4.4.1 Lake C

This lake provided sediment from above the tree-line to be sampled, with the aim of producing a record of vegetation change primarily influenced by climatic change rather than anthropogenic activity.

4.4.2 Sediment coring

The largest lake of Buni i Jezerces is Liqueni Madhe (Abraham, 2007). This was unsuitable for sediment extraction because of; limited water retention, presence of rubble across the surface from rock-falls which may have compressed or disturbed sediments. While evidence of seasonal drying suggested fossil pollen grains in the upper units of the core were likely to have been desiccated. These factors suggested that the stratigraphic order may have been compromised, particularly in the upper units which would be useless for this late Holocene study. The only lake suitable for coring was Lake C (42^{o} 28’ 01.91” N, 19^{o} 48’ 43.0” E; See Figures 3.6, 3.7 Chapter 3) with limited clastic material on the lake bed surface and shallow water present making desiccation and disturbance of the upper sediments less likely. There was evidence of drying and cracking on part of the lake bed, but the site was less affected by this than the surrounding water bodies. Furthermore, as the lake was situated at just above the tree line (1754 m a.s.l.) in this north-east facing valley, it appeared to be suitable for a primarily climatically driven palaeoenvironmental record.

Sediments from Lake C (core BJC1) were extracted using a 20 mm gauge auger, in 500 mm sections following the methods of Aaby and Digerfeldt (1986). Slow sedimentation rates have been shown in similar small high mountain lakes (Stefanova *et al.,* 2006a, b; Tonkov *et al.,* 2002, 2006) and thus core BJC1 was extracted to a depth of 1.45 m.

4.4.3 Maja e Koljaet glacier

As one of the few current glaciers in Prokletije and Southern Europe this glacier would enable a record of glacial advances and recessions to the modern day to be completed. The glacier is located in a north-east facing niche, providing ideal conditions for preservation of accumulated snow (Kruss and Hastenrath, 1987; Chueca and Julian, 2004), at an altitude of 1993 – 2029 m a.s.l. (42^{o} 27’ 12.5” N, 19^{o} 47’ 52.8” E; Figures 3.3, 3.4, Chapter 3). Maja e Kokervhakes and Maja e Jezerces lie to the east and south-east respectively of the glacier, providing further shade for the development and maintenance of a small high mountain glacier (Figure 3.6, Chapter 3).

4.4.4 Glacial geomorphological mapping

Benn and Evans (1998) described glacial landscapes as, ‘consisting of superimposed depositional systems of different ages’ (p. 535). Mapping of these systems/landforms and glaciers has occurred across the Mediterranean since the late 1700s, with geomorphological mapping to assess the past extent of glaciers occurring since the late 1800s (Carbonnières, 1789; Schrader, 1894 cited in Cía *et al.*, 2005). Glacial geomorphological mapping relies upon the correct identification of glacial landforms such as moraines and trim lines (e.g. Sharp, 1985; Benn and Evans, 1998; Ballantyne, 2007) to allow reconstruction of past glacial extents. Features must then be dated using e.g. cosmogenic dating or lichenometry, to provide a chronology for palaeoclimatic data created using degree day modelling (Nesje and Dahl, 2000; see section 4.13). Moraines provide topographic evidence of past glacial extent, but other features of glacial activity should also be mapped, such as dolines and ice moulded bedrock, to ensure full understanding of the glacial processes at a site. Dolines have been described as 'sink holes' and are common features in karstic topographies (Hughes *et al.*, 2006), often located at former glacial margins in the form of doline fields and offer evidence of proglacial meltwater discharge or trapped windblown snow (Hughes *et al.*, 2006; Hughes *et al.*, 2007). Geomorphological mapping in Montenegro began with Cvijić's (1907, 1913, 1917) work, before Auer (2007) and Milivojević *et al.* (2008) continued the study of Pleistocene glaciations around Prokletije. Hughes (2007, 2008, 2009a, b) is currently the only researcher to provide work on more recent, LIA to present, glacial changes in Montenegro around Mount Orjen (112 km west of Lake Plav) and the Durmitor Massif (90 km north west of Lake Plav, Figure 4.4). Studies across Montenegro and beyond rely on mapping glacial features onto 1:10,000 or 1:25,000 scale base maps. Moraine locations, other glacial features and the current glacier extent can then be mapped directly onto the base maps, from which glacial geomorphological maps may be created. Hughes (2009b) used additional measurements of sediment ridges and dolines with an abney level and 50 m measuring tape.

For this research project the glacial features of the Maja e Koljaet glacier were marked on to a 1:10, 000 base map and GPS readings were taken to locate glacial features. Sketch maps and photographs aimed to aid development and interpretation of the resultant map. The glaciers past course was constrained by the local geomorphology, with high ridges and peaks associated with Maja e Bojes, Maja e Kokervhakes and Maja e Ete (Figure 3.6, Chapter 3). Glacial landforms in the area were sometimes obscured by recent debris from the surrounding slopes and this debris was identified by: active source areas above, a lack of weathering on debris material and conflicting lichen sizes when compared to clasts beneath or surrounding debris material.

Figure 4.4: Map identifying areas of previous glacial geomorphological research in Montenegro.

4.4.5 Dating of glacial features

Dating glacial features is vital to constrain the time at which moraines were uncovered by ice and provide an accurate reconstruction of glacial advances and recessions (Cerling and Craig, 1994; Everest and Bradwell, 2003; Hughes, 2009b; Zahno *et al.*, 2009). Two approaches are available for direct dating, cosmogenic isotope dating and lichenometry. However, as cosmogenic dating incurs large errors for very young surfaces (<1000 years), due to low concentrations of nuclides, it was not suitable for application in this study (Gosse and Phillips, 2001). An earlier approach for dating young moraines is lichenometry, which is a method for determining surface age developed by Beschel (1954, 1961). Although the method is non-destructive to the dated surface, it assumes that the surface being dated is no younger than the oldest individual lichen thallus (Matthews, 2005; McCarthy, 2007). Lichen types used in lichenometric studies must grow radially, producing oval/circular thalli, with direct measurement of the radial growth providing a record of age (Beschel, 1954; 1961). The lichen used in a study is dependent on the environment, for example the commonly used *Rhizocarpon geographicum* is rare on calcareous rocks due to an intolerance of calcium, and is thus favoured for use on basaltic rocks (Luckman and Osborn 1979; Chenet *et al.*, 2010). Lichen types tolerant of calcareous substrates include *Aspicilia calcarea* and *Xanthoria elegans*[2], which have been used from the Canadian Rocky Mountains and Scandinavia, and recently in Montenegro (e.g. Osborn and Taylor, 1975; Luckman and Osborn, 1979; Matthews and Shakesby, 1984; Karlén, 1988; McCarthy and Smith, 1995; Winkler, 2003; Hughes, 2009b). Lichen used in lichenometric dating must exist on rocks away from vegetation cover or depressions where snow-cover may persist, as these conditions hinder lichen growth providing unreliable growth data (Denton and Karlén, 1973; McCarthy, 2003; Matthews, 2005). To ensure robust lichenometric data, surface ages should be carefully interpreted as lag time for colonization of a surface by lichen (ecesis), is generally 10 - 25 yrs, but may reach 100 yrs. Ecesis affects surface age estimates and thus limits the age range at which lichenometry can be used (Armstrong, 2004; McCarthy, 2007). *Xanthoria elegans* has an ecesis of between 10 - 30 yrs (McCarthy and Smith, 1995) and *Aspicilia* type between 10 - 20 yrs (Miller, 1969; Hughes, 2009b). Hughes (2009b) highlighted the problem of lichen ecesis, with a ridge below Minin Bogaz in Durmitor Massif lacking lichen, suggesting the ridge was formed ***c.***10 to 20 yrs prior (Hughes, 2009b). Growth of the thallus is rapid at the beginning of colonization (first few decades), becoming steadier before it declines beyond a maximum of around 700 years (Matthews, 2005). *Aspicilia calcarea* has an average annual radial growth of around 1.1mm yr^{-1} (Armstrong and Bradwell, 2010). Lichen growth rates will differ slightly as a result of their position, thus the average of several lichen must be calculated to account for this. Surface ages may only be determined once an accurate lichen growth curve has been established. This is a numerical relationship between lichen size and surface age, thus surfaces of known age are required for its development (e.g. gravestones, dated buildings). Once completed, a lichen may be measured and their size used to assess the age of a surface (McCarthy and Smith, 1995). Lichenometry is an inexpensive, non-destructive and relatively efficient surface dating tool compared to cosmogenic dating and provides the most reliable

---

[2] *Xanthoria elegans* was one of the first lichen used by Beschel (1954) in lichenometric studies.

method for dating from the LIA to present (Miller, 1969; Andersen and Sollid, 1971; Mottershead and White, 1972; Matthews and Shakesby, 1984).

Dated monuments and gravestones are usually an ideal location for the growth of lichen and thus a source of lichen data for producing a lichen growth curve. Gravestones in the Lake Plav catchment were dominated by polished granite gravestones, which are not conducive to lichen growth. Coupled with the limited number of other dated memorials, the construction of a statistically robust local lichen growth curve was not permitted. Therefore, the growth curve created by Hughes (2009b) for *Aspicilia calcarea* around Durmitor was used as it was the closest data set available (Figure 4.5).

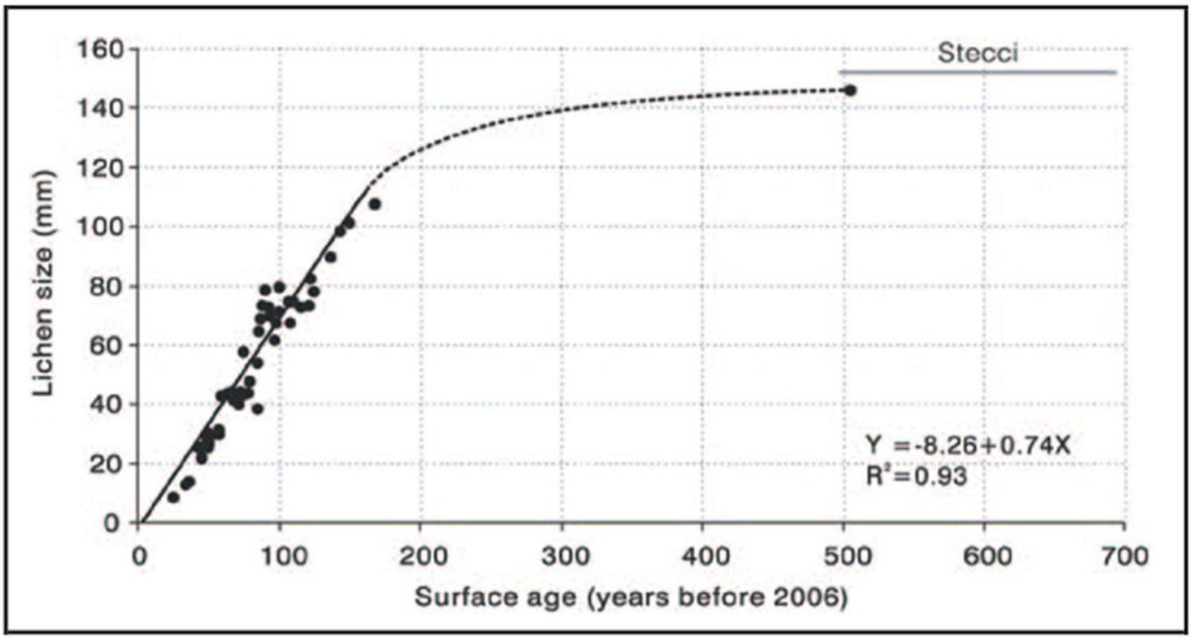


Figure 4.5: Lichen growth curve developed for Durmitor using lichen situated on a variety of surfaces (taken from Hughes, 2009b).

The equation associated with the Figure 4.5 lichen growth curve:

$$Y = -8.26 + 0.74X$$

Where: ***Y*** is the mean size of lichen (mm) and ***X*** years before 2006.

At the Maja e Koljaet glacier it was expected that the largest 30 thalli of *Aspicilia calcarea* type (Figure 4.6) could be sampled on moraine crests to avoid depressions and vegetation cover, but the number of available lichen was limited. This is a common problem on limestone, as the carbonate rich rocks produce sparse colonization of lichen, possibly exacerbated by high weathering rates and persistent snow cover. Therefore, 5 - 25 lichen were sampled at all but one glacial feature, enabling the mean of the five largest to be calculated for most features and used against the lichen growth curve developed by Hughes (2009b: Figure 4.5).

Figure 4.6: An *Aspicilia calcarea* lichen upon a boulder at the Maja e Koljaet site.

4.5 LABORATORY METHODS

Palaeoenvironmental records throughout the Balkans, Mediterranean and the rest of the world have continually relied on multi-proxy analyses to ensure that an accurate record of environmental, climatic and anthropogenic change is produced (e.g. Wijmstra, 1969; Guiot *et al.*, 1989; Prentice *et al.*, 1996; Denefle *et al.*, 2000; Smol, 2002; Birks and Briks, 2006; Ariztegui *et al.*, 2010; Fouache *et al.*, 2010). The following sections (4.6 to 4.11) provide a brief review of each method and outline why the methods were used during this study.

4.6 POLLEN ANALYSIS

Pollen analysis was developed in 1916 by von Post and Lagerheim, applying the technique to Quaternary lake sediments and peat deposits (von Post, 1916 cited in West, 1964). The method has been widely used in a variety of subjects from palaeoenvironmental studies (palaeopalynology) to the analysis of pollen in honey (melitopalynology) or animal dung (copropalynology) and in criminological studies (Erdtman, 1969; Moore and Webb, 1978). The identification of gymnosperms, angiosperms, spores and aquatics is used to produce a pollen stratigraphy relating to trees, shrubs, herbs, aquatics and spores. Extensive use of the method is due to the excellent preservation and abundance of pollen grains in sediments and the structural and sculpting features of the grains that makes them highly recognizable and identifiable to species level (Moore and Webb, 1978). Furthermore, fossil pollen records provide a statistical representation of past vegetation in an area at a particular time, allowing reconstruction of the temporal and spatial vegetational changes at a site and, spatial and temporal comparison of results. These changes can then be linked to climatic and anthropogenic impacts on a system over time (e.g. Lowe and Walker, 1984; Tzedakis, 1993; Oldfield *et al.*, 2003; van der Linden *et al.*, 2008).

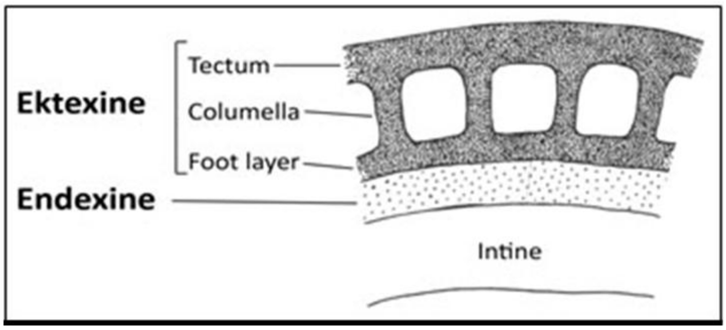


Figure 4.7: The structure of a pollen grain (adapted from Moore and Webb, 1978).

Pollen is produced in varying quantities, with 71,000 grains found upon a single *Typha latifolia* inflorescence (Smirnov, 1964). Wind pollinated species generally produce greater volumes of pollen than insect pollinated varieties. Further types include cleistogamous species which are a non-opening/self-pollinating types, entomophilous types which are pollinated by insects and anemophilous species which are wind pollinated types (Birks and Birks, 1980). Estimations of pollen production were first made by Pohl (1937) and others since have provided information on pollen production and concentrations of certain types. These studies have defined *Pinus*, *Alnus*, *Betula* and *Fagus* as high pollen producers, whilst low pollen producers include *Ilex*, *Tilia* and *Hedera* (e.g. Andersen, 1974; Middledorp, 1982 cited in Bohncke and Wijmstra, 1988; Lau and Stephenson, 1994; Bottema, 2001; Fernández-Illescas *et al.*, 2010). The high production rate of pollen grains and small size leads to an abundance of grains within sediments. Researchers such as Smirnov (1964) suggested 6kg/hectare per year of pollen grains were deposited at the Rybinsk Reservoir in Russia, and Pohl (1937 cited in Smirnov, 1964) who suggested that within a hectare of forest, up to a billion pollen grains can be released by angiosperms. Pollen production is not the only factor affecting the abundance to which grains are deposited, but their dispersal is an important aspect to include and as they are small (5 µm - ~120 µm) pollen grains can be transported via wind, dispersed in the atmosphere and deposited as pollen rain. Various models for pollen dispersion have been created, such as Tauber's dispersion model for forested areas, Sutton's model for particle dispersal from ground level (Sutton, 1947; Tauber, 1965, 1967) and more recent models that have enabled hypothesis based reconstructions of a landscape by Prentice (1985) and Sugita (1993). The reliability of each model has been questioned due to the variety of factors that may impact on how far pollen grains may travel, with varying weather conditions aiding/hindering dispersion, the height at which pollen is released and whether an area is in leaf determining how easily pollen grains are dispersed (Holmes and Bassett, 1963; Tauber, 1965; Birks and Birks, 1980; Jackson, 1990). Pollen sources have ranged from as little as 50 m to hundreds of kilometres (e.g. Perry, 1978; Jackson, 1990; Punning and Koff, 1997; Koff *et al.*, 2000; Sjögren *et al.*, 2008), but further consideration must be made to stream-borne pollen, especially in a lake system. Stream-borne sources include: fall from stream-side plants (e.g. *Salix, Alnus*), catchment or bank erosion events and local surface run-off, which has been described by Birks and Birks (1980) as a major source of pollen. Unaccountable sharp rises in certain stream-side plants may thus be indicative of a palaeoflood, depending on indications from other independent data sources. The influence of catchment wide scale events such as floods versus changes within the lake environment requires unravelling, thus pollen records from lake sediments must be carefully interpreted to ensure changes are deciphered as accurately as possible. By using a multi-proxy approach these variations may be more easily understood as interpretations may be supported by alterations in other proxy records. Although pollen assemblages are complex in nature and can represent varying spatial scales, this has not hindered the application of the method in a variety of studies across the world in many different settings (e.g. Bartlein *et al.*, 1984; Follieri *et al.*, 1989; Bonnefille and Mohammed, 1994; Björkman *et al.*, 2003; Pensa *et al.*, 2005; Fouache *et al.*, 2010).

Internal factors within sediments, do not affect pollen deposition, but can cause post-depositional re-distribution of pollen (Moore and Webb, 1978; Birks and Birks, 1980). The primary cause of disturbance in lacustrine sediments is by detritus feeding animals mixing the surface layers, but unlike peat there is no down-washing of pollen (Moore and Webb, 1978). Accurate dating of the upper sediments can provide evidence of biotic disturbance, with divergence of a caesium-137 (^{137}Cs) profile suggesting possible physical and biological mixing (Longmore *et al.*, 1983, 1986). This must not down-play the valuable fossil pollen data lacustrine sediments can provide, as the majority of lake material is allochthonous the spatial scale that the data represents will always be representative of catchment wide changes to a degree (e.g. Moore and Webb, 1978; Edwards and Whittington, 2001; Vannière *et al.*, 2003; Chiverrell *et al.*, 2008). Furthermore, with accurate dating of the sediments, the possibility of pollen re-distribution can be identified.

4.6.1 Pollen grains, identification and indicators

The preservation of pollen grains within sediments owes to their resistance to decay in non-oxidising situations, as each grain is composed of a central living section (including the intine) and an outer section, exine (Figure 4.7). The cellulose intine decomposes rapidly on deposition/fossilisation. In contrast the exine consists of sporopollenin, an inert substance more resistant to decomposition and chemical action (Zetsche, 1932; Lowe and Walker, 1984). These exine traits have been exploited during the preparation process of pollen samples. An acetolysis step to remove organic material is included, but as the pollen exine is resistant to acetolysis samples are free from other organics. Research has shown that pollen grains with higher sporopollenin contents are more resistant to other chemical actions such as oxidation e.g. Lycopodium (23.4%), *Tilia* (14.9%) and *Ulmus* (7.5%; Kwiatkowski and Lubliner-Mianowska, 1957). Ensuring that the exine remains intact following sampling is the most important aspect of pollen analysis, as it includes the structural and sculpting features making pollen identification possible (Figure 4.7). These morphological and structural features include apertures and surface pattern. Apertures are described as, 'any thin or missing part of the exine which is independent of the pattern of the exine' by Moore and Webb (1978) and can take the form of colpus and/or porus. Each single grain can have upwards of three furrows (e.g. *Salix*) and one pore (e.g. *Poaceae*). Surface patterns must be identified to ensure accurate results, which describe the morphology of a grain. The serebrate pattern is indicative of the *Ulmus* (Elm) grain and striate suggests *Saxifraga* (Saxifrage). The size of grains is also considered during identification as cereal types (e.g. *Avena*) are around 40 µm to 45 µm (Moore and Webb, 1978; Lowe and Walker, 1984; Joly *et al.*, 2007). Certain grains are released as tetrads (e.g. Ericaceae) making them obvious during identification (Birks and Birks, 1980). It must be mentioned that pollen grains are not always in 'mint' condition and may be crumpled, corroded, broken or degraded making identification difficult (Cushing, 1967). Grains laid down in river clays are more likely to be severely corroded (with etching/pitting) and in silty lake sediments the morphology of grains may be degraded, but major structural features will remain recognizable (Havinga, 1967). The site of deposition also plays a part in the pollen record retrieved. Corroded grains found in Pleistocene ice-pushed moraine were attributed to the formation of the moraine (Polack *et al.*, 1962). Wilmhurst and McGlone (2005a, b) have suggested corroded grains in lake sediments are associated with catchment disturbances. However, the effect of in-washed eroded soils and littoral sediments can produce pollen data misrepresented by re-working and thus a distorted signal of catchment erosion or disturbance events (Wilmhurst and McGlone, 2005b).

1) Cereal type:

- Long axis diameter > 40 µm
- Mean grain size (average of long and short axes) > 37 µm
- Annulus diameter must be twice the pore diameter and ≥ 8 µm
- Protruding pore* with diameter > 4 µm

2) cf. cereal type:

- Long axis diameter ≥ 40 µm
- Mean grain size (average of long and short axes) > 37 µm
- Annulus diameter must be twice the pore diameter and > 8 µm
- Protruding pore* with diameter ≥ 3.5 – 3.9 µm

3) Possible cereal type:

- Long axis diameter ≥ 37 µm
- Annulus diameter > 8 µm

4) Wild grass (Poaceae):

- Long axis diameter ≥ 37 µm
- Annulus diameter < 8 µm

Figure 4.8: Categories defined by Edwards *et al.* (2005) for classification of cereal types. Indicates the additional feature used to identify cereal types as stated by Leroyer (1997).

Indicator grains have played a vital part since palaeopalynological research began as various species can be indicative of particular changes (e.g. West, 1964; Moore and Webb, 1978; Behre, 1986). This provides a large database of information to aid the interpretation of pollen sequences, with three different categories of indicator species available:

(1) Indicators of anthropogenic influence
(2) Thermophilous indicators
(3) Climatic indicators of elevated wet/dry periods

Anthropogenic indicators rely on the accurate classification of arable and pastoral grains.

Arable indicators are recognised as cultivated grasses i.e. cereal types, which are morphologically dissimilar to other monoporate grains i.e. wild grasses. Wild grasses are less than 37 µm in long axis diameter, whereas the size threshold for cereal types is at least 37 µm (threshold now 45 µm), with an annulus diameter of >8 µm (Edwards and Hirons, 1984; Edwards *et al.*, 2005; Joly *et al.*, 2007; Joly and Visset, 2009). Beug (1961) summarised the different cereal types, providing a key to determine the species found, whilst Edwards *et al.* (2005) provided four categories for the differentiation of wild grass and cereal types (Figure 4.8). The categories defined by Edwards *et al.* (2005) provide a good base for cereal and wild grass differentiation, but Joly *et al.* (2007) state that for a pollen grain to be identified as cereal type the pore must be protruding and the exine thick following research by Leroyer (1997; Figure 4.8, 4.9).

Cereal types can be used alongside other weed pollen, associated with ground disturbance, to provide evidence of arable farming with the concentration of grains employed to assess the extent of cultivation within an area (e.g. Hibbert, 1978; Jansenn, 1981; van den Brink and Janssen, 1985; Bakels, 1991; Bottema, 1992; Poska *et al.*, 2004; Joly and Visset, 2009). Table 4.1 highlights the different pollen species and types that can be used to infer varying changes within a palaeopalynological record. Forest clearance by means of a reduction in total arboreal pollen, provides another sign of anthropogenic activity and the type *Epilobium augustifolia* (Rosebay willowherb) has been used as an indicator of fire during forest clearance (Schimmel and Granstrom, 1996). Climatic inferences can similarly be made with particular thermophilous/mesothermic pollen types (Table 4.1). Arboreal pollen types such as *Corylus, Ulmus*, *Tilia*, *Carpinus/Ostrya* and *Juglans* are less tolerant to cold climates (Belis *et al.*, 2008; Jiménez-Moreno *et al.*, 2010). Čarni *et al.* (2009) mapped the extent of various thermophilous deciduous woodland types in Southern Europe. Durmitor massif in Montenegro (Figure 4.4) was included, which was indicative of higher altitudes and included Balkan endemic types: *Campanula*, *Lathyrus*, *Trifolium* and *Veronica* (Čarni *et al.*, 2009). The '*Quercion petraceae-cerris*' and '*Carpinion orientalis*' groups were recorded on the Albanian-Serbian border, with the latter dominated by *Quercus pubescens* in the tree layer and was found along the Adriatic coast or where the maritime influence extended inland (Čarni *et al.*, 2009). The Lake Plav catchment may well fit into the *Carpinion orientalis* group, but as *Fagus* and *Pinus nigra* (both dominant within the catchment) were not used in the research by Čarni *et al.* (2009), as they were seen as mesophilous climazonal, this is difficult confirm. *Pinus nigra* is a species that can tolerant moderate soil moisture and nutrients and has been actively planted across Karst environments since the 1850s making its presence an indicator of anthropogenic change rather than climatic variation (Kranjc, 2009).

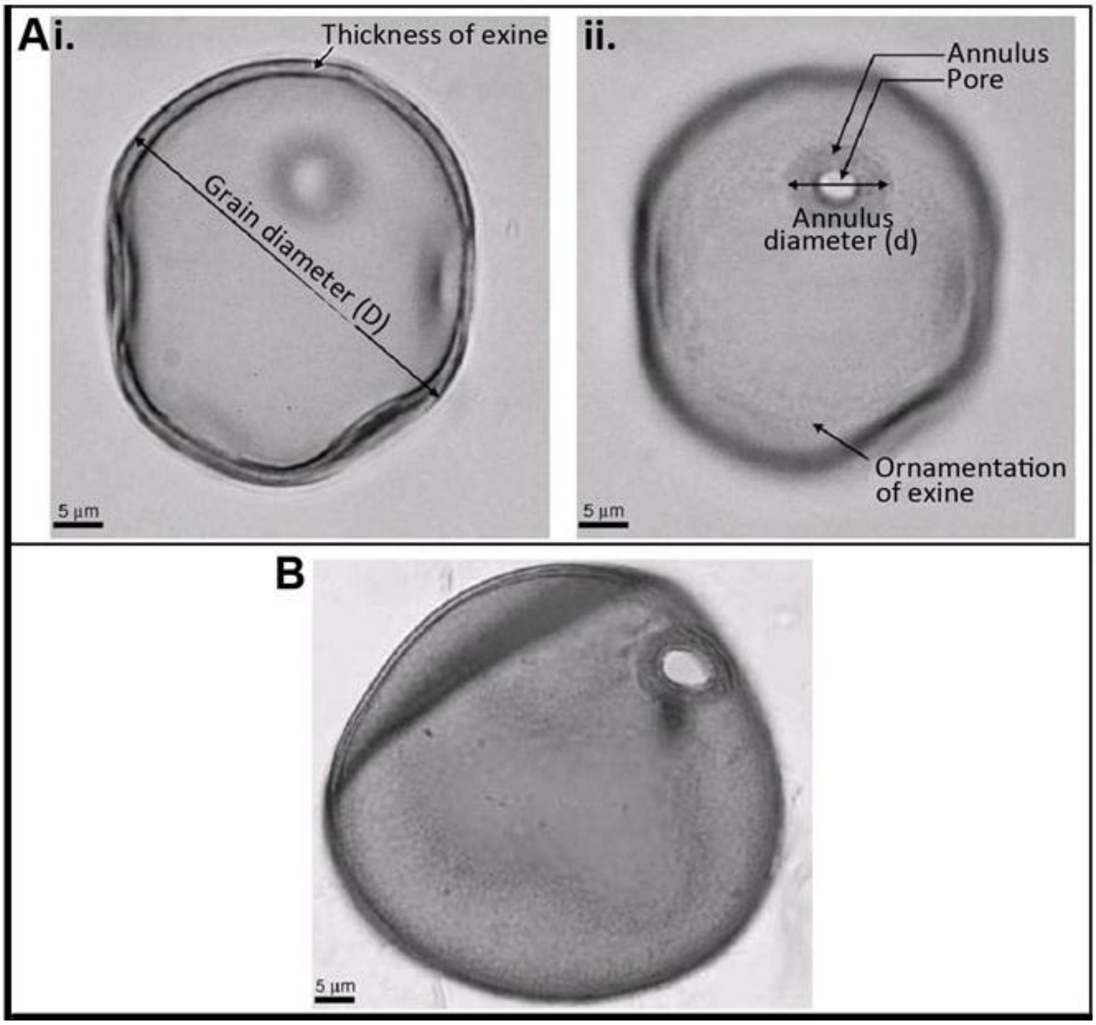


Figure 4.9: (A) i. and ii. Morphology of a wild grass pollen (*Bromus hordeaceus subsp. hordeaceus*) including suggested positions for identification measurements, and (B) cereal type (*Triticum spelta*) with measurements of: grain diameter 61 µm, pore size 6.6 µm, annulus diameter 15 µm (Adapted from Joly *et al.*, 2007).

Activity or change indicated	Examples of indicator pollen type/species	References
Arable farming	*Secale cerale*; Cerealia type; Triticum; Oleaceae; Vitis; Castanea; *Juglans regia*; *Vicia*; *Chenopodium*; Brassica; *Plantago major/media*; *Polygonum*	Sergerstrom, 1991; Bottema, 1992; Mighall *et al.*, 2006; Deforce, 2010
Pastoral farming	*Plantago lanceolata*; *Campanula*; *Trifolium repens*; *Viola*; *Rumex*; Ranunculaceae; Lactuceae	Behre, 1981; Jassen, 1981;
Climate	*Corylus; Ulmus; Tilia; Ostrya/Carpinus;* Juglans*; Hedera; Ilex; Viscum; Sedum*	Iversen, 1944; Moore and Webb, 1978; Belis *et al.*, 2008
Terrestrial to swamp/aquatic environment	**(ordered: wet/swamp land to species tolerant to 1-2 feet standing water)** *Equisetum; Cyperaceae; Typha latifolia*	Polunin, 1969; Grace, 1989; Kercher and Zedler, 2004; Mountford *et al.*, 2006; Theocharopoulos *et al.*, 2006

Table 4.1: Herb pollen indicators used in European palaeoenvironmental studies for different activities and changes (various sources).

4.6.2 Pollen extraction

Removal of pollen grains from lake sediments has been achieved in a variety of ways with Kidson and Williams (1969) developing a sodium pyrophosphate suspension method and Faegri and Iversen (1975) a hydrofluoric acid (HF) method. The HF preparation is one of the most widely used as it can be used to remove silicates and other minerogenic material, but still retain pollen grains because of their resistance to chemical action (Faegri and Iversen, 1975). Additions to the preparation have included sieving steps and use of a preliminary $Na_4P_2O_7$ wash for clay or slit-rich sediment (Kidson and Williams, 1969 method, see Appendix IV). Cwynar *et al.* (1979) suggested that without sieving, clay and silico-fluoride precipitates would hide 'critical features'. More recently heavy liquid separation using zinc chloride ($ZnCl_2$) or sodium polytungstate ($Na_6(H_2W_{12}O_{40})$) has been developed (Gregory and Johnston, 1987; Zabenskie *et al.*, 2006). The latter method, first discussed in the mid-1990s for application in palynological research, uses the specific gravities of minerogenic particles against that of pollen grains. Sporopollenin contained within each pollen grain has a specific gravity of ~1.45g/cm^3, whereas minerogenic material has a greater specific gravity. By using the reagent sodium polytungstate (SPT) at a specific gravity of between 1.95g/cm^3 and 2.1g/cm^3 the pollen will 'float off' during preparation, whilst the minerogenic material will sink and be discarded. SPT is less dangerous than HF (Meyer and Gross, 1985) and is thus a safer alternative to the Faegri and Iversen (1975) technique. 20 samples from core LPCA were used to trial this method, but the samples processed were not of the quality expected. Counting of these slides would have been overly time-consuming and the accuracy of grain identification would have been severely hindered. Therefore, the traditional Faegri and Iversen (1975) method was used, but due to the high minerogenic content within the sediments three treatments of HF were applied to ensure the production of 'cleaner' samples.

Pollen was extracted from all three cores, LPCA, LPCB and BJC1, with varying degrees of success. Samples every 20 cm through the length of core LPCB were taken, but they were generally devoid of pollen and further sampling was not pursued. Preparation of samples from core LPCA were taken at 10 cm intervals throughout the length of the core. As BJC1 was shorter in length, samples were taken contiguously for the first 30 cm and then every 5 cm. The samples were measured by volume displacement following the standard Faegri and Iversen (1975) method (Appendix V), modified by exposing samples to three HF treatments (at least one hot treatment) and with the additions of a Lycopodium exotic marker grain (13,400 spores/tablet).

4.6.3 Analysis of pollen samples

Birks and Birks (1980) suggest that between 300 and 500 grains should be counted, excluding aquatics and spores, before the results offer statistically robust data. Some studies have used fewer than 200 grains in samples with low pollen concentrations

(Tzedakis, 1993; Karayigit *et al.*, 1999), providing less taxonomically diverse samples which can provide misleading results (Birks & Line, 1992). Poor pollen preservation in particular stratigraphic units will reduce the reliability of data within that sample, with the percentage of indeterminable grains generally relating to reliability of data for that unit (Cushing, 1967). Quantities of grains found in samples have also been related to the depositional environment in Mediterranean palaeoecological studies, where Tzedakis (1993) noted that glacial-age sediments had lower pollen concentrations, around 200 to 300 pollen grains per sample, but other sediments sampled resulted in a maximum of 1,033 pollen grains being counted from one sample. Generally more than 500 grains (excluding aquatics and spores) were counted for LPCA samples and between 500 and 1500 for the BJC1 samples.

Pollen concentrations were established using marker grains from Lycopodium spore tablets with classification and identification of pollen taxa throughout this research based on type slides and the texts of Reille (1992), Moore *et al.* (1991) and Faegri and Iversen (1975). While, Joosten and Klerk (2002) found that the keys of Faegri and Iversen (1975) and Moore *et al.* (1991) contain errors, the application of these keys was minimal as Reille (1992) offered a more comprehensive collection of pollen grains found in the Mediterranean region.

4.6.4 Displaying the palynological record

Within this project calculations and preparation of pollen diagrams were completed using the programmes Tilia, Tilia. Graph and TGview developed by Grimm (1987, 1991). Pollen assemblages provide a reflection of vegetation development in an area over time, but depending on the influences affecting the area different pollen sums have been suggested to reduce statistical errors within the sum. Standard pollen sums were used for the Lake Plav catchment; total arboreal pollen, shrub pollen, herb pollen, aquatics and spores and total land pollen sum (which excludes the aquatic and spore fraction; Craigie and Sugden, 2008). Once diagrams were created, pollen zonation exposed periods of significant change. Visual examination of the pollen profile is the most common form of assigning pollen zones and is the first necessary analysis of the pollen profile, but other methods of quantitatively delineating zones are available (Moore and Webb, 1978; Birks and Birks, 1980). Dale and Walker (1970) suggested the use of a dissimilarity coefficient, but this failed to maintain the stratigraphic order of the profile and clustered samples that were stratigraphically separated. CONISS is a program that quantitatively defines similar adjacent stratigraphic zones within a profile and was developed by Grimm (1987) to be used in conjunction with Tilia and its associated files. The data can be transformed in three different ways. Following the Edwards and Cavalli-Sforza chord distance, a square root transformation gives less weight to more abundant types which may otherwise dominate the pollen sequence. Prior to running CONISS on the desired Tilia file (.til file), pollen types with an abundance <2% are discarded in the .til file as their inclusion would give too great an influence to rare types which may relate to the effects of pollen rain rather than catchment wide change. Generally types not included within the total land pollen sum i.e. aquatics and spores are excluded, but it was felt these may represent a history of lake level change and should be included within the first CONISS run on the LPCA data. The program was run again for LPCA and BJC1 excluding aquatics and spores to ensure that any changes in terrestrial pollen types were not overlooked during the period of greatest change in aquatic/spore taxa.

1) What taxa are present?
2) What were the past relative abundances of the taxa present?
3) What plant communities were present?
4) What space did each community occupy?
5) What time did each community occupy?
6) What were the other factors operating in the eco-system in that time and space?

Figure 4.10: 6-step query of pollen data developed by Birks and Birks (1980).

4.6.5 Palaeoecological reconstructions using palynological data

Birks and Birks (1980) proposed a 6-step query of pollen data once a pollen profile has been completed to identify and reconstruct the palaeoecology of an area (Figure 4.10), but there are limitations to answering each of these questions. Varying pollen production could lead to some 'low producing' pollen types being overlooked, especially when the size of the catchment is far greater than the discrete sample taken - $269km^2$ versus $1cm^3$. Furthermore, the identification of pollen taxa to species level is not always possible and only the genera can be achieved (Birks and Birks, 1980). Therefore, the researcher must not assume that the vegetation suggested by a pollen profile is an exhaustive list of the taxa found in the natural environment. The second question can be answered by using correction factors as discussed by Iversen (1944) and Anderson (1970), but if correction factors are only available for a limited number of taxa this will provide an irregular analysis of results. Therefore, the application of the CONISS program seemed most appropriate as the pollen profile is not interpreted on the basis of dominant species alone (Grimm, 1987). This leads to questions 3 and 4 which can be completed by assessing the local or regional origin/ similar ecology of taxa and using the relative abundance of these categories to suggest the extent of a plant community (Berglund and Ralska-Jasiewiczowa, 1986). Pollen influx can further support suggestions of the space that a community occupied, by direct or indirect methods. The former relying upon calculation of total pollen in a sample, whilst indirect methods rely on determination of pollen concentrations using marker grains (Birks and Birks, 1980). Pollen influx diagrams require calculation of sediment accumulation rates from a series of dates plotted against depth (Berglund and Ralska-Jasiewiczowa, 1986). An independent chronology is required to assess the time over which a plant community occupies a particular site and vegetation succession that may occur (see section 4.11). Finally question 6 requires a multi-proxy resource to explore the different possible contributing factors to change within a pollen record. Palynological reconstructions may provide a history of lake level fluctuations as a change to more terrestrial pollen taxa may suggest progradation, whilst alteration to greater dominance of aquatic species may suggest greater water depth or a possible increase in flood incidence (Lundqvist, 1925, 1927; Digerfeldt, 1986; Dearing, 1997). Changes in pollen concentrations have been used as a direct link to the past vegetation in an area by Oldfield (1996) and Williams *et al.* (1996). Oldfield (1996) proposed that dilution of pollen concentration can suggest erosion within a catchment, but this must be verified within other records before it can be confirmed. Both Oldfield (1996) and Williams *et al.* (1996)

indicated that although a pollen profile can provide a useful data set on past environmental change, it is vital to use a multi-proxy approach to ensure accurate interpretation of any palynological variations (Birks and Birks, 2006).

4.7 OSTRACODA ANALYSIS

Ostracoda are small bi-valve crustaceans that live in saline and freshwaters and their fossils have a temporal range of 500 million years. Adult and juvenile ostracoda co-habit and each secretes a calcite bi-valved carapace that encloses the body and appendages. Upon death the body and appendages decompose by bacterial action, resulting in the separation of the two valves (Horne *et al.*, 2002; De Deckker, 2002; Boomer *et al.*, 2003). The current ecological requirements of particular ostracoda can be used to infer lake development and past changes in lake level, lake environment (chemical composition and acidity/alkalinity), depositional environment, sedimentation rate and the palaeoclimate (Bridgewater *et al.*, 1999; Mourguiart and Montenegro, 2002; Ito, 2002; Mischke and Holmes, 2008). Modern analogues for the depositional environment indicate that rates of sedimentation can be inferred by the condition and proportion of disarticulated carapaces (Oertli, 1971), as rapid burial of a carapace is likely to result in no disarticulation and less dissolution (De Deckker, 2002; Horne *et al.*, 2002). The age structure of fossil assemblages can be used as a tool for assessing the palaeoenvironment of a site, for example a lack of juveniles is indicative of reworking as the heavier adult carapaces are left (e.g. Pokorny, 1965; Whatley, 1988; Boomer *et al.*, 2003; Ruiz *et al.*, 2003). Lake development at freshwater sites can be inferred by species such as *Candoninae* and *Cytheroidea* which are both endemic to the Eastern Mediterranean and are characteristic of ancient lake basins (Forester, 1991; Meisch, 2000). Juvenile *Candona* sp. are indicative of shallow littoral habitats, and *Limnocythere inopinata* suggests shallow limnic environments (Frogley *et al.*, 2001). Ostracoda have been used to assess lake development across the Mediterranean at sites such as Ioannina in Greece (Lawson *et al.*, 2004) and Lago di Albano in Italy (Belis *et al.*, 1999). Palaeoclimate inferences can be made from reconstructed palaeo-productivity of water bodies, by the use of a Mutual

Temperature Range method[3] and analysis of the $\delta^{18}O$ oxygen isotope composition of valves, but a high number of carapaces are required (e.g. Belis *et al.*, 1999; Bridgwater *et al.*, 1999; Ito, 2002; Reed *et al.*, 2008).

Ostracoda data analysis generally works on the assumption that counts of > 350 valves per sample provide statistically significant and reliable results (Bridgwater *et al.*, 1999; Griffiths and Holmes, 2000). Throughout palaeoecological work many studies have found that preservation is far lower than this and valves may even be absent (e.g. Belis *et al.*, 1999; Frogley *et al.*, 2001; Ruiz *et al.*, 2003). Poor counts have been attributed to factors such as, low concentrations of oxygen and calcite ($CaCO_3$), changes in sedimentation rate and grain size, the effect of anthropogenic activity on lakes and an abrupt shift in lake conditions during the life cycle of the ostracoda, restricting its progression into subsequent stages (De Deckker, 1988; Belis *et al.*, 1999; Griffiths *et al.*, 2002; Ruiz *et al.*, 2003). Current research still seems to lack any real understanding of the exact factors that are affecting a particular system to cause a reduction/absence of ostracoda in sediment records. Different processes during and after deposition of carapaces can affect the level of identification and the fossil record found. If the reticulate of the valves is changed or the area around sieve pores thinned, valves become more susceptible to breakages. Following burial of a valve, diagenesis may occur leaving a grainy or chalky appearance which can hinder identification to species level (Oertli, 1971; De Deckker, 1988; 2002; Reeves *et al.*, 2007). Valves are also affected by predators, as live ostracoda can be eaten and the shells crushed as they pass through the guts of fish and water birds (Kornicker and Sohn, 1971; Horne *et al.*, 2002). Finally, the different life stages of ostracoda are notoriously difficult to distinguish and identification of 'congeneric' juveniles has been noted as a problem throughout studies, with many studies grouping types together (Griffiths *et al.*, 2002; Reed *et al.*, 2008).

The most common preparation of ostracoda samples is sediment removal followed by wet sieving to collect different size fractions of ostracoda. Sample sizes work on around

[3] The Mutual Temperature Range method was developed by Horne and uses the Non-marine Ostracoda Distribution in Europe (NODE) database. NODE provides the different conditions in which specific ostracoda can survive and a modern climate dataset to infer past climatic changes for a site (Horne, 2007).

0.5 cm to 1 cm thick samples (3 g to 8 g) as this generally yields high enough concentrations of ostracoda carapaces (e.g. Rutka and Eyles, 1989; Bridgwater *et al.* 1999; Griffiths and Holmes, 2000; Reeves *et al.*, 2007; Reed *et al.*, 2008). However, sample sizes have been quoted as large as 100 - 250 g of wet sediment (Kilenyi, 1969; Ruiz *et al.*, 2003). Breakdown of the sample sediment to allow removal of valves can be done by a variety of methods from drying/rehydration, freeze-thaw and freeze drying to the use of hydrogen peroxide (H_2O_2) or calgon ($(NaPO_3)_6$) to disaggregate the material. Many researchers prefer the simple physical cleaning of the valves using high-purity de-ionised water and a fine paint brush, as this is less likely to affect the shell chemistry if trace element work is to be completed on the shells (Holmes, 1996). Griffith and Holmes (2000) suggest employing the method of Smith (1993) with a stack of three sieves at 850 µm, 150 µm and 63 µm, but many prefer to substitute the 850 µm sieve for a larger 2 mm sieve (e.g. Rutka and Eyles, 1989; Vannier and Abe, 1995; Crasquin-Soleau and Kershaw, 2005; Hiruta *et al.*, 2007). The adult carapaces are caught in the larger sieves, while the small sieves capture juveniles separately to provide an accurate representation of the population age structure of the assemblage (e.g. Whatley, 1988; Delorme, 1989; Smith, 1993).

Cores LPCA and LPCB were analysed for ostracoda during this study, but the high proportion of coarse grained material in LPCB is thought to have resulted in poor preservation of valves. Samples were taken at 20 cm intervals throughout LPCA. The simple extraction technique was used (Horne *pers. comm.*), using around 30 g sediment and high purification de-ionised water before gently heating samples for 4 hours to disaggregate the material. Samples were then passed through three sieves; 2 mm, 150 µm and 63 µm and the sieve residues kept for analysis. Ostracoda valves were then identified following the work of Meisch (2000) and Griffiths and Holmes (2000).

4.8 MINERAL MAGNETIC ANALYSIS

Mineral magnetic measurements are rapid, simple, non-destructive and can be used on a variety of materials, from lacustrine sediments to loess deposits (e.g. Thompson *et al.*, 1975; Edwards and Rowntree, 1980; Heller and Tungsheng, 1984; Feng and Johnson, 1995; Korhola *et al.*, 2002; Oldfield *et al.*, 2003; Quigyu *et al.*, 2010). Magnetic analyses have been used in palaeoenvironmental studies since the 1970's and used as a proxy indicator for temporal variations within sediment cores to suggest past processes (Dearing, 1999). Thompson and Oldfield (1986) termed this application of mineral magnetic analysis, *environmental magnetism*, whereby, processes such as erosion, sedimentation, anthropogenically influenced change (e.g. fire, agriculture, land clearance) and climatic variations could be indicated by variations in the magnetic record. However, a multi-proxy approach is required during such studies to ensure accurate interpretation of magnetic variations (e.g. Thompson and Morton, 1979; Edwards and Rowntree, 1980; Dearing, 1983; Oldfield, 1988; Williams *et al.*, 1996; Doner, 2003; Oldfield *et al.*, 2003; Vannière *et al.*, 2003; Chapron *et al.*, 2007a).

4.8.1 Mineral magnetic behaviour

Magnetisation of a material occurs when a sample is placed within a magnetic field becoming positively (in the direction of the applied field) or negatively (against the direction of the applied field) magnetised. Once removed from the magnetic field magnetisation is lost except in the instance of ferro- and ferrimagnetism. There are five types of magnetic properties; diamagnetism, paramagnetism, ferromagnetism, canted antiferromagnetism and ferrimagnetism (Table 4.2). Magnetic susceptibility relates to the concentration of magnetic crystals within the sample and assumes that the mineral composition controls the magnetic susceptibility of the material (Thompson and Morton, 1979; Dearing, 1999). Although ferromagnetic materials have the greatest magnetic susceptibility and remnant magnetization potential, they are rarely found in the natural environment and rather the concentrations of ferrimagnetic minerals are generally seen to be represented by magnetic susceptibility records.

Associated magnetism	Mineral/material	Mass specific magnetic susceptibility (10^{-6} m^3 kg^{-1})
Ferromagnetism	Iron	276000
Ferrimagnetism	Magnetite	1116 - 580
Canted antiferromagnetism	Hematite	1.19 - 0.6
Paramagnetism	Dolomite	0.011
	Biotite (Mg, Fe, Al silicates)	0.05 – 0.95
	Montmorillonite (clay)	~0.05
Diamagnetism	Calcite	-0.0048
	Quartz	-0.0058
	Water	-0.0090

Table 4.2: Examples of different magnetic minerals and their mass specific susceptibilities (adapted from Thompson and Oldfield, 1986; Dearing, 1999).

4.8.2 The application of environmental magnetism

Magnetic measurements have been used in many palaeoenvironmental studies to provide an assessment of processes that have occurred at a site or catchment i.e. environmental magnetism (Thompson and Oldfield, 1986). Loess deposits have provided a wealth of data using their magnetic susceptibility records, from; correlation between low-field susceptibility in loess samples and the $\delta^{18}O$ signature of marine sediments, evidence of pedogenesis during interglacials marked by enhanced magnetic susceptibility, and climate regimes and their links to pedogenic enhancement (e.g. Kukla *et al.*, 1988; Maher and Thompson, 1991; Feng and Johnson, 1995; Vidic *et al.*, 2004; Qingyu *et al.*, 2010).

Enhanced magnetic susceptibility has been associated with enhanced erosion within a catchment, as increased minerogenic magnetic material is inputted (Mullins, 1977; Dearing, 1991; O'Sullivan, 1994). This can be tested using a multi proxy approach.

Different processes within catchments are interlinked and it must not be taken that enhanced magnetic susceptibility is the determinant of change. Rather each proxy should be evaluated separately and then as a whole group to ensure assumptions made from a change in one proxy can be supported by those in another. Magnetic susceptibility results can suggest periods of possible erosion within a catchment, but as Thompson and Morton (1979) showed particle size analysis and loss-on-ignition can then provide information on the local soil development, sedimentological environment and detrital inputs into the system to ensure that the proxies are interpreted correctly. Investigations of lacustrine sediments have been used in a similar way to loess deposits, but with greater reliance on a multi-proxy approach, predominantly because of the good preservation of biological and sedimentological proxies within lake sediments allowing reconstruction of climate, palaeoenvironment and catchment processes using the mineral magnetic record (e.g. Lanci et al, 2001; Korhola *et al.*, 2002; Vannière *et al.*, 2003; Basavaiah *et al.*, 2004). Chapron *et al.* (2007a) interpreted higher magnetic susceptibility records alongside reduced total organic carbon (TOC) and greater minerogenic input (identified in spectrophotometry data) as indicators of enhanced erosion and allochthonous input into Lake Huez. Particle size analysis and core lithology are then complementary to such magnetic measurements as the depositional environment and type of material (autochthonous or allochthonous) can be ascertained (Williams *et al.*, 1996; Lotter *et al.*, 2002). Leng *et al.* (2010) suggested that inputs of detrital carbonates may also cause rises in magnetic susceptibility.

Climatic instability of the LIA has been identified in magnetic susceptibility records alongside greater variation in particle size due to variable erosion events within catchments (Doner, 2003; Chapron *et al.*, 2007a), resulting in higher detrital inputs and magnetic susceptibility results (Williams *et al.*, 1996). Some research has suggested that an increase in the finer clay fraction and enhanced magnetic susceptibility can be seen as a reduction in the input of pedogenic material, but a multi-proxy approach is vital to ensure accurate interpretation (Mullins, 1977; Thompson and Morton, 1979; Maher, 1986). Vannière *et al.* (2003) demonstrated this at the Lower Doubs Valley site, where sedimentological, magnetic susceptibility, charcoal and palynological analyses were used to provide an interdisciplinary synthesis of change within the catchment. Palynological indicators of human influence e.g. *Secale cereale* and *Cannabis* (arable), were synchronous with enhanced charcoal and magnetic susceptibility records, suggesting the use of fire during cultivation of the area leading to increased catchment erosion and detrital inputs (Vannière *et al.*, 2003). By using palynological records periods of greater magnetic susceptibility can be used to identify periods of soil erosion and/or intensification of agriculture (O'Sullivan, 1994; Vannière *et al.*, 2003). Recent rises in magnetic susceptibility in many lakes have been attributed to anthropogenic influences associated with combustion products and pollution (Cameron *et al.*, 2002; Lotter *et al.*, 2002). However, at the Lower Doubs Valley site enhanced particle size records were associated with lower magnetic susceptibilities, due to the high calcium carbonate content of the local bedrock 'diluting' the magnetic measurements (Vannière *et al.*, 2003). Dilution of magnetic measurements is a common problem in limestone dominated areas, as calcite is diamagnetic and has a similar effect to material rich in organic matter or with high water content as the magnetic susceptibility recorded is dampened (Williams *et al.*, 1996).

Overall, magnetic susceptibility records provide the opportunity to clarify changes within other proxy records. In this study low frequency (LF) magnetic susceptibility of sediment was measured to provide an assessment of erosional events and longer-term changes in catchment scale sediment availability.

4.8.3 Magnetic susceptibility measurement

Material must be placed in a uniform magnetic field (H) to allow the magnetic minerals within a sample to acquire a magnetisation per unit volume (M). The magnetic susceptibility can then be defined as the magnetisation acquired per unit volume (K) following:

$$\mathbf{K = M/H}$$

(Evans and Heller, 2003).

To allow calculation of mass specific susceptibility and ensure changes in core volume did not affect the results a Bartington MS2B dual frequency susceptibility sensor was used (e.g. Dearing, 1999; Björck *et al.*, 2000; Doner, 2003; Hu *et al.*, 2007; Roucoux *et al.*, 2008). Samples were weighed prior to measurement, ensuring **K** was corrected according to differences in sample density (Thompson and Oldfield, 1986; Dearing, 1999). The correction was completed by dividing **K** by the mass of the sample (ρ) to calculate mass specific magnetic susceptibility (X; known as magnetic susceptibility from here on), or specific susceptibility of the sample:

$$\mathbf{X = K/\rho}$$

(Evans and Heller, 2003).

4.8.4 Preparation and analysis of mineral magnetic properties

Samples from LPCA, LPCB and BJC1 were prepared for measurement following the procedure of Dearing (1994, 1999) and Higgitt *et al.* (1991). A minimum 1 cm^3 of sample was taken at 5 cm increments for the length of cores LPCA and LPCB. To allow preparation of ^{137}Cs and magnetic susceptibility samples simultaneously for core BJC1, increments of 2.5cm were used for the top 20 cm of sediment, providing 9 samples for this section. The discrete samples were dried in an oven at 40^{o}C for 72 hours, before being disaggregated and analysing samples for low frequency magnetic susceptibility using the Bartington MS2B dual frequency susceptibility sensor.

Mass specific susceptibility was then calculated following the equation of Evans and Heller (2003):

$$\mathbf{X = K/\rho}$$

4.9 SEDIMENT GRAIN SIZE ANALYSIS

Analysis of sediment grain size has been used for over 100 years to provide information on the depositional environment, entrainment and transport of sediments in fluvial and marine settings (Friedman, 1979; Shu and Collins, 2001; Sperazza *et al.*, 2004; Parris *et al.*, 2010). Palaeohydrological changes can be reconstructed, and where there is evidence of climatic causation, palaeoclimatic changes within a catchment. Lundqvist (1925, 1927 cited in Digerfeldt, 1986) pioneered studies into lake sedimentation processes and sedimentology of lakes less than 100 ha (1km^2), to provide information on past lake level. Digerfeldt (1986) and Dearing (1997) have since modified the method using variations in particle size to suggest displacement of the shoreline, with coarse grains associated with the near-shore or littoral lake zone and finer fractions with central lake sediments (Digerfeldt, 1986). Particle size fluctuations may also be a product of catchment changes with increased erosion resulting in greater coarse minerogenic inputs. Finer sediments may infer periods of catchment stability or proximity to the centre of the lake where finer sediments accumulate (e.g. Menzel *et al.*, 1985; Lobo *et al.*, 2001; Punning *et al.*, 2006; Wantzen *et al.*, 2008; Thornes *et al.*, 2009; Marion *et al.*, 2010; Waele *et al.*, 2010). This range of possible interpretations of a particle size record highlights the need for multi-proxy and/or multi-core analysis to understand the processes occurring.

A variety of methods can be used to reveal the grain size distribution, such as the traditional sieve/pipette or hydrometer method, or more recent methods including X-ray attenuation, scanning electron microscopy and laser diffraction methods (e.g. Jordan *et al.*, 1971; Weiss and Frok, 1976; Cheetham *et al.*, 2008; Sorti and Balsamo, 2009). The sieve/pipette or hydrometer method is time consuming and dependent on techniques used and operator error results can be greater than 40% (Syvitski *et al.*, 1991; Sperazza *et al.*, 2004). Investigations into analysis by X-ray attenuation, laser diffraction and scanning electron microscopy techniques against the traditional sieve/pipette method have been conducted by Beuselinck *et al.* (1998) and Cheetham *et al.* (2008). These studies found that the X-ray attenuation method was impractical for analysis of sand-dominated fluvial sediments as it cannot be used for grains larger than 250 µm. Larger fractions proved to be the limitation for the scanning electron microscopy method too. The laser diffraction method was shown to provide replicable results efficiently, as silts to coarse sands may be analysed (Cheetham *et al.*, 2008) and on the basis of this the method was used for this research project.

4.9.1 Laser diffractometry

Laser diffractometers operate on the principle that sediment particles of a given size will diffract light at a particular angle. As the particle size decreases the angle of diffracted light increases. The sample travels through a sample cell where a beam of monochromatic light passes through the sample. The diffracted light is then focused on to an array of detectors and the sediment grain-size distribution is calculated by the intensity of light reaching the detectors (Weiss and Frock, 1976; Beuselinck *et al.*, 1998). Using inconsistent methods or sample preparation has been shown to highly affect results when using the laser diffraction method (e.g. Sperazza *et al.*, 2004; Cheetham *et al.*, 2008; Sorti and Balsamo, 2009). To ensure this did not occur during this project a consistent method was maintained.

The laser diffractometer used during analysis was the Malvern Mastersizer 2000G, which has been tested by Sperazza *et al.* (2004), and Sorti and Balsamo (2009) on clay-rich sediments, sands and catalclastic breccias. These studies provided information on good sample preparation and the optical and different machine parameters to use. To ensure reliable processing of the data an optical model must be selected which is based on one of two diffraction theories; the Mie or Fraunhofer theory. Mie theory assumes grain size is: <10 µm and particles are not opaque, but this model is inaccurate for grains >10 µm (de Boer *et al.*, 1987; Blott and Pye, 2006). The Fraunhofer model assumes that all particles are opaque discs, but more importantly can be used on samples with a variety of particle sizes from fine to coarse grains (Weiss and Frock, 1976; Pye and Blott, 2004) and was thus appropriate for the Lake Plav sediments. To ensure comparable results are produced throughout analysis consistent sample concentrations must be used i.e. obscuration of the narrow light beam passed through the sample must remain at around 15%. Obscuration above 15% has been found to decrease the median grain size of finer sediments, providing inaccurate results (Pye and Blott, 2004; Sperazza *et al.,* 2004). Guaranteeing representative results required consideration of the material strength, disaggregation and optical properties of the sediments prior to analysis. As field-based textural analysis of the Lake Plav sediment indicated high clay content, samples required a pre-analysis treatment to disperse these particles. Flocculated clay particles present a larger target for the optical laser skewing results to larger particle sizes (McCave *et al.*, 1986). Sodium hexametaphostphate (Calgon) was identified by Tchillingarian (1952) as an effective dispersion agent for clays and research by Pye and Blott (2004) and Sperazza *et al.* (2004) advocated a concentration of 5.5 g/l before leaving samples to disaggregate for at least 24 hours following. However, prior to dispersion of samples it is important to remove any obvious organic material with hydrogen peroxide (H_2O_2; Robinson, 1922; Lewis, 1984), as organics much like flocculated clays, provide a larger target for the beam of light, skewing results. Furthermore, pre-treatment with H_2O_2 has been suggested to aid de-flocculation (Gray *et al.*, 2010). Many authors also suggest using an HCl wash to eliminate carbonates (Allen, 1997; Murray, 2002) although any signal of catchment disturbance associated with the carbonate fraction is then destroyed (Murray, 2002). Therefore, a pre-treatment of HCl could not be used on the Lake Plav sediments. Instead to a pre-measurement application of sonication was used to ensure clay particles were fully dispersed before analysis. As sonication for a period longer than 60 seconds and greater than 10 μm tip displacement has been shown to cause particle disintegration (Pye and Blott, 2004; Sperazza *et al.*, 2004; Storti and Balsamo, 2009), the Lake Plav sediments were tested to ensure the correct period of sonication was used during analysis.

To test the level of sonication required during this project, two samples were analysed; (1) a silty clay, PSA1, from core LPCA at 167 cm and (2) a sand, PSA2, from core LPCB 465 - 470 cm. Each sample was treated with hydrogen peroxide and Sodium hexametaphostphate before sonication was applied for; 10, 20, 25, 30, 35, 40, 50 and 60 seconds consecutively. Following each application the particle size distributions were measured.

Sample PSA1 was affected by the length of sonication applied more so than sample PSA2 due to the dominance of the fine fraction within the sample. Sonication at 10 and 20 seconds produced rather uni-modal results with a limited clay fraction contrary to field based textural analysis. However, sonication at 25 to 35 seconds and 60 seconds produced results that did go some way to account for the clay fraction in the sample

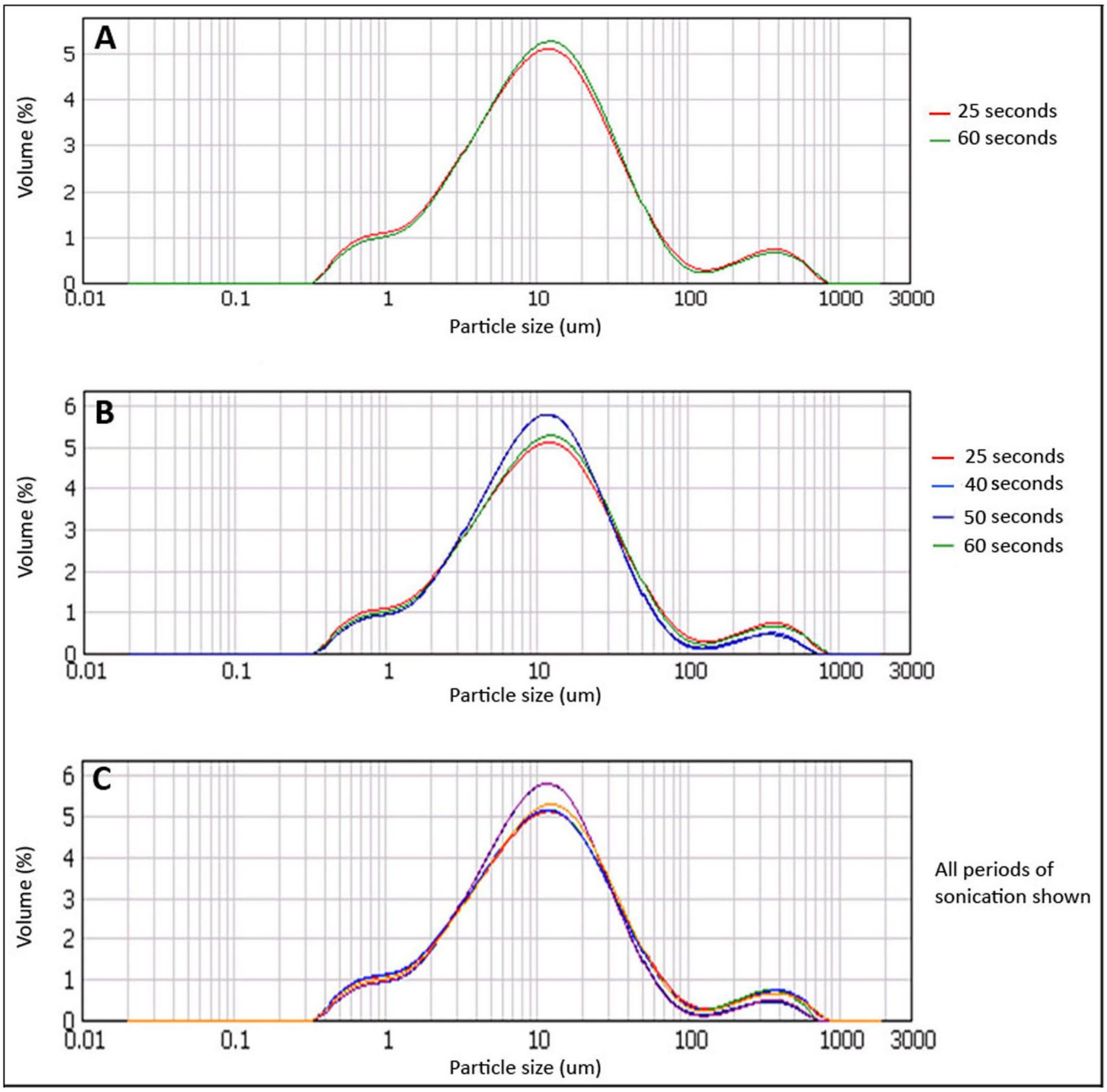


Figure 4.11: Malvern Mastersizer screen shots of sonication test results for silty clay, PSA1. (A) 25 and 60 seconds sonication, (B) 25, 40, 50 and 60 seconds sonication and (C) all sonication lengths tested.

suggesting replicable results could be gleaned using sonication for any of these periods (Figure 4.11). Between 40 and 50 seconds pre-measurement sonication produced results suggesting the silt fraction was larger than during any other period of pre-measurement sonication. Sample PSA2 produced stable results throughout the tested use of sonication and thus it was assumed that sonication would not cause particle disintegration in more sandy samples up to 60 seconds (Figure 4.12), in agreement with previous studies (Pye and Blott, 2004; Sperazza *et al.*, 2004; Storti and Blasamo, 2009). Therefore, the decision to use 25 seconds at 10 µm tip displacement pre-measurement sonication was a consequence of the synchronisation between the majority of results on all but the 10, 40 and 50 second tests.

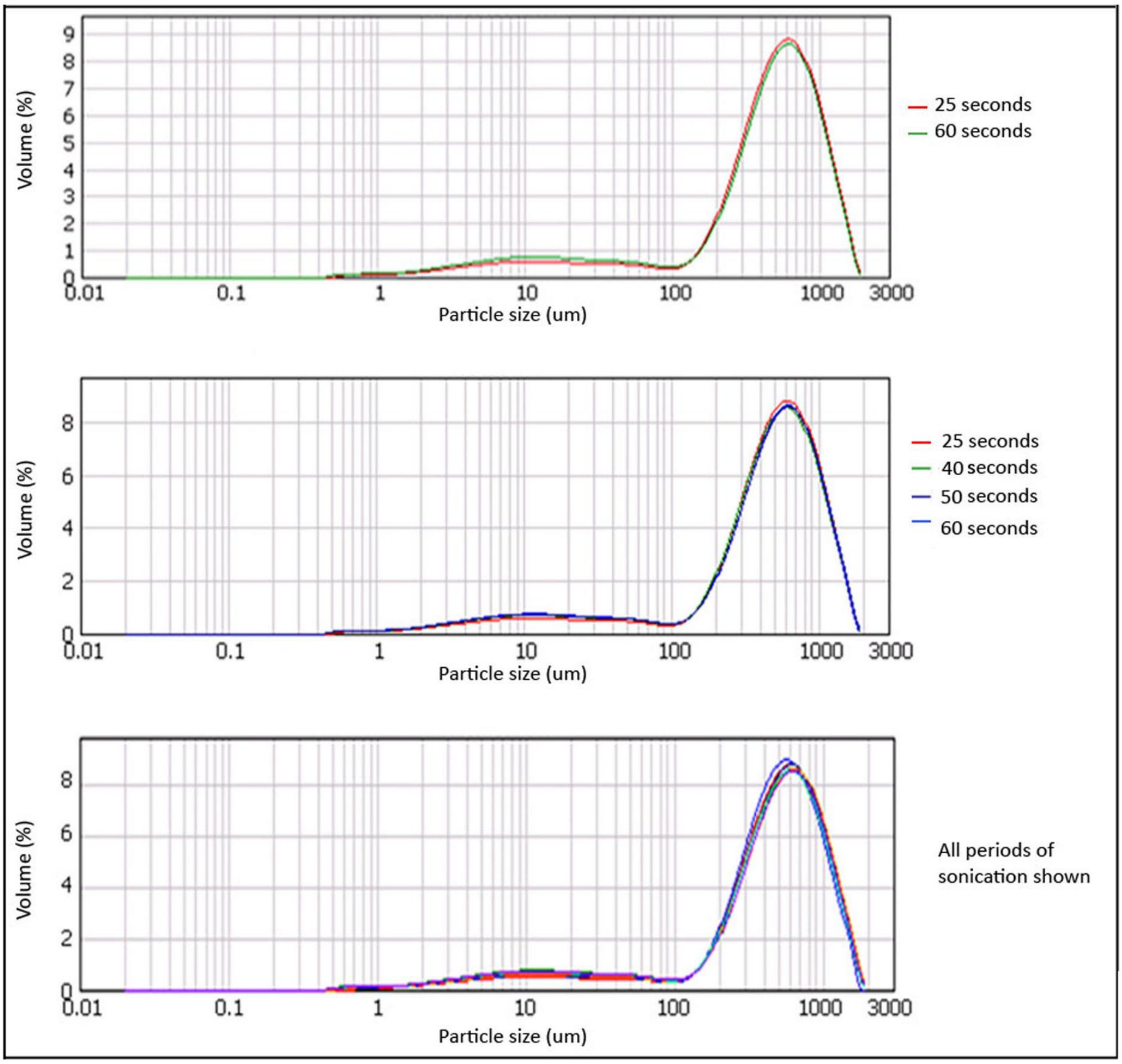


Figure 4.12: Screen shots of Malvern Mastersizer outputs for sandy PSA2. (A) 25 and 60, (B) 25 40, 50 and 60, (C) All.

Obscuration levels were also tested at this time and it was found that wet samples of 6 g L^{-1} for clays and 18 g L^{-1} for very coarse sands produced an acceptable obscuration (~15%) for analysis. Before samples could be analysed a standard operating procedure (SOP) was created, this is the method that the laser diffractor will use throughout the analysis and ensures the use of a particular optical model and consistent levels of sonication, pump and stirrer speeds. A method that accounted for variation in grain size and optical properties of the sediments from Lake Plav was developed following the study on lacustrine sediments by Sperazza *et al.* (2004). As previously mentioned the Fraunhofer model was employed due to the range of grain sizes to be analysed and the pump speed selected was 2100 RPM. This speed followed research by Sperazza *et al.* (2004) and Storti and Balsamo (2009) that suggested pump speeds between 2000 – 2300 RPM would ensure re-circulation of coarser materials, without significantly reducing the precision of the finer grains. Analysis times of less than 5 seconds were found to produce less reliable results (Storti and Balsamo, 2009), thus the Malvern Mastersizer default was used (12 seconds). The Lake Plav SOP settings can be summarised as:

- Fraunhofer optical model.
- Pre-measurement sonication (see above).
- Pump speed 2100RPM.
- Stirrer speed 850RPM (Malvern Mastersizer default).
- Sediments analysed for 12 seconds, repeated 3 times for average measurement.

Finally, during the testing of the Malvern Mastersizer results highlighted the dominance of silt sized particles. Following the textural analysis results it was thought that the Malvern Mastersizer was overlooking the variability in finer grain sizes and not differentiating between clay and silt fractions, a problem noted by Murray (2002) and Palmer (*pers. comm.*). Therefore, a 'clay dominated' sample from core LPCA at 300 cm was analysed at x40 magnification and the results suggested that the Malvern Mastersizer was correct as the sample was dominated by silt sized particles (see Figure 4.13). In field analysis of clays and silts in the field can be difficult if sediments are wet or the fraction sizes are close to the boundary between the two types i.e. 2 µm. Differentiation between textural analysis results and Malvern Mastersizer results are therefore indicated during the presentation of sediment composition results in Chapter 5.

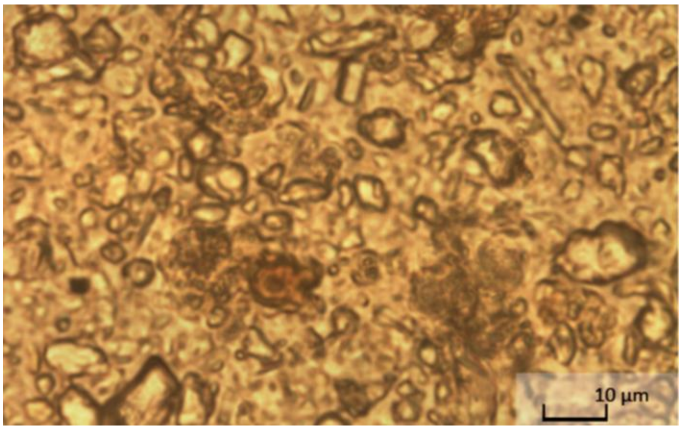


Figure 4.13: Sediment sample from core LPCA at x40 magnification.

4.10 ORGANIC CONTENT AND LOSS-ON-IGNITION

Organic content of lacustrine sediments or percentage loss-on-ignition (% LOI), as it is commonly referred to, provides a record of lake productivity and allochthonous inputs into a lake (e.g. Davis and Ford, 1982; Nesje and Dahl, 2000; Shuman, 2003; Dalton *et al.*, 2005). LOI determination is a simple, cost effective and time efficient preparation, in which the percentage of organic carbon in a sample can be determined (Heiri *et al.*, 2001; Wang *et al.*, 2010). The influence of allochthonous inputs into the littoral zones of lakes has been identified by many authors as suppressed LOI results, whereas rises in LOI are generally associated with periods of greater lake productivity and terrestrial sites (Stevens *et al.*, 2001; Reinemann *et al.*, 2009). Dalton *et al.* (2005) suggested that less pronounced fluctuations in LOI of littoral sediments can be associated with increased allochthonous inputs and climatically unstable periods. Lake evolution and periods of climatic stability conducive to re-vegetation (organic rich or 'swamp' areas indicated by pollen types such as *Typha*) and shoreline stability may be associated with higher LOI results (Lanci *et al.*, 1999; Dalton *et al.*, 2005; McGowan *et al.*, 2008; Margielewski *et al.*, 2010; Panek *et al.*, 2010). As with the interpretation of magnetic parameters and particle size analysis the variety of influences upon LOI underlines the need for a multi-proxy analysis, to determine the factors inducing variations.

LOI was calculated for cores LPCA and LPCB following the method of Bengtsson and Enell (1986) and Heiri *et al.* (2001), with samples placed in a furnace at 550^{o}C for 4 hours.

4.11 SEDIMENT DATING TECHNIQUES

Dating of sediments is vital to understand the temporal scale on which a particular record is based. A range of methods are available from qualitative/comparative to radioactive nuclide and radiative dosimetry methods (Jull, 2007). Birks and Birks (1980) suggested that radiocarbon dating was the main method for creating an independent chronology. To ensure an appropriate dating method is selected, where possible, preliminary dating of sediments is required. Qualitative/comparative methods and radioactive nuclide dating were used during this study, as other methods do not provide a dating scale suitable for this project.

4.11.1 Comparative and qualitative dating methods

General comparisons between transect cores can occur with lithological, magnetic susceptibility or palynological changes correlated to similar changes in dated cores (e.g. Kukla *et al.*, 1988; Maher and Thompson, 1991; Williamson *et al.*, 1991; Chen *et al.*, 1999; Machlus *et al.*, 2000; Ito, 2002). The transect cores extracted from Lake Plav were not independently dated due to cost restraints, but independent dating chronologies for cores LPCA and LPCB were used as a comparative dating technique where applicable.

4.11.2 Radioactive nuclide dating

Radioactive nuclide dating relies upon the radioactive decay of unstable elements and the associated half-life (Bell and Walker, 1992). The concentration of a particular radioactive nuclide is used to estimate the age of a deposit, as nuclides will increase or decrease over time (Jull, 2007). Dating of fine-grained sediments and organics are particularly suited to AMS radiocarbon dating (^{14}C dating), lead-210 dating (^{210}Pb) and caesium-137 dating (^{137}Cs), all of which can be used to cover Holocene deposits (Table 4.3; e.g. Ely *et al.*, 1992; Wohlfarth, 1996; Thorndycraft *et al.*, 2005; Arnaud *et al.*, 2006; Enters *et al.*, 2006).

Isotopic system	Half-life	Timescale	Material
^{14}C	5730 yr	250 yr – 50 kyr	Organic material
^{210}Pb	22.3 yr	0 – 150 yr	Fine grained sediments
^{137}Cs	30 yr	Post 1950	Fine grained sediments

Table 4.3: Dating methods suitable for the Little Ice Age deposits (adapted from Jull, 2007).

4.11.3 AMS radiocarbon dating

Birks and Birks (1980) suggested the use of radiocarbon dating provides a reliable means of dating. The traditional method of radiocarbon analysis relies on the detection of decaying ^{14}C atoms in the sample (Linick *et al.*, 1989). In contrast, AMS (Accelerator Mass Spectrometry) radiocarbon dating uses an accelerator mass spectrometer to count all ^{14}C atoms within the sample. Conventional radiocarbon dating requires between 20 – 100 g of sample material, but for AMS dating only 20 – 50 mg are needed (following Beta Analytic Inc.). Radiocarbon dating has been used on a variety of sediments from lacustrine sediments to peaty or marine material and has an upper limit of around 40,000 to 50,000 yrs BP (e.g. Bell and Walker, 1992; Jorissen *et al.*, 1993; Björck and Wohlfarth, 2001; Feurdean *et al.*, 2007; Webb, 2001; Breitenlechner *et al.*, 2010; David, 2010).

Core code	Laboratory code	Depth extracted (m)	Type of material	Sample weight (g)	Date sent for analysis
LPCA	RWLPC1760	7.60	Plant fragment, possibly *Equisetum*	0.0700	09.10.2008
LPCB	RWLP5401	5.40	Wood	0.2100	09.10.2008
LPCB	RWLP2675	26.75	Herbaceous plant fragment	0.2550	16.01.2009
BJC1	RWBJ113	1.13	Possibly dung, coarse textured organic material	0.0446	21.11.2010

Table 4.4: AMS radiocarbon sample information.

Both traditional and AMS dating methods do, however, require calibration due to variations in atmospheric ^{14}C throughout the Holocene (Bell and Walker, 1992). Calibration was undertaken by Beta Analytic Inc. and for this project the Calib 4.1 radiocarbon calibration program was used (Stuiver and Reimer, 1993). Although the applications of AMS radiocarbon dating are far reaching, dating errors of between 40 and 100 years can occur. The method is also inaccurate for more recent samples, with a lower limit of around 250 years i.e. AD 1750 to present (Beta Analytic Inc. *pers. comm.*). Four samples were sent for AMS radiocarbon dating, one from LPCA at 7.60 m, two from LPCB at 5.40 m and 26.75 m and a single sample from BJC1 at 1.13 m. AMS dating was chosen due to the limited organic fragments preserved in each of the cores and small sample masses required (Table 4.4). The fourth sample was extracted from core BJC1 at 1.13 m, as sedimentation rates for similar 'high mountain' situations are between 40 – 200 cm/1000 years (e.g. Tonkov *et al.*, 2002; 2006; Lotter *et al.*, 2006; Stefanova *et al.*, 2006a,b; Romera *et al.*, 2010).

All samples were sent to Beta Analytic Inc., Miami in USA for AMS radiocarbon dating and prepared using an 'acid/alkali/acid' method, the full method is provided in Appendix VI. Following the ^{14}C dating results ^{210}Pb and ^{137}Cs dating was pursued for all three cores to provide an alternative chronology for the upper sediments.

4.11.4 Short-lived radionuclide dating

^{210}Pb and ^{137}Cs are two common methods for calculating recent geochronologies in lacustrine environments (Benoit and Rozan, 2001). Both isotopes are comparatively short-lived to ^{14}C (Table 4.3) allowing sediments up to 150 years old to be dated (Birks and Birks, 1980). Vertical mixing of sediments can cause re-distribution of radionuclides within the sediment column, by biological means or chemical diffusion, but this is identified by a flattening of ^{210}Pb activity or degradation of ^{137}Cs activity peaks. These complementary dating methods allow recent changes in ^{210}Pb to be verified with ^{137}Cs analysis (Ritchie *et al.*, 1974; Appleby, 2001). The application and process of each dating method is discussed below.

Lead-210 dating

The ^{210}Pb method was devised by Goldberg in 1963 and relies upon the radioactive decay of radon gas (^{222}Rn) which is part of the U-series dating decay chain (Lowe and Walker, 1984; Olsson, 1986). Upon decay of ^{222}Rn, ^{210}Pb is produced as a daughter nuclide, but it also occurs naturally in sediments which contain Uranium-238 (^{238}U) and in the atmosphere. This dating method has proven particularly useful where accumulation rates are uniform and little biological disturbance occurs in the upper surface layers (Appleby, 2001), allowing sediment accumulation rates since ***c.*** AD 1850 to be calculated (Appleby *et al.*, 1979; Lowe and Walker, 1984; Humphries *et al.*, 2010). Different approaches to ^{210}Pb estimation have been used with the gamma-ray (γ-ray) spectrometry being the simplest method, allowing direct interpretation of results (Tanner *et al.*, 2000). However, some methods for ^{210}Pb dating can be destructive e.g. the approach using chemical preparation of a sample and in-growth of polonium-210 (^{210}Po), the alpha-emitting granddaughter of ^{210}Pb (Appleby *et al.*, 1979). Finally, ^{210}Pb has been used as a tool for estimating biogeochemical cycles within lacustrine and other systems because of its unique properties that allow tracing of atmospheric and aquatic depositional process since the 1850s (Robbins, 1978; Krishnawami *et al.*, 1980).

^{210}Pb dating was completed at the University of Lancaster under the supervision of Dr. Jackie Pates, employing the 'total dissolution' method, a procedure that assumes equilibrium with ^{210}Po. Small amounts of ^{210}Pb in sediments originate from the decay of uranium or its isotopic daughters and this is termed supported ^{210}Pb. Unsupported ^{210}Pb is that produced in the atmosphere, thus in the total dissolution method, both supported and unsupported ^{210}Po are analysed to produce a ^{210}Pb profile. Only LPCA was analysed for ^{210}Pb as this core had been a focus of much study and required an accurate assessment of sedimentation rates. Further ^{210}Pb analyses could not be completed on LPCB or BJC1 due to cost constraints. Samples for ^{210}Pb analysis were taken at 5 cm intervals up to 8 0cm and an outlier sample taken at 100 cm, to ensure the beginning of the unsupported ^{210}Pb curve was captured. Two further samples were extracted at 7.5 cm and 12.5 cm for ^{137}Cs analysis and were included in the ^{210}Pb analysis to ensure comparability between both data sets.

Samples were dried and ground to a homogeneous composition before weighing 0.5 g sediment and adding 0.1 g of ^{209}Po yield monitor. 10 ml of 6M HCL was then added and each sample taken to dryness, before adding 20 ml H_2O_2 and taken to dryness. The latter step was repeated until the initial reaction was weak. 10 ml HF, 20 ml HNO_3 + HCl were added, before covering the samples with a watch glass and allowing reflux for 4 hours. Samples were taken to dryness before adding 50 ml 6M HCl, to ensure the polonium was put into solution, and then filtered to remove the solid fraction. Plating of the samples using copper "discs" required samples to be taken to dryness, adding 100 ml Milli-Q, 5 ml 30% hydroxyl ammonium chloride (NH_3OHCl) and 5 mg bismuth nitrate. Each 2.5 x 2.5 cm copper "disc" could then be briefly immersed in 2M HNO_3 and rinsed with de-ionised water before threading on to 10 cm lengths of fishing line. A copper discs was hung into each sample and the beakers heated at a constant temperature and stirred for 3 hours. The disc was then removed and dried over-night. Dr. Jackie Pates then determined activity of ^{210}Po and ^{208}Po by alpha (α-) spectrometry.

Caesium-137 dating

Unlike ^{210}Pb, ^{137}Cs is generated exclusively by anthropogenic means as a result of thermonuclear weapons testing. Dating relies upon evidence of spikes in the radionuclide activity record to provide spot dates within a profile rather than continuous dating throughout a profile, as per ^{210}Pb (e.g. Pennington *et al.*, 1973; Ritchie *et al.*, 1973; Lowe and Walker, 1984). Within the literature there are conflicting dates for the period when the first significant quantities of ^{137}Cs were produced, with Lowe and Walker (1984) suggesting AD 1945 and Pennington *et al.* (1973) and Ritchie *et al.* (1973), recommending ***c.*** AD 1954. These early literature do though, agree on the maximum peak occurring at AD 1963, but a more recent marker is the product of the Chernobyl accident in AD 1986 (Callaway *et al.*, 1996; Carroll *et al.*, 1999a; Erlinger *et al.*, 2008). The Chernobyl marker provides far higher levels of ^{137}Cs, between 150 - 400 Bq/kg in many sites across Europe (e.g. Pinglot and Pourchet, 1995; Jensen *et al.*, 2002; O'Reilly *et al.*, 2010). The ^{137}Cs radioisotope is deposited on the ground by rainfall, where its decay can be evaluated using γ -ray spectrometry. Samples were prepared from LPCA, LPCB and BJC1 for analysis at the University of Lancaster by Dr. Jackie Pates. 1cm^3 samples were taken from core LPCA at 5 cm intervals to 25 cm, with two further samples at 7.5 cm and 12.5 cm. Four samples were sent for analysis from core LPCB; 0 – 5 cm, 5 – 10 cm, 10 – 15 cm and 15 – 25 cm, and core BJC1 was sampled at 1 cm contiguous increments to 10 cm. The majority of samples for LPCA and all for LPCB were 10 – 20 g. The remaining samples from cores LPCA and BJC1 were 1.5 g. Samples were packed into small labelled petri dishes and evaluated using gamma-ray (γ) spectrometry by Dr Jackie Pates.

4.12 PALAEOCLIMATIC MODELLING USING GLACIERS

Palaeoclimatic inferences can be made from glacial geomorphological reconstructions to provide a direct source of climatic data (Nesje and Dahl, 2000). For the purposes of this research a glacier-climate model was used to reconstruct past temperatures during the LIA at the Maja e Koljaet glacier, enabling comparisons to present-day conditions. Two approaches for palaeoclimatic modelling were utilised for reconstructing climate over this glacier: (1) a degree-day model (DDM) (Brugger, 2006; Hughes 2008) and (2) a polynomial regression equation (Ohmura *et al.*, 1992). DDMs have been used to determine the required temperature change to maintain an independently identified ELA or to assess the variations in temperature needed for a steady-state mass balance (e.g. Brugger, 2006; Hughes, 2008, 2010; Hughes and Braithwaite, 2008). A DDM calculates the amount of melting that occurs under a given temperature regime. The known or reconstructed mean annual temperature can then be distributed over 365 days (1 calendar year) using a sine curve distribution following Brugger (2006):

$$T_d = A_y \sin(2\pi/\lambda - \phi) + T_a$$

Where:

T_d is the mean daily temperature

A_y relates to the annual temperature range

λ is the period covered i.e. 365 days

Φ represents the phase angle (taken as 1.93 radians to assume that January is the coolest month)

T_a is the mean annual air temperature

The volume of snow melt per day resulting from the reconstructed mean daily temperature is then calculated using a degree day factor, where accumulation and daily snow melt are equal. Hughes (2008, 2010) used a degree-day factor of ***c.*** 4 mm day^{-1} K^{-1} following research by Braithwaite (2008) and Braithwaite *et al.* (2006), that established degree-day factors of snow between 3.5 $\pm$ 1.5 mm day^{-1} K^{-1} and 4.5 $\pm$ 1.4 mm day^{-1} K^{-1} using 66 glaciers, with a mean of 4.1 mm day^{-1} K^{-1}. For the purposes of this study a simple DDM was used in Excel, employing the Brugger equation along with user-defined inputs of mean annual temperature and mean annual temperature range to reconstruct the mean daily temperature distribution across 365 days. A degree-day factor of 4 mm day^{-1} K^{-1} (Braithwaite *et al.*, 2006) was then multiplied by each individual daily temperature mean to calculate daily melt, the sum of which provided the mean annual melt. In order to maintain a glaciers survival mean annual melt must be equalled by annual accumulation under any given temperature regime i.e. the calculation of annual melt equates to the annual accumulation needed to maintain the ELA/glacier at a particular point (Hughes, 2008).

The second method of Ohmura *et al.* (1992) employs a polynomial regression equation based on the relationship between winter balance + summer precipitation and ablation (winter balance + summer precipitation is effectively close in value to the annual precipitation, except that in some situations local controls such as avalanching and windblown snow can affect winter balance). Ohmura *et al.* (1992) obtained data from direct monitoring of 70 contemporary glaciers mass-balances in polar and mid-latitude regions. The study resulted in the production of a best-fit polynomial regression curve for the ELA of each glacier (Figure 4.14) in relation to winter balance + summer precipitation and average June/July/August temperature experienced at each site, developing the following equation:

$$P = a + bT + cT^2$$

Where:

P *is* winter balance + summer precipitation *expressed in mm w.e.*

T *is temperature* (^oC)*,* a = 645, b = 296, c = 9

Using this equation, average June/July/August temperature data can be inputted to the regression to calculate the winter balance + summer precipitation (or annual precipitation) required to maintain a glacier with a particular ELA.

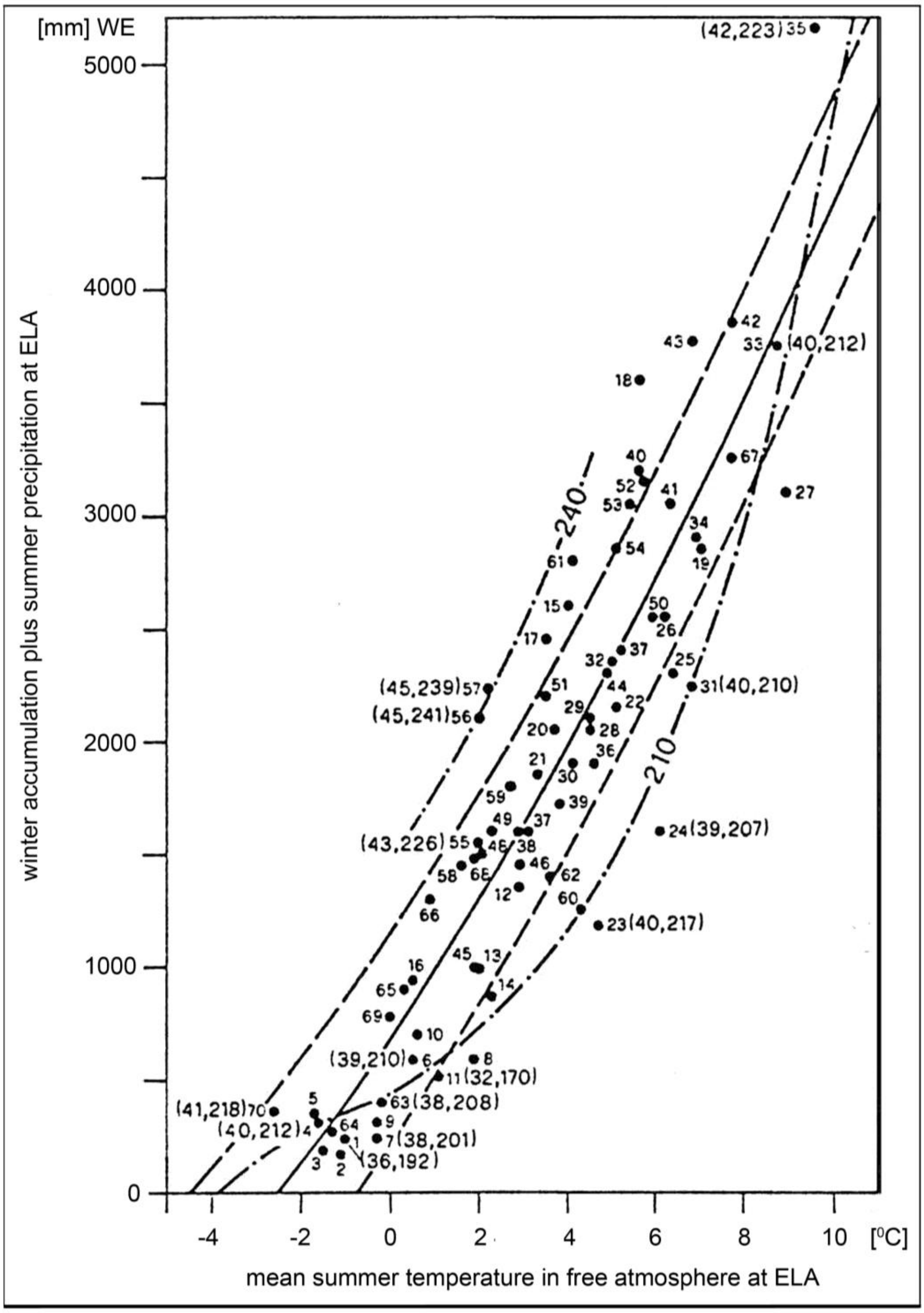


Figure 4.14: Best-fit polynomial regression curve for the ELA of 70 glaciers (taken from Ohmura *et al.*, 1992).

The DDM method can be used to assess the climate at any section of a glacier from the ELA to frontal moraine as it is not designed to be used specifically at the ELA of a glacier.

However, if the palaeoclimate at an ELA is needed, an estimation of the ELA can be calculated using methods such as the standard accumulation-area method (AAR; Benn and Evans, 1998; Brugger, 2006), or via the median elevation (Braithwaite and Raper, 2007). This separates a glacier into two equal parts and has been shown by Hughes (2008, 2010) to provide the best ELA estimation for small glaciers in the Durmitor massif in Montenegro. On small high mountain glaciers there is little altitudinal variation across the glacier, unlike larger mountain glaciers (Benn and Lehmkuhl, 2000) causing the position of the ELA to vary annually (Hughes, 2008). Small glaciers may, therefore, experience periods of negative net balance (i.e. greater ablation than accumulation), triggering the ELA to rise and in some cases causing the whole glacier to reside below the theoretical ELA. Conversely, if a positive net balance is experienced (accumulation > melting), a glacier may actually exist entirely above the ELA. When comparing the ELA changes on a glacier through time, even the median elevation of the glacier surface can be problematic. The hypsometry (shape-altitude-area distribution) of such glaciers may vary dramatically through time, depending on the shape of the host cirque basin. Thus, for very small glaciers (<0.5 km^2) the elevation of the moraine front is likely to provide the most representative altitudinal value for the ELA (Hughes *pers. comm.*). As altitudinal distribution of very small glaciers is small, any error in estimating the ELA is likely to be low no matter where it is positioned. Contemporary temperature data is also required to allow palaeoclimatic reconstructions using the DDM or Ohmura *et al.* (1992) method, but no such records were available for Maja e Koljaet glacier or Plav. Records were available for Vermosh within the Lake Plav catchment (Figure 3.3, Chapter 3), but these were incomplete and insufficient for input into a DDM (Palmentola *et al.*, 1995). The most appropriate records available were from Kolašin (954 m a.s.l.) 40 km north-west of Plav, covering the period AD 1973 – 2008, with relatively complete monthly data (Appendix VII). Average annual temperatures were calculated for Kolašin between AD 1973 - 2008 providing an annual temperature range of 19.4^{o}C (Tutiempo, 2009). Kolašin is around 1000 m lower than the Maja e Koljaet glacier and therefore, temperature data were extrapolated using a standard atmospheric lapse rate of 0.6^{o}C/100 m (Table 4.5, Barry and Chorley, 1987). These data were inputted into the relevant models to determine the annual accumulation (or precipitation) required to maintain a glacier and thus infer the temperatures at past ELAs.

Altitude (m a.s.l.)	Mean annual temperature data for DDM (°C)	Altitude (m a.s.l.)	JJA temperture data for Ohmura *et al.* (1992) model
954 (Kolašin)	8.5	954 (Kolašin)	17.2
1800	3.42	1800	12.12
1850	3.12	1850	11.82
1900	2.82	1900	11.52
1950	2.52	1950	11.22
2000	2.22	2000	10.92

Table 4.5: Extrapolated Kolašin temperature data for various altitudes around the Maja e Koljaet cirque basin.

5.1 LAKE PLAV CATCHMENT VEGETATION SURVEY RESULTS

During the contemporary vegetation assessment of the Lake Plav catchment 29 sites were investigated. The sampling network incorporated a range of elevations and aspects to ensure that all vegetation zones were analysed. Ten sites downstream (north-east) of Lake Plav were analysed to ensure that areas with greater anthropogenic modification and elevation below 900 m a.s.l. were captured. Sites were generally focused at the River Ljuča floodplain and near-by slopes of Visitor Mountain, as these sites were most accessible. Tributary valleys were investigated where possible, but the surrounding high, steep and rocky slopes could not be surveyed. The vegetation recorded was checked using a combination of remote field observations and maps of the area, allowing the dominant vegetation cover and characteristics to be identified. Using a DEM for the area (supplied by Robert Aleksić at the United Nations Development Programme Montenegro) the dominant vegetation types and altitudinal range covered by each were then inputted into ArcGIS. This enabled a contemporary catchment vegetation map to be created in which 14 vegetation zones were identified (Figure 5.1; Table 5.1).

The most species diverse zones were I to V as anthropogenic clearance seemed less conducive to the monoculture of *Fagus* and coniferous woodland that dominated other zones. Floodplain vegetation was dominated by wildflowers and *Poaceae* (see Figure 5.1), but around Lake Plav extensive wetlands existed with types such as *Phragmites*, *Cyperaceae* and *Equisetum*. Adjacent to most homes in the catchment small plots of land were cultivated with types such as cabbages (*Brassica*), corn (*Zea* type) and fruit trees (*Prunus*; see Figure 5.3) favoured. This is an example of 'garden agriculture' following McNeil (1992) and was focused in zones I and II. Anthropogenic activity was particularly evident around Plav, but also Vusanje and Gusinje. Figure 5.2 highlights the cleared land used for grazing livestock, which rises to around 1650 m a.s.l. around Plav.

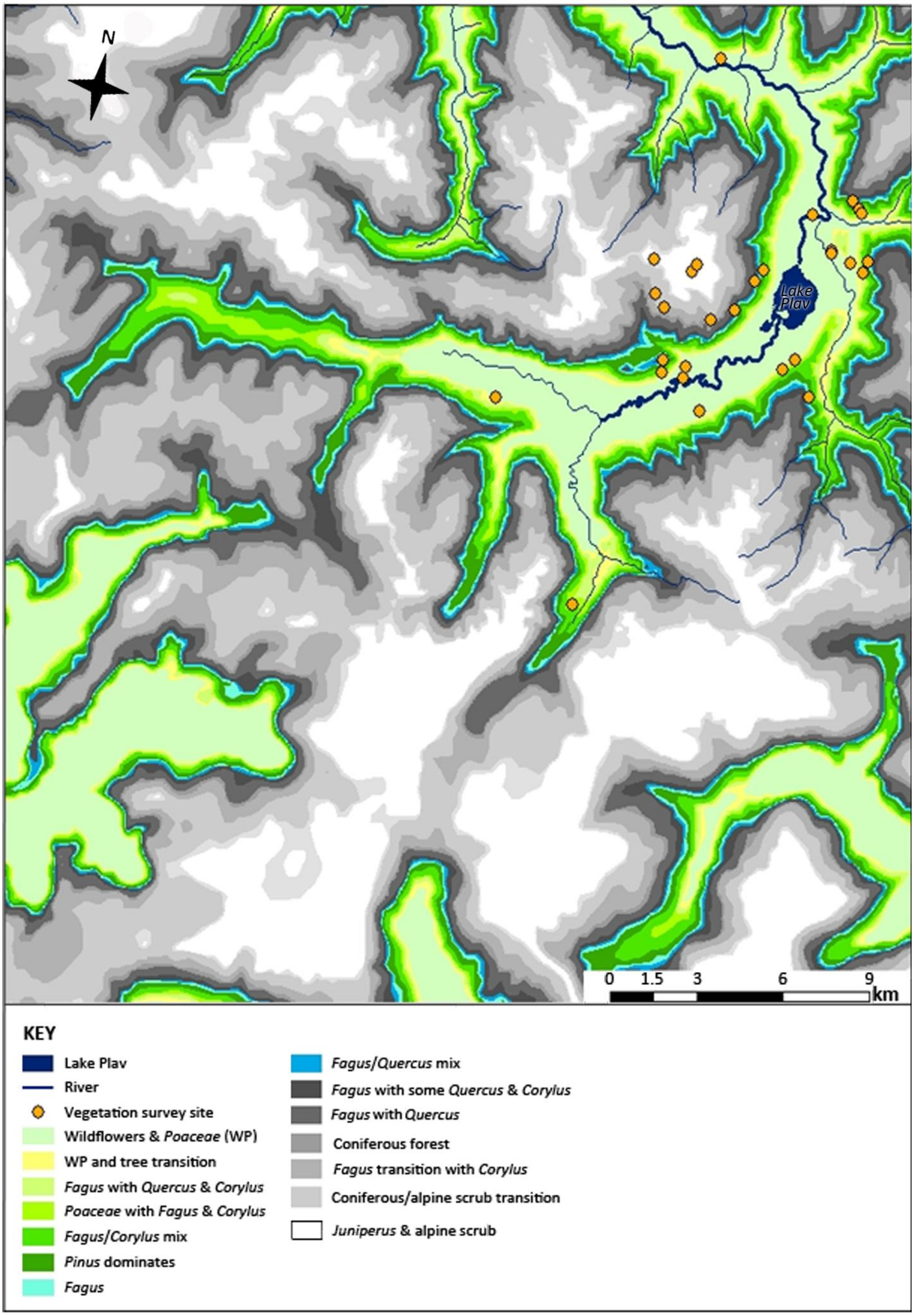


Figure 5.1: Contemporary vegetation map of the Lake Plav catchment, created using ArcGIS.

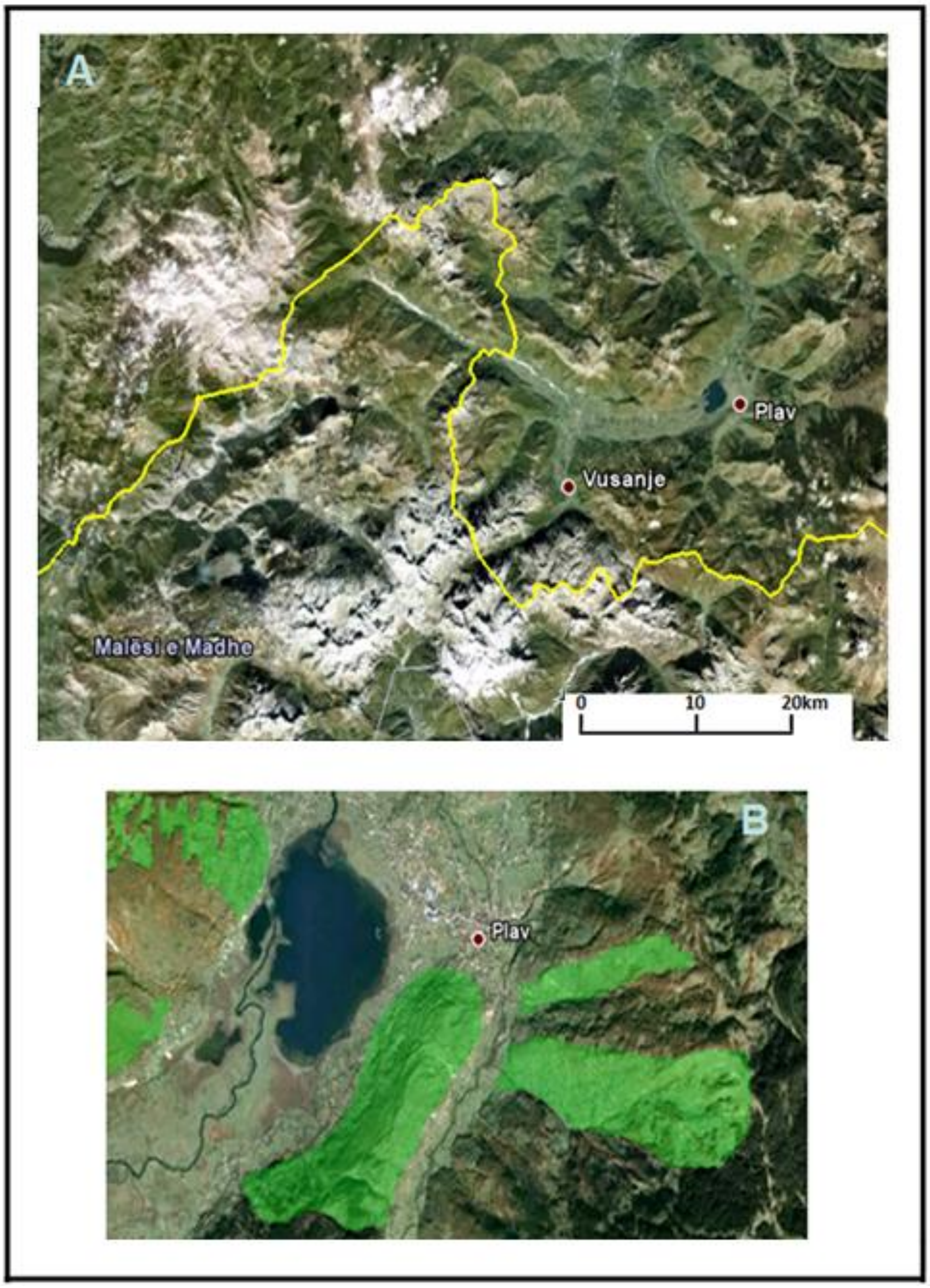


Figure 5.2: (A) Google earth image of the Lake Plav catchment. (B) Area around Plav, green shading represents the areas that have been anthropogenically cleared around Plav town (Google earth, 2010).

In the higher sections of meadowland, >1500 m a.s.l., large *Juniperus communis* shrubs were scattered within the cleared areas (see Figure 5.3). The *Fagus* woodlands of the area are actively managed by coppicing to provide a local source of firewood, with the trees growing up to zone XII (<1800 m). Pastoral farming dominates anthropogenic activity in the catchment with areas around Lake Plav being cleared for grazing < 1650 m a.s.l. (Figures 5.2, 5.4), where types such as, Vetch (*Vicia*), Scabious (*Scabiosa*), bellflower (*Campanula*) and clover (*Trifolium*) exist. This vegetation is not indicative of the majority of the catchment, were generally a mixture of deciduous and coniferous forests exists. The anthropogenic activity in the area seemed conducive to the diversity of the vegetation in places, with types such as *Verbascum*, only seen around cleared fields (Figure 5.3). Land above the tree-line (1800 – 1900 m) e.g. the Bun i Jezerces site, was used for grazing where alpine meadow was possibly maintained by the grazing sheep. Overall, anthropogenic activity was dominant in the lower catchment and across the whole area humans interacted with the local vegetation. The 'natural' vegetation was not cleared in more remote valleys and upper catchment areas.

Figure 5.3: Various examples of vegetation associated with cleared land in the Lake Plav catchment. (A) *Verbascum*, (B) *Juniperus communis*, (C) *Prunus* trees within zone I, (D) mixed deciduous woodland of zones III and IV.

Vegetation zone	Zone	Altitudinal range (m)
Wildflower and *Poaceae* (WP)	I	850 – 980
WP and tree transition	II	980 – 1000
Fagus with *Quercus* and *Corylus*	III	1000 – 1040
Poaceae with *Fagus* and *Corylus*	IV	1040 – 1060
Fagus/Corylus mix	V	1160 – 1130
Pinus dominates	VI	1130 – 1215
Fagus	VII	1215 – 1250
Fagus/Quercus mix	VIII	1250 – 1370
Fagus with some *Quercus* and *Corylus*	IX	1370 – 1460
Fagus with *Quercus*	X	1460 – 1500
Coniferous forest	XI	1500 – 1650
Fagus transition with *Corylus*	XII	1650 – 1800
Coniferous/alpine scrub transition	XIII	1800 – 1900
Juniperus and alpine meadow/scrub	XIV	> 1900

Table 5.1: Altitudinal range attributed to the different vegetation zones identified across the Lake Plav catchment.

Table 5.1 highlights the dominance of *Fagus* in the catchment, with 7 out of 14 zones containing this type. This dominance adheres to the general vegetation trends in the Balkans outlined by Polunin (1980; see Chapter 3). Within the floodplain of Lake Plav various wildflowers and *Poaceae* dominated as land had been cleared to provide grazing land, construction of buildings and small scale cultivation (Figure 5.4). *Salix* trees occurred throughout the cultivated floodplain area, but were concentrated along river banks < 1000 m a.s.l.. *Fagus* woodland began to dominate the landscape above 1000 m a.s.l., but along disturbed areas such as roadsides, paths or around housing a more open canopy was created with: oak (*Quercus*, deciduous), hazel (*Corylus*), hawthorn (*Crataegus monogyna*), dog/wild rose (*Rosa canina*), buttercups and hellebores (Ranunculaceae), Mullein (*Verbascum*), sowbread/cyclamen (*Cyclamen*), spurge (*Euphorbia*), and liverwort (*Hepatica*). Above 1100 m a.s.l. *Fagus* woodland dominated in areas remote to Plav, but coniferous forests and deciduous types were interspersed particularly around Plav and Gusinje. Coniferous forests in the catchment were similar to *Fagus* woodland in that they existed as a relative monoculture of *Pinus nigra/sylvestris* with associated needles providing the dominate ground cover. Zone X, *'Fagus* with *Quercus',* describes a brief zone where deciduous *Quercus* was present sporadically throughout a predominantly monoculture of *Fagus* forest. Forest/woodland persisted to 1750 – 1800 m a.s.l., the local tree line, where alpine meadow became dominant with some J*uniperus communis* (subsp. nana) specimens. *Abies* and *Picea* dominated the tree line at both Visitor Mountain and the Buni i Jezerces complex, but along the Grnčar valley, *Fagus* woodland lay directly adjacent to alpine meadow/scrub at the tree line. The *Juniperus* and alpine scrub zone varies across the catchment, with some *Juniperus* shrubs dotted above 1900 m a.s.l.. *Poaceae* and short alpine types such as *Ranunculus crenatus* (crenate buttercup), *Genista sericea* (silvery broom) and *Hyacinthella Dalmatica* (like *Hyacinthus*) were present throughout the alpine scrub.

During the vegetation survey a downstream site, close to Murino town, provided vegetation data from a site around 100 m lower than the Lake Plav floodplain (Figure 5.5). This was to ensure that major changes at lower elevations were not overlooked, as these may indicate vegetation that once existed or may exist in the future at the Lake Plav catchment. The lower site formed part of a Quaternary river terrace seen in Figure 5.5. Although, the floodplain vegetation of this site was very similar to that around Lake Plav, the forested areas on the valley side contained sparse evergreen specimens such as *Quercus ilex* (evergreen Oak).

Figure 5.4: Evidence of cultivation (fruit trees), land clearance and development around Plav town.

Figure 5.5: Downstream of Lake Plav (812 m a.s.l.). The wide meandering tree lined River Lim (outflow of Lake Plav) can be seen in the middle ground.

Maps of the area date from at least the 1980s, thus satellite images provided the most recent and extensive record of vegetation and provided a good base for testing the GIS modelled data. The variety of zones identified during the vegetation survey was not obvious within the satellite images, as the resolution was not high enough to verify vegetation identification. Areas where monocultures persisted in arboreal species were more easily identified. The modelled data in Figure 5.1 was most comparable to satellite imagery in the upper and lower most areas of the catchment. Overall, the tree lines and zones XIV and XIII were in good agreement with the satellite images, but some detail was lost especially around snow covered or shaded high altitude areas which became homogenous in appearance. The differences between the GIS modelled and satellite images are predominantly the level of homogeneity, with the modelled vegetation data suggesting greater variety in the ecological assemblages throughout the catchment. However, the satellite image (Figure 5.2) does highlight the anthropogenic activity throughout the catchment with evidence of woodland clearance and fenced plots of land across the floodplain as seen in Figures 5.3, 5.4. The modelled data provides a general interpretation of the vegetation in the catchment without mapping the woodland clearances across the area. Sections of cleared land were more apparent within the satellite imagery, but these overlooked woodland management practices such as coppicing which creates a denser canopy. The GIS map and satellite images should be used in conjunction as the former provides greater detail on the variety of vegetation, but generalises altitudinal ranges without accounting for clearance events. Satellite imagery provides a good record of anthropogenic management, but changes in high altitude snow covered or shaded areas are lost.

5.2 AGE DEPTH MODELS FOR THE CATCHMENT CORES

Three cores were extracted for this project, see Figure 5.6. Two from the lower catchment, cores LPCA (base 8.3 m) and LPCB (2 sections: (i) base 7.5 m, (ii) base 27.4 m) and a single core from the upper catchment, core BJC1 (base 1.45 m). AMS radiocarbon and short-lived radionuclide techniques (^{210}Pb, ^{137}Cs) were used to develop a chronology for these cores. Table 5.2 presents all dating results and AMS radiocarbon dates for each core were calibrated using CALIB 5.1.0 BETA and the INTCALO4 dataset (Reimer *et al.*, 2004). AMS radiocarbon dates are presented as the calibrated range for the 2σ conventional radiocarbon age using AD 1950 as the present (0 yrs BP) and in AD/BC to allow comparison to the lead-210 (^{210}Pb) and caesium-137 (^{137}Cs) data (presented in AD).

Core	Depth (cm)	Dating method	Lab code	Date determined	Un-calibrated age	2σ calibrated age range
LPCA	760	^{14}C	RWLPC1760	110.4 +/- 0.4 pMC	110.4 +/- 0.4 pMC	Modern
LPCA	13.5	^{137}Cs	RW/LPCA33	AD 1986	n/a	n/a
LPCA	17.5	^{137}Cs	4A	AD 1963	n/a	n/a
LPCA	25.5	^{137}Cs	9A	AD 1945	n/a	n/a
LPCA	65.5*	^{210}Pb	RW/LPCA16	AD 1887	n/a	n/a
LPCB	2.5	^{137}Cs	1B	AD 1986	n/a	n/a
LPCB	7.5	^{137}Cs	2B	AD 1965	n/a	n/a
LPCB	17.5	^{137}Cs	4B	AD 1945	n/a	n/a
LPCB	540	^{14}C	RWLP5401	AD 1575	270 ± 40 yrs BP	470 - 280
LPCB	2675	^{14}C	RWLP2675	BC 3275	4820 ± 40 yrs BP	5320 - 5130
BJC1	113	^{14}C	RWBJ113	BC 1750	3450 ± 40 yrs BP	3740 - 3660
BJC1	4.5-9.5	^{137}Cs	RW/BJ 5-10	AD 1986	n/a	n/a

Table 5.2: Dates for sediment cores for cores extracted in the Lake Plav catchment. *Base of dated ^{210}Pb profile.

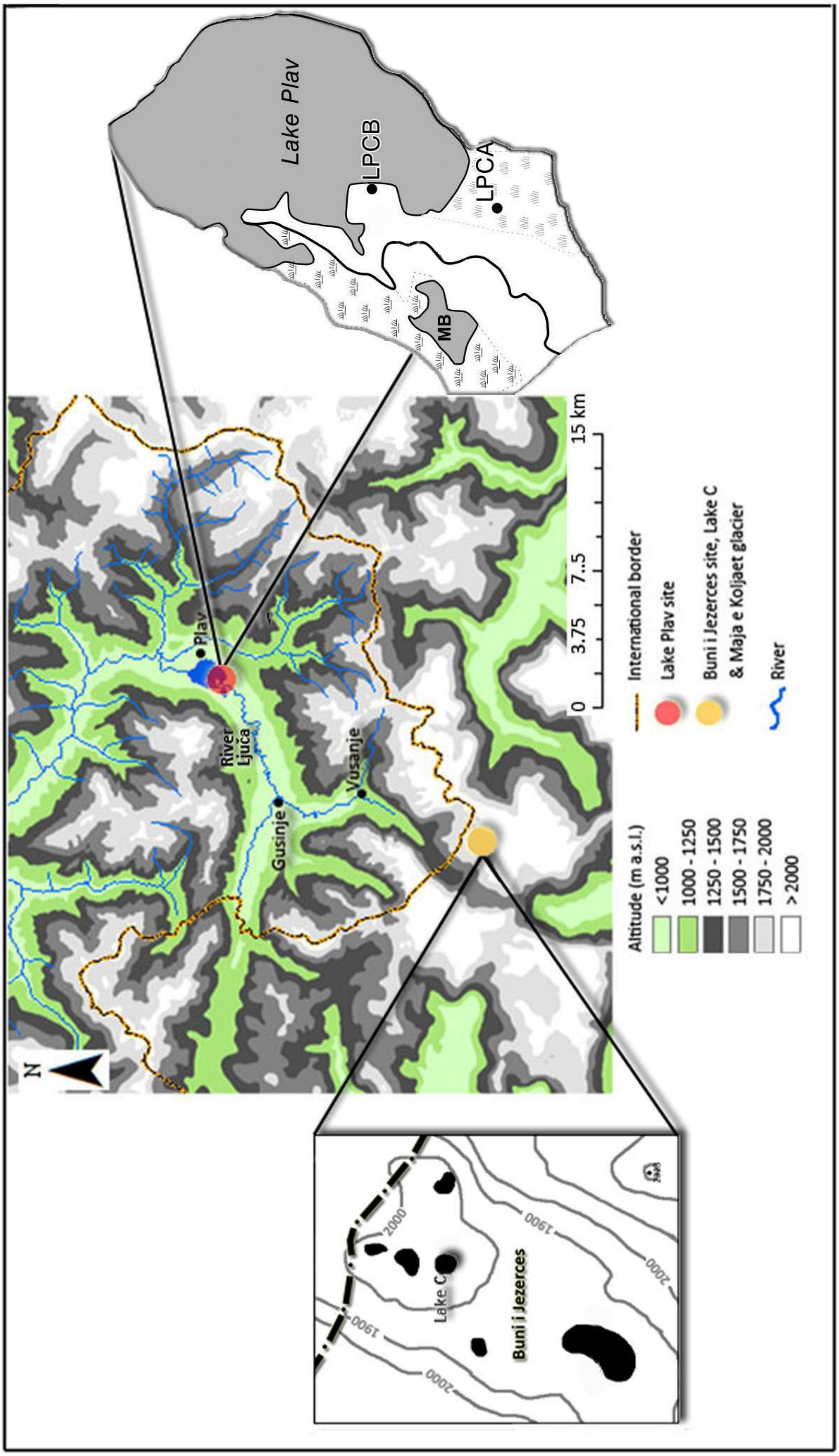


Figure 5.6: Locations of core extractions within the Lake Plav catchment.

5.2.1 Dating results for core LPCA

Material extracted from core LPCA at 7.6 m (RWLPC1760) was suggested to be of a modern age, i.e. post-1950. The material contained a higher 'percentage of modern carbon (pMC)' than the reference standard used by BETA Analytic - referred to as 100pMC i.e. 0 yrs BP or AD 1950. Beyond AD 1950 thermonuclear testing during the 1960s added 'bomb carbon' into the atmosphere producing values greater than 100pMC and levels of subsequent radiocarbon have been reconstructed indicating that during the 1980s levels reached 114 pMC (Genty and Massault, 1999). This gradually decreased to the present day level ***c.*** 106 - 107.5pMC (Genty and Massault, 1999; Gilbert *et al.*, 1999; Moreton *et al.*, 2004; Beta analytic, 2009). Sample RWLPC1760 contained 110.4 ± 0.4 pMC, which lies between the AD 1980 and present day values (see Table 5.2). This suggests that the organic material extracted from core LPCA was most likely deposited between AD 1980 and the present day. If this date is seen as a true reflection of the sediment age at 7.6 m, it would indicate an annual sedimentation rate of ***c.*** 29 cm yr^{-1}. Compared to further dates obtained from core LPCA and those from core LPCB, this seemed unrealistic and therefore, this date has not been used in the age-depth model. Instead two dating methods used for modern samples were pursued and both ^{210}Pb and ^{137}Cs were used in conjunction; as previous studies suggest these dating techniques are complementary with ^{137}Cs providing validation of the model-determined ^{210}Pb results by way of marker date/s (e.g. Carroll *et al.*, 1999a, b; Xiang *et al.*, 2002; Doner, 2003; van Welden *et al.*, 2008; Appleby, 2008).

Lead-210 results for core LPCA

The supply of ^{210}Pb to lake sediments is split into two categories; (i) atmospheric fallout ($^{210}Pb_{excess/unsupported}$) and (ii) in situ decay of natural ^{226}Ra to ^{210}Pb. However, a number of factors influence the supply of ^{210}Pb to lakes such as: atmospheric flux, transportation from around the catchment, the fraction of radionuclide attached to sediment particles, mean particle settling velocity, variations in sediment ^{210}Pb as a result of sediment diagenesis and post depositional processes including sediment focusing and mixing (Carroll *et al.*, 1999a; Appleby, 2008). Another suggested influence upon supply is the residence time within water bodies before deposition. Krishnaswamy *et al.* (1971) discussed the efficient deposition of ^{210}Pb within freshwater and as Lake Plav is a freshwater lake it seemed conducive to efficient ^{210}Pb deposition. Bioturbation and sediment mixing is easily identified within ^{210}Pb profiles as they become rather homogenous in activity (Crusius and Kenna 2007; van Welden *et al.*, 2008). Traditionally ^{210}Pb curves have been interpreted following the assumption of cumulative accumulation and depending on the determined effects of sedimentation processes different models can be used. Models such as the sediment isotope tomography (SIT) method, assumes that sediment accumulation rates and isotopic fluxes both vary with time (Carroll *et al.*, 1999b). Other models available are the constant rate of supply (CRS) model (Tobin and Schell, 1988 in Sikorski and Bluszcz, 2003), constant initial concentration (CIC) model (e.g. Oldfield *et al.*, 1978; Appleby, 2008) and constant flux/constant sedimentation (CF/CS) model (O'Reilly *et al.*, 2010). The CF/CS model assumes that the initial concentration of $^{210}Pb_{excess}$ i.e. unsupported ^{210}Pb from atmospheric fallout will remain the same throughout each sediment layer and that the sedimentation rate remained constant during the period of study (Krishnaswamy *et al.*, 1971; O'Reilly *et al.*, 2010). Figures 5.7 and 5.8 provide evidence of three clusters with similar levels of ^{210}Pb excess at: 65.5 – 40.5 cm (cluster 1), 35.5 – 13 cm (cluster 2) and 10.5 – 0.5 cm (cluster 3). Cluster 1 occurs between two peaks in coarse grained material (see Figure 5.21), thought to be a period of waning sediment supply to the site, which may have caused this plateau then decline. The remaining clusters both coincide with increases in coarse grain material suggested to be periods of increased fluvial activity and sediment supply which may have caused focusing of the sediments at this point. Therefore, the clustering of data may have been focused by the hydrological and sedimentological variations at Lake Plav, causing ^{210}Pb from similar ages to be deposited rapidly at sites during periods of increased sediment deposition. Without any other independent dating to verify this hypothesis the results are interpreted in the traditional cumulative accumulation manner. This is supported by relatively smooth exponential decline produced when $^{210}Pb_{excess}$ profile produced was plotted against accumulated dry mass which was taken as an excellent fit with an r^2 =

0.895. Overall, the CF/CS model produced an average sediment accumulation rate of 0.15142 ± 0.02 g cm^{-2} yr^{-1} with an associated regression analysis providing an r^2 of 0.885. Figure 5.8 highlights the 'baseline' of $^{210}Pb_{excess}$ at 65.5 cm where atmospheric inputs of ^{210}Pb began to be registered, dated to AD 1887 ± 15 yrs using the CF/CS model. Table 5.3 provides dates and sedimentation rates throughout core LPCA which averaged 0.54 cm yr^{-1}.

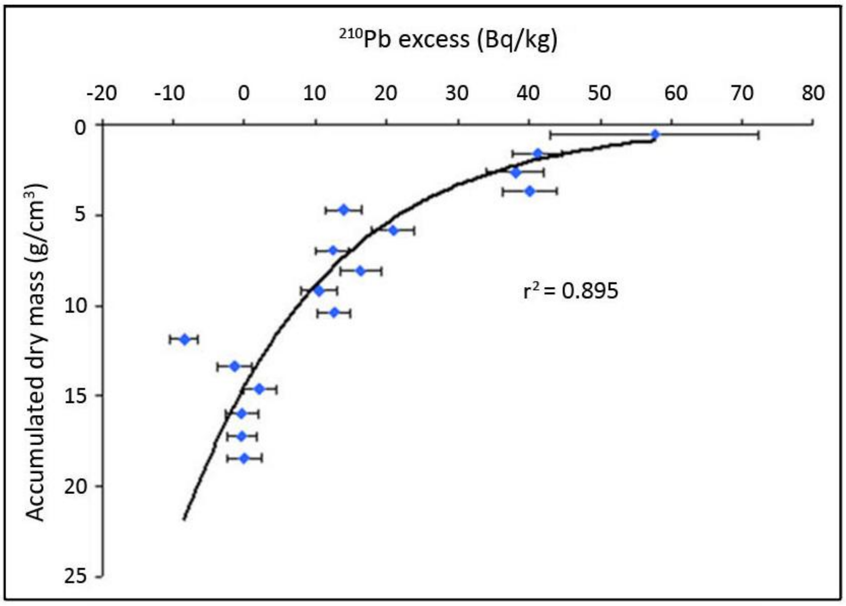


Figure 5.7: Exponential decline of ^{210}Pb.

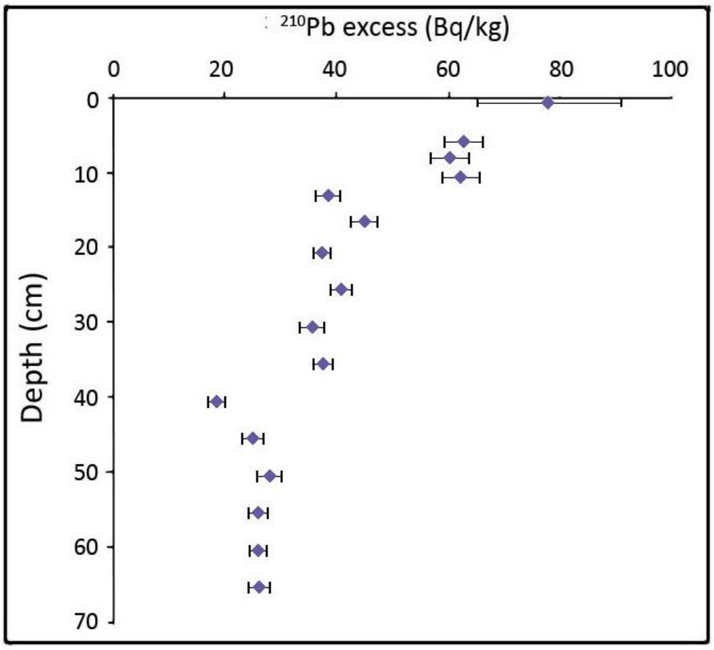


Figure 5.8: Figure identifying $^{210}Pb_{excess}$ plotted against depth for core LPCA.

Sample code	Mid depth (cm)	Date (AD) using CF/CS model	Sedimentation rate ($cm\ yr^{-1}$)
RW/LPCA1	0.5	2006	n/a
RW/LPCA2	5.5	1999	0.79
RW/LPCA3	8.0	1992	0.57
RW/LPCA4	10.5	1985	0.50
RW/LPCA5	13.0	1978	0.46
RW/LPCA6	16.5	1971	0.47
RW/LPCA7	20.5	1963	0.48
RW/LPCA8	25.5	1956	0.51
RW/LPCA9	30.5	1948	0.53
RW/LPCA10	35.5	1940	0.54
RW/LPCA11	40.5	1931	0.54
RW/LPCA12	45.5	1921	0.54
RW/LPCA13	50.5	1912	0.54
RW/LPCA14	55.5	1904	0.54
RW/LPCA15	60.5	1895	0.55
RW/LPCA16	65.5	1887	0.55

Table 5.3: Dates applied to core LPCA using the CF/CS model and the associated sedimentation rates.

Caesium-137 results for core LPCA

Both atmospheric and catchment sources input caesium into lakes (Xiang *et al.*, 2002), similar to ^{210}Pb input. Once in contact with soil caesium associates readily with clay particles, which enter a lake system via erosion or surface runoff (Ilus and Saxén, 2005).

The ^{137}Cs profile produced from analysis of core LPCA (Figure 5.9) suggests the initiation of thermonuclear testing occurred between 25.5 – 20.5 cm, indicated by rising activity. The continued use of thermonuclear testing (1950s to 1960s) is then thought to be represented between 20.5 – 16.5 cm, but it is difficult to confidently date the peak in thermonuclear testing (AD 1963) without further analysis at a finer resolution. In further studies the use of other radionuclides associated exclusively with this period e.g. americium-241 (^{241}Am; van Welden *et al.*, 2008) or iron-55 (^{55}Fe; Mackenzie, 2000) would allow this period to be more confidently constrained.

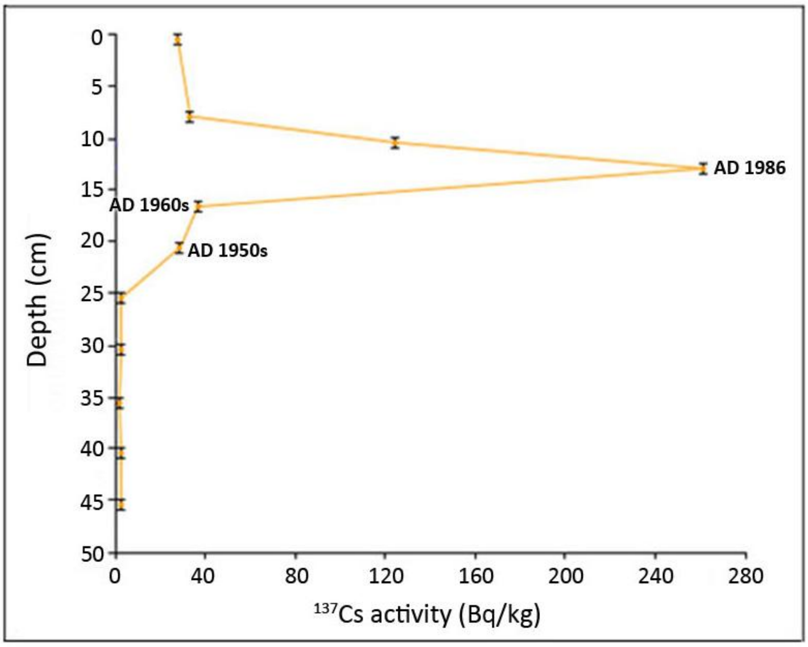


Figure 5.9: ^{137}Cs curve for core LPCA.

As Chernobyl fallout has been registered across Europe and the Eastern Mediterranean e.g. in Northern Greece (Simopoulos, 1989) it is not unreasonable to assume that ^{137}Cs fallout reached the Lake Plav site especially as the nearly 7-fold ^{137}Cs increase at 13.5 cm is characteristic of the disaster (e.g. Pinglot and Pourchet, 1995; Jensen *et al.*, 2002; O'Reilly *et al.*, 2010). Unlike the 'bomb spikes', ^{137}Cs activity associated with the Chernobyl disaster rises and falls more abruptly. Sedimentation rates using the 'bomb spikes' and 'Chernobyl spike' are 0.40 $\pm$ 0.04 cm yr^{-1} and 0.65 $\pm$ 0.1 cm yr^{-1}, respectively. This assumes that the spike in activity at 13.5 cm could represent a date from AD 1982 -

1988. Overall, the average sedimentation rate calculated using the ^{137}Cs activity was 0.525 $\pm$ 0.125 cm yr^{-1} which was in good agreement with the average sedimentation rate suggested by the ^{210}Pb modelled data (0.54 cm yr^{-1}).

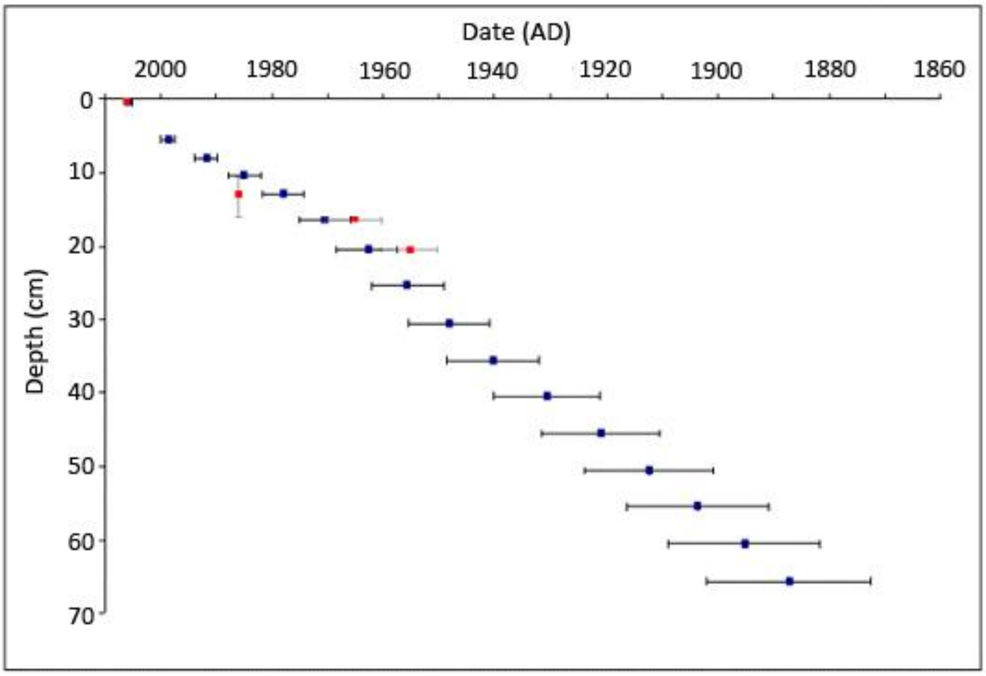


Figure 5.10: ^{210}Pb (blue square) and ^{137}Cs (red square) dating results.

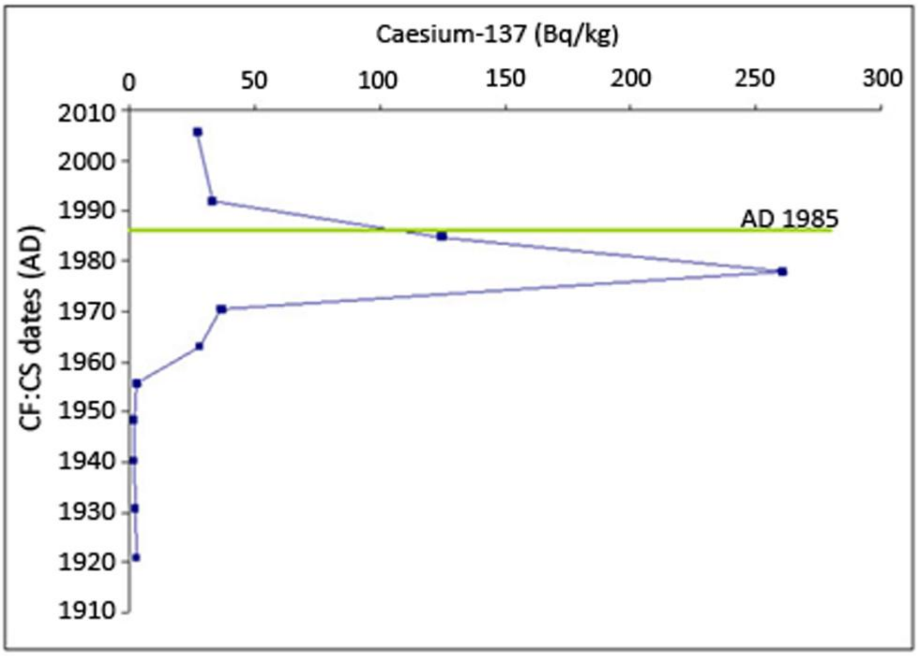


Figure 5.11: ^{137}Cs activity plotted against the modelled ^{210}Pb dates (Green line identifies a modelled ^{210}Pb date).

Age depth model for core LPCA

The ^{210}Pb and ^{137}Cs data for core LPCA are plotted together in Figure 5.10, highlighting variation in dates assigned to each sample. Average sedimentation rates were similar at 0.54 cm yr^{-1} to 0.524 cm yr^{-1}, with both suggesting a recent increase in the rate of sedimentation during the late 20th – 21st century. Comparison of the modelled ^{210}Pb dates and ^{137}Cs marker dates provides a generally good agreement between the two dating methods. The error associated with the earliest rises in ^{137}Cs activity overlaps that of the modelled ^{210}Pb data, but this may be a consequence of sampling resolution or core stratigraphy. Sediments between 70 – 27 cm had a small fraction of fine sand and as ^{137}Cs preferentially bonds to finer particles the level of caesium may have been underestimated, suggesting a younger date. The ^{210}Pb dates lag that of the ^{137}Cs markers by a maximum of 8 yrs at 13.5 cm with ^{210}Pb results suggesting sediments at 10.5 cm would have been laid down ***c.*** AD 1985 (see Table, 5.3; Figure 5.11). A similar lag was identified by Doner (2003) during a study of North-West Iceland where ^{210}Pb dates lagged ^{137}Cs bomb-spike ages by 4 – 10 yrs. Doner (2003) analysed similar clay and silt rich lake sediments to the Lake Plav site and accepted the lag between ^{210}Pb and ^{137}Cs under the assumption that no ^{137}Cs had migrated through the profile. With little homogeneity within the ^{137}Cs data (see Figure 5.9) it is reasonable to assume no migration has occurred within core LPCA and that the lag is a product of the distinct analyses and thus the ^{137}Cs peaks are used to verify and correct the modelled ^{210}Pb data (Figure 5.10). As no further radionuclide analysis was possible due to monetary restrictions the lead of other researchers was taken, and the 'Chernobyl spike' was used as an independent radionuclide marker (e.g. Carroll *et al.*, 1999a; Jensen *et al.*, 2002; O'Reilly *et al.*, 2010; Figure 5.9). The previous 'bomb spikes' were not used as markers as the deeper ^{210}Pb results seemed to correspond well to the remaining ^{137}Cs data (Figure 5.10).

The final age-depth model for core LPCA is presented in Figure 5.12 and indicates that sediments at 65.5 cm are dated to AD 1887 $\pm$ 15 yrs with the rate of sedimentation following that of Table 5.3 until AD 1986 when the rate of sedimentation increases to 0.675 cm yr^{-1}.

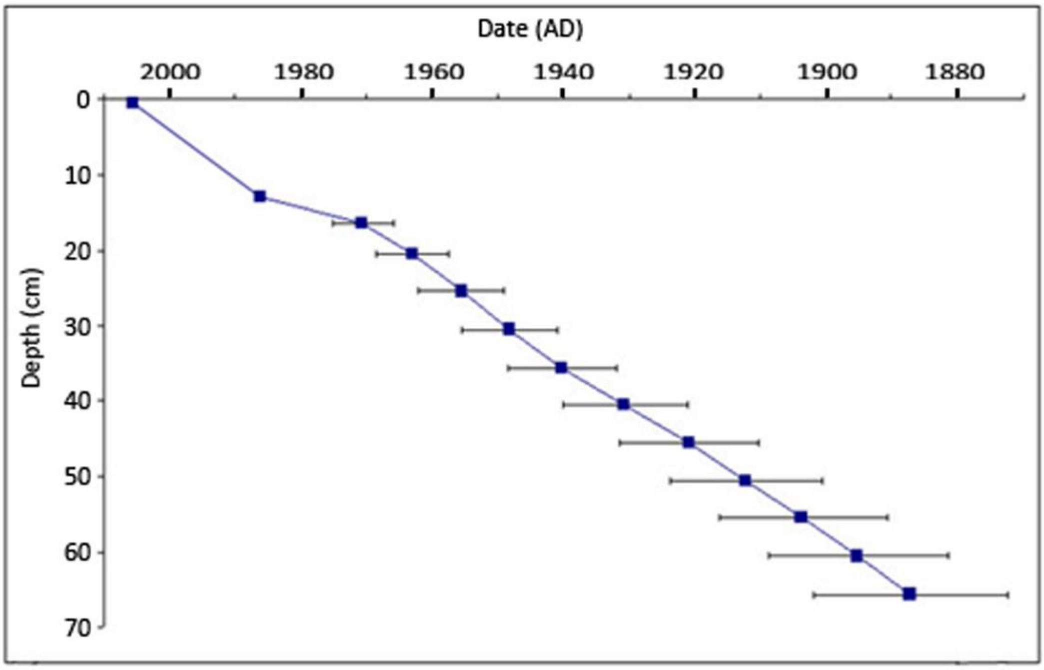


Figure 5.12: Age-depth model for core LPCA.

5.2.2 Dating results for core LPCB

The dating analyses undertaken on core LPCB included ^{137}Cs measurements in the upper sections of the core and radiocarbon dating for the lower sediments. AMS radiocarbon measurements for core LPCB are presented in Table 5.2, represented in terms of AD and BC; 540 cm dated to AD 1575 $\pm$ 40 yrs (375 $\pm$ 40 yrs BP) and 26.75 m dated to 3275 BC $\pm$ 40 yrs (5225 $\pm$ 40 yrs BP). These data suggest that sedimentation rates rose 3-fold between the mid-Holocene to late 1500s from 0.44 cm yr^{-1} to 1.41 cm yr^{-1}, possibly as a result of the LIA deterioration. Figure 5.13 plots the radiocarbon results alongside a value of AD 2008 for 0 cm, representing the surface layer of sediment and the date of core extraction. A polynominal regression was then applied, which gave an r^2 = 0.999, suggesting that the data is non-linear as the regression produces a near perfect fit.

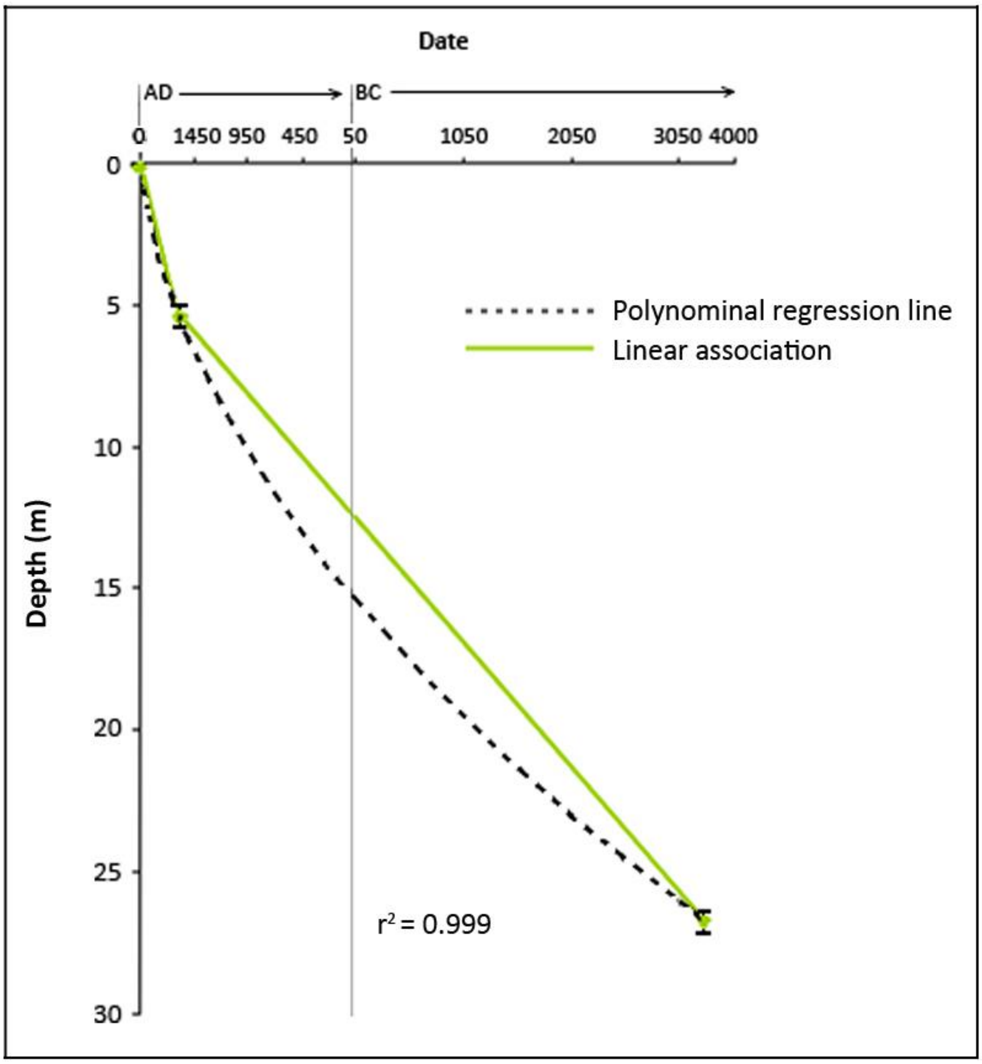


Figure 5.13: Polynominal regression of radiocarbon dates collected from core LPCB.

The ^{137}Cs results for core LPCB (Table 5.2; Figure 5.14) provided a means of calculating recent sedimentation rates. Sample ranges for core LPCB are broad compared to those of core LPCA, due to the core being extracted in 5 cm sections on site. Evidence for the initiation of thermonuclear testing is identified in sample 3B (10 – 15 cm) indicating sediments ***c.*** 60 years old i.e. 1950s. The following sample is likely to include the 'bomb spike' of AD 1963 and an AD 1960s date is applied. Finally, sample 1B (0 – 5 cm) is thought to include the activity peak associated with the Chernobyl disaster as ^{137}Cs activity rises to 198.94 Bq/kg, similar to the activity levels seen throughout Europe as previously mentioned. Although this level of activity is below that of core LPCA at the same time this is thought to be a product of the larger core LPCB sampling range 'diluting'

^{137}Cs activity within the samples. A polynominal regression line was fitted to the ^{137}Cs data, to enable comparison to the radiocarbon dates, producing an $r^2 = 0.9888$, indicative of a very good non-linear correlation between the data (Figure 5.14). A linear regression was also fitted, but produced a poorer r^2 value of 0.929. Finally, the ^{137}Cs dates were plotted with the most recent radiocarbon date for the core and application of a polynomial regression produced an r^2 value of 1. These data suggest that between AD 1575 - 1945 sedimentation rates slowed from 1.41 $\pm$ 0.17 cm yr^{-1} to 0.277 $\pm$ 0.03 cm yr^{-1}; the lowest rate seen at the site since the mid Holocene. The rate of sedimentation has continued to slow to a possible minimum of 0.11 cm yr^{-1} at the current time, causing rates to fall below those of site LPCA by 0.5 cm yr^{-1} since AD 1986.

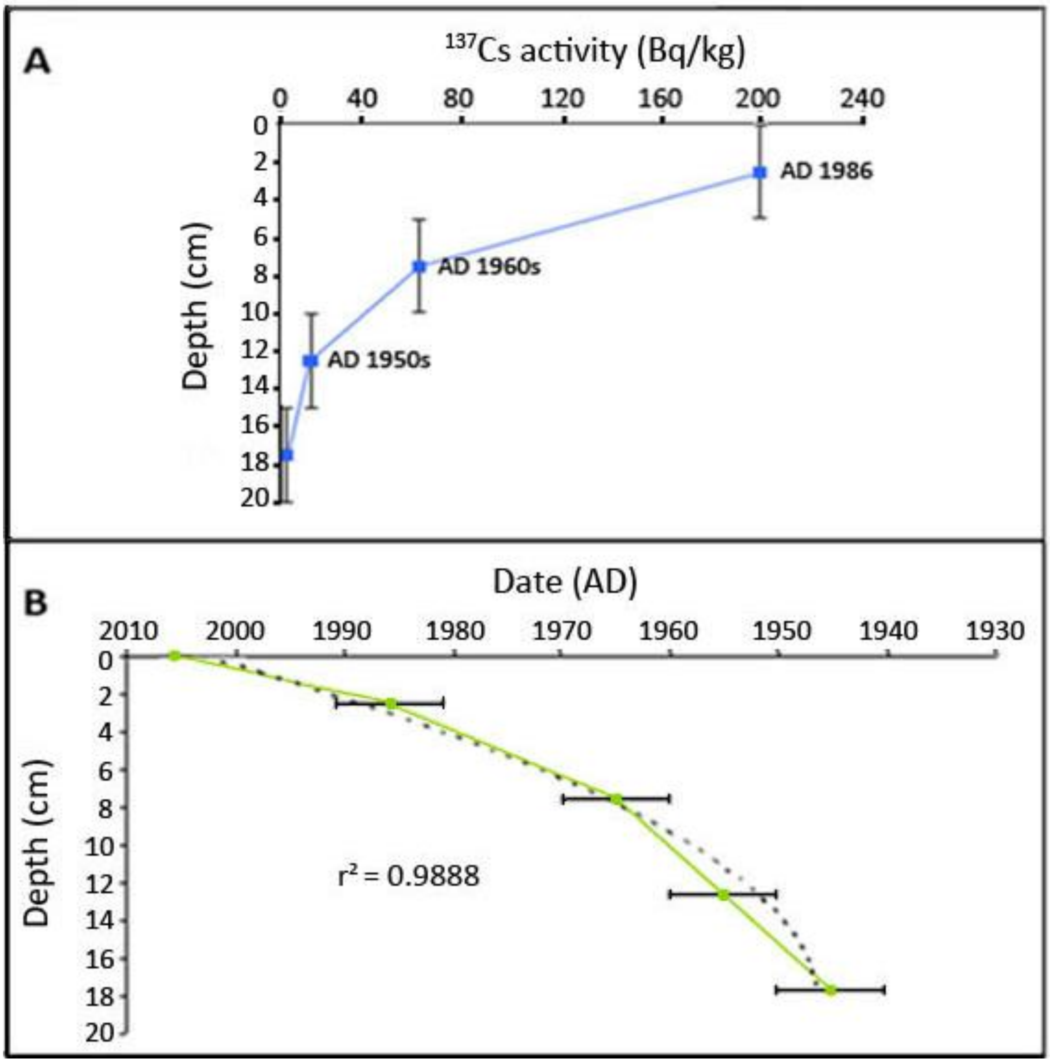


Figure 5.14: (A) ^{137}Cs activity profile for core LPCB. (B) Polynomial regression curve applied to the dated profile.

Age depth model for core LPCB

As both the ^{137}Cs and radiocarbon data for core LPCB suggest good fit with a polynomial regression, the regression lines created were used to produce the age-depth model for the core. The model is presented in Figure 5.15, with the upper and lower sections of the cores separated due to the different scales used. Although, the model provides the basis for dating core LPCB sediments it does not highlight the hiatus between 7.5 – 25.5m within the core.

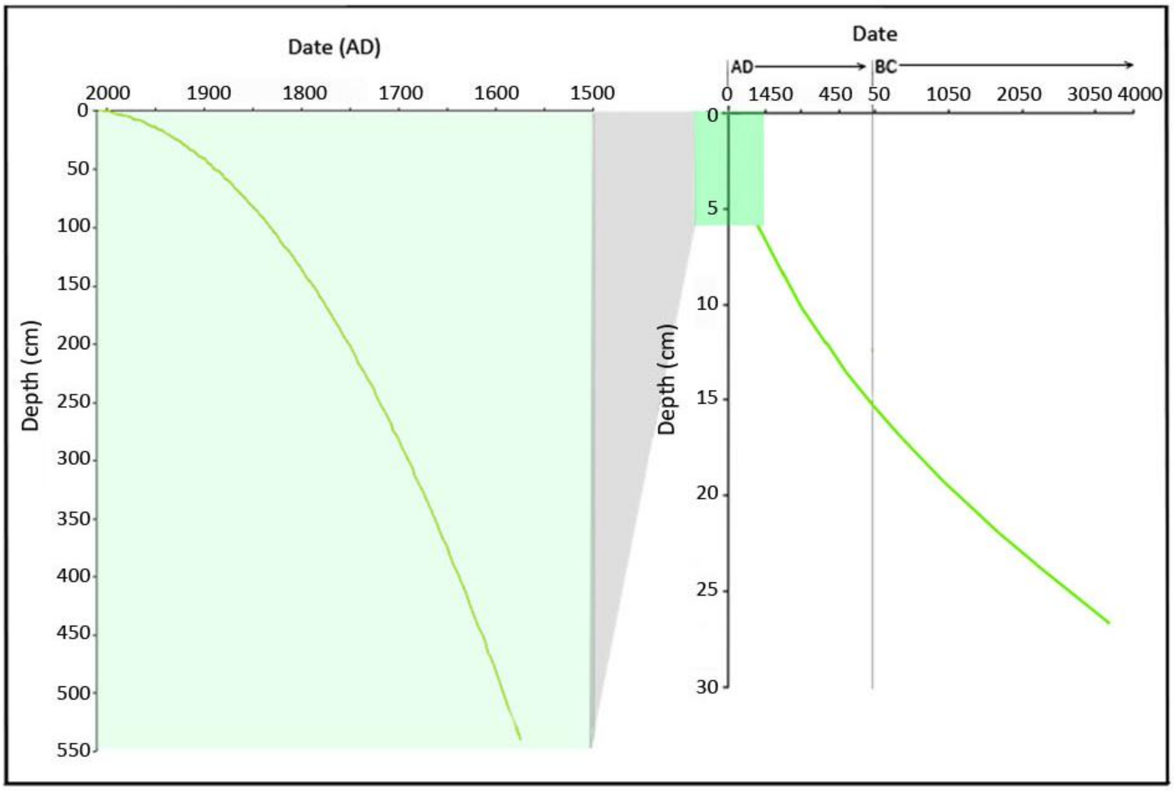


Figure 5.15: Age-depth model for core LPCB.

5.2.3 Dating results for core BJC1

Dating of core BJC1 utilised the same two techniques as that of core LPCB (see Table 5.2). Radiocarbon analysis of an organic fragment extracted at 113 cm thought to be dung rather than an aquatic or terrestrial plant macrofossil (see Appendix VIII). Analysis was undertaken exclusively upon the coarse fraction of the sample, with the aim of minimising any hard water effects that may have occurred. This is a common inaccuracy for samples extracted from a karst lacustrine environment as ‘old’ carbon from dissolved bedrock carbonates has an affinity to build up in lake sediments (e.g. Birks, 2001; Walker, 2005; Hajdas, 2006). An age of 1750 BC ± 40 yrs (3700 ± 40yrs BP) at 113 cm was provided (see Figure 5.16), indicating sedimentation rates at the site were 0.0305 ± 0.003 cm yr^{-1}. Although this is nearly four times slower than the present lower catchment rates, it is comparable to other high altitude lakes in the Eastern Mediterranean region. Sedimentation rates at Preluca Tiganului (730 m a.s.l.), Gutaiului Mountains in North-Western Romania and Lake Besbog, Bulgaria have been ~0.05 cm yr^{-1} throughout the Holocene (Björkman *et al.*, 2002; Stefanova *et al.*, 2006a). Similar rates were identified at Lake Ohrid (693 m a.s.l.) on the Macedonia/Albania border and Lake Maliq in Albania at ~0.07 cm yr^{-1} during the late Holocene (Wagner *et al.*, 2008; Denèfle *et al.*, 2000) and both lakes had a larger catchment area than Lake C. Therefore, the rates seen at Lake C are thought to be comparable to other sedimentation rates around the Eastern Mediterranean during the late Holocene.

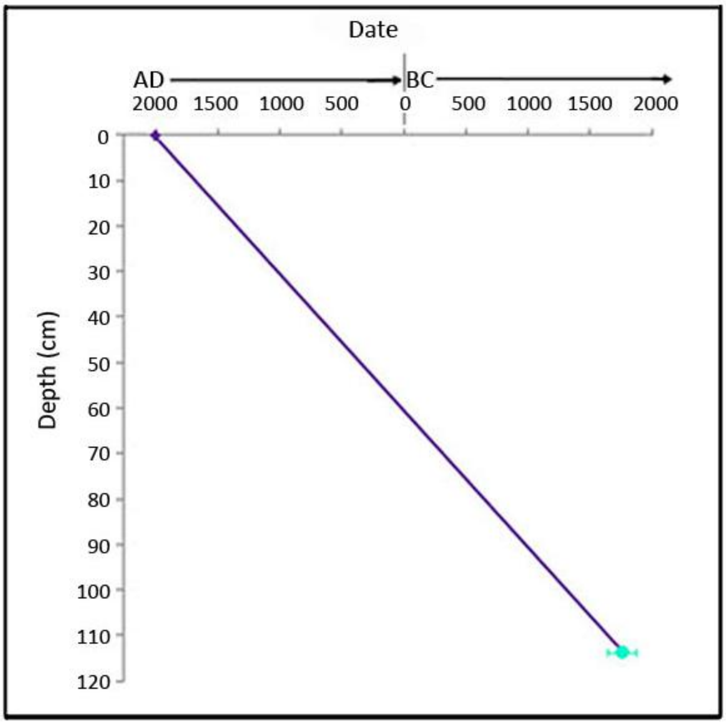


Figure 5.16: Radiocarbon profile for core BJC1, the AMS radiocarbon date is represented by a turquoise circle and the present day by a purple diamond.

Caesium-137 dating for core BJC1

As previously stated the identification of spikes within the ^{137}Cs record is used to assess the date at which sediments were laid down. The record produced for core BJC1 (Figure 5.17) was very different to the lower catchment records. Activity levels reached 446.53 Bq/Kg at 4.5 ± 0.5 cm (RW/BJ5) and remained at 395 – 430 Bq/Kg until the final sample at 9.5 ± 0.5 cm (RW/BJ10). This decline is not linear, with sample RW/BJ8 (7.5 ± 0.5 cm) registering a lower level of activity than the two samples beneath it. The record is indicative of post-depositional mobility, which may have been exacerbated by the predominantly silty nature of the sediments, as ^{137}Cs readily associate with smaller clay particles (Ritchie and McHenry, 1990). This may have caused the ^{137}Cs to be more mobile within the profile, producing this broader record of radionuclide activity. Researchers have indicated a variety of reasons for the apparent post-depositional mobility of ^{137}Cs through a sediment profile, including: bioturbation (Robbins *et al.*, 1977; Fisher *et al.*, 1990), delayed inputs of ^{137}Cs from a catchment (Miller and Heit, 1986) and competition of ^{137}Cs ions with Na^+ and K^+ within sediments (Longmore *et al.*, 1986; Foster *et al.*, 2006). A tentative date of AD 1986 is applied at 4.5 ± 0.5 cm (RW/BJ5) as this sample registers the highest level of activity in the profile at > 400 Bq/kg, the maximum associated with the Chernobyl disaster in European samples (e.g. Pinglot and Pourchet, 1995; Jensen *et al.*, 2002; O'Reilly *et al.*, 2010). The high level of ^{137}Cs compared to the lower catchment cores, is thought to be the result of Lake C being far smaller with little or no currents as it appears stagnant, which was shown by Shukhorukov *et al.* (2000) to concentrate the levels of ^{137}Cs within lake sediments. Without an initiation of activity visible and indication of down core ^{137}Cs leaching a further depth cannot be associated with the early 1950s thermonuclear testing. Overall, this dating suggests that sedimentation rates at Lake C in the Buni i Jezerces complex have been around 0.195 cm yr^{-1} since AD 1986. Uncertainty is associated with this rate due to the level of post-depositional mobility seen within the record and is therefore disregarded for the age-depth model of core BJC1 (see Figure 5.16).

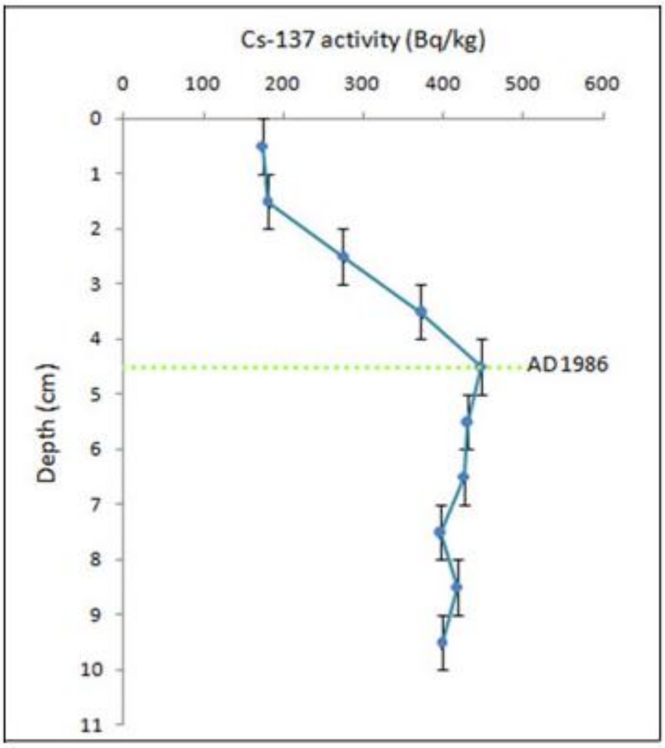


Figure 5.17: ^{137}Cs curve for core BJC1.

Age depth model for BJC1

The age depth model follows Figure 5.16, but when calculating sediment ages a modification is required i.e. the known age of 0 cm is AD 2009 following the date of extraction, but if the zero radiocarbon age were to be used, 0 cm would be assigned an age of AD 1950. This would be contradictory to the sedimentation rates calculated and thus 2 cm was used as AD 1950, as this was felt to go some way to placing sediment ages in perspective especially with ^{137}Cs detected in the uppermost sediments.

5.3 THE LAKE PLAV LOCALITY

A total of eight sediment cores were extracted from the edge of Lake Plav at locations identified in Figure 4.1 (Chapter 4). Six were analysed in the field, whilst the remaining two cores (LPCA and LPCB) were brought back to the UK for multi-proxy analysis. The stratigraphies of these cores are presented in the following sections.

5.3.1 The stratigraphic record

As no previous palaeoenvironmental studies had been conducted at Lake Plav, a transect of cores were used to assess the most beneficial area for sediment coring. The South Eastern area of the floodplain was less affected by seasonal flooding and thought to provide a record where the upper units were less disturbed. Cores extracted along the transect were analysed in the field and a full stratigraphic records of these cores is provided in Appendix IX.

Descriptions of the sediment cores extracted are divided into three sections, with the first detailing the six transect cores and the following describing the sediment cores used for multi-proxy analysis i.e. LPCA and LPCB. Results are based upon field textural analysis for all but core LPCB, which was analysed in the laboratory. Data presented for the core stratigraphies are discussed using the following code system for each unit identified within the stratigraphy: Bxu$^{\text{unit number}}$ (e.g. B1u^{1}, LPCAu^{1}) and presented in Figure 5.18.

5.3.2 Transect cores data

Core 1 (B1) was predominantly sand and divided into 9 units. B1u^{1} was a dark grey (5Y 4/1) medium grained sand (238 - 350 cm) with an abrupt boundary to B1u^{2}. B1u^{2} was composed of finer grained sediments with *Substantia humosa* and woody fragments. An inorganic sand with clay and silt fractions characterised B1u^{3}. B1u^{4} contained white/cream mottled clay silt between 152 - 156.5 cm, but B1u^{5} had greater silt content. B1u^{6} became more clay rich, with brief fine sand alternations. The remainder of the core was predominantly a dark grey clay with silt and fine sand. Fossil molluscs were recorded within B1u^{8} at 55 - 58 cm and the final unit, B1u^{9}, was an organic rich soil with many rootlets and matted grass roots in the upper 2 cm sediment.

B2 was also divided into 9 units the lowermost unit, B2u^{1}, similar to the lower unit of B1. B2u^{1} contrastingly contained wood fragments ~ 5 mm in size at 545 cm. Then units B2u^{1}

to B2u^{4} were dominated by medium to coarse sand and fine grained grit, differentiated by the grain size recorded. The boundary between B2u^{6} and B2u^{7} was sharp, indicative of an abrupt change in depositional environment. B2u^{8} was a very dark grey (2.5Y 3/1) at the base, with a gradual transition to a dark grey (2.5Y 4/1) clay silt with white mottling in the centre of the unit. Organic fragments and staining occurred in the upper units, which remained clay silt (see Figure 5.18). Horizontally bedded organic fragments, possibly *Phragmites*, were recorded within B2u^{8} at 64 - 109 cm. The transition to B2u^{9} was sharp giving way to a silt unit which included small wood fragments.

B3 was unlike the previous cores as dark grey (5Y 4/1) dominated the base of the profile (B3u^{1} - B3u^{2}). Slightly stratified layers were recorded in B3u^{1}, whilst an abrupt coarse sand layer at 283 - 285 cm was recorded in B3u^{2}. The fourth and sixth units were distinct with at least 50% silt, whilst B3u^{7} contained organic staining. By B3u^{7} silty clay dominated, but B3u^{8} consisted predominantly of clay. B3u^{9} was indicative of a more turbulent environment as medium sand dominated with silty loam organic horizons throughout. The final unit consisted of brown-grey clay silt with increasing organic content to the surface (see Figure 5.18).

Clays and silts continued to dominate the sediments in B4. Sparse mollusc remains were found in B4u^{1}, a dark grey (5Y 4/1) stratified silt, suggesting relatively calm lake waters. This was continued into B4u^{2} with molluscs, fine sand and some stratified sediments. B4u^{3} was predominantly silt, before sediments returned to a silty clay with fine sand and molluscs. By B4u^{5} sediments became olive brown-grey in hue and slightly mottled. Occasional horizontally bedded organics were logged in the olive grey clay silt sediments of B4u^{6}, whilst B4u^{7} was characterised by alternations of dark grey organic rich clays with a grey clay silt or black brown detrital peat. By B4u^{8} clay dominated once more and although less frequent and thick, organic rich layers persisted. Overlying these units was a dark grey clayey silt with some orange mottling at 50 – 70 cm. This graded onto B4u^{10} which was a dark olive brown clay silt, similar to the upper units of B1 - B5.

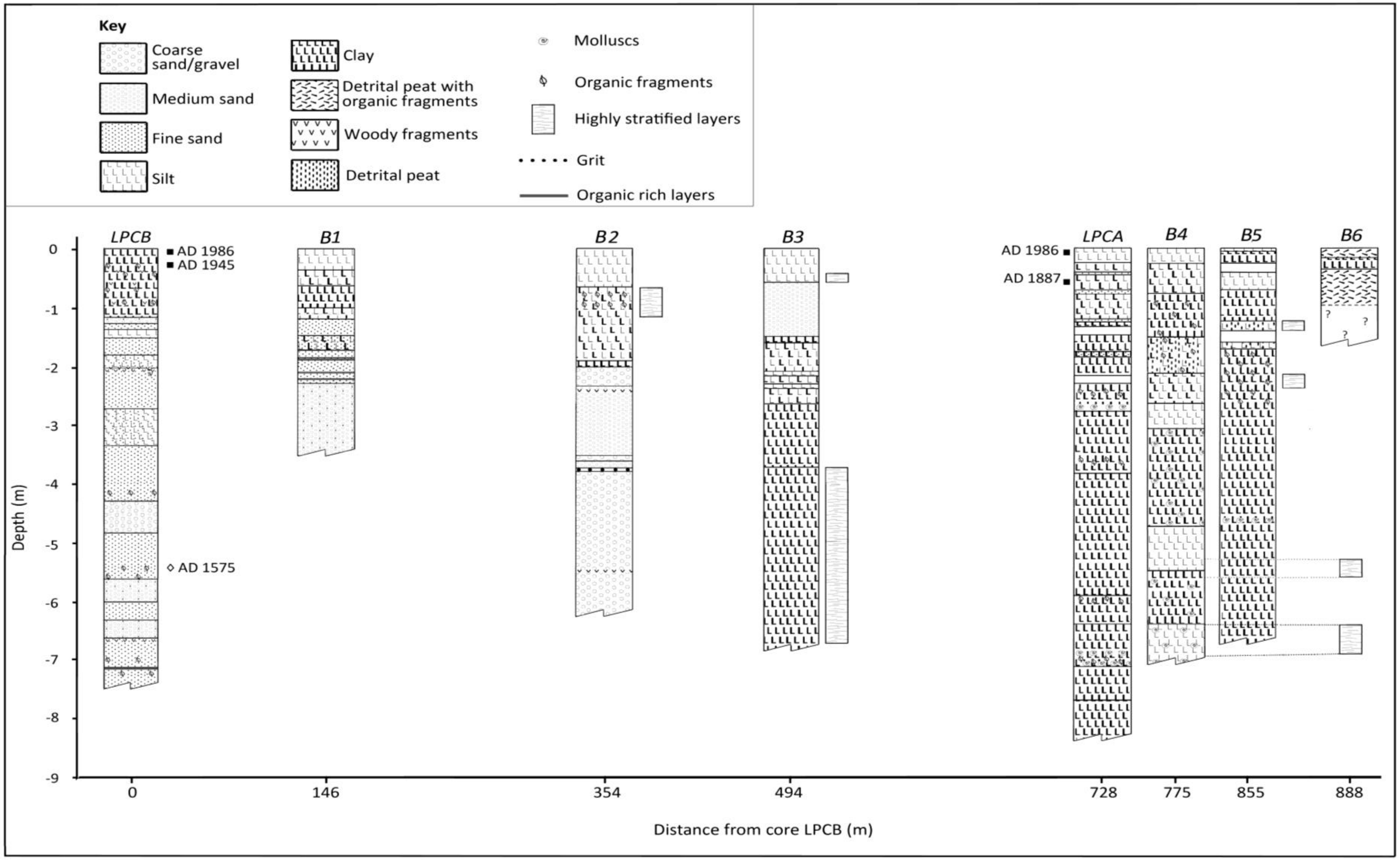


Figure 5.18: Stratigraphic transect of the Lake Plav shoreline using AMS radiocarbon dates (open diamond), ^{210}Pb and ^{137}Cs dates (square blocks).

The penultimate core, B5, was similar to core LPCA with occasional organic and mollusc remains recorded in a predominantly clay matrix. Unlike previous clay rich units, the colour of B5u^{1} was a dark greenish grey, becoming lighter up core. Variations within B5u^{2} included peaty silt at the base of the unit before clay rich and silty sediments returned by 171 cm. Between B5u^{2} and B5u^{3} a hiatus was encountered, but B5u^{3} continued as a stratified silty peat (B5u^{2}). The fourth unit was dominated by clay, but stratified silty peat returned by 133 – 139 cm. A second hiatus was present between B5u^{4} and B5u^{5}. The final sediment unit was characterised by a grey brown silty clay with mottling (2.5Y 4/1), but the uppermost 3 cm was an organic rich clay silt seen at the surfaces of most cores.

Finally, B6 was the most organic-rich core extracted at Lake Plav. This was divided into 4 units and B6u^{1} sufficed as a depth test, indicating the base of the core at 169 cm. A dark brown (2.5Y 3/1) detrital peat composed B6u^{2} with silt /sand fractions and *Phragmites* macrofossils, but silty clay dominated B6u^{3} with the occasional rootlet. Finally, B6u^{4} was a very dark greyish brown (2.5Y 3/2) with many intertwined rootlets and some silt.

5.3.3 Core LPCA stratigraphy

LPCA was divided into 18 units, with the lowermost eight units dominated by a generally dark grey (5Y 4/1) clay. Subtle variations within the stratigraphy are not visible in Figure 5.18, but are apparent in Table 5.4. A minor hiatus occurred at 700 – 719 cm. LPCAu^{3} was more diverse with a small sand fraction, mollusc remains and organic macrofossils. LPCAu^{4} was similar but without a sand fraction. Above, LPCAu^{5}, a dark/olive grey silty clay, contained some small (1 – 2 mm) organic fragments at 580 – 591 cm and 457 – 557 cm. LPCAu^{6} began with an abrupt transition to a brief coarse sand layer at 386 – 387 cm, but then returned to dark grey silty clay similar to LPCAu^{5}. Clay continued to dominate the profile through LPCAu^{7}, with sharp transitions to sediments including mollusc remains and organic fragments (see Table 5.4). Another minor hiatus was encountered, but the sediments above remained a dark grey silty clay (LPCAu^{8}). LPCAu^{9} included three abrupt transitions to: (i) fine sand layer 182 – 184 cm, (ii) clay silt 180 – 182 cm and (iii) fine sand layer 178 – 180 cm. Sediments returned to a clay silt in LPCAu^{10}, but became a lighter. LPCAu^{11} was identical to LPCAu^{10}, but was overlain by a fine sand layer at 120 – 124 cm. The pattern from LPCAu^{13} to LPCAu^{17} was a constant shift between a grey clay silt to fine sand, with all but the LPCAu^{17} (grey/brown, 2.5Y 4/1) being dark grey to grey in colour (5Y 4/1, 5Y 5/1). Finally, LPCAu^{18} had greater organic content with intertwined rootlets and similar to cores B1 to B3.

Core LPCA stratigraphy

Depth (cm)	Description and composition	Boundary	Colour
0 – 27	Brown grey clay silt with intertwined rootlets $Ag^3\,As^1Th^+$	VG	5Y 4/2
27 – 40	Clay:silt with fine sand $Ag^2\,As^2\,Ga^+$	S	
40 – 46	Fine sand Ga^4	S	5Y 5/2
46 – 70	Clay:silt with fine sand $Ag^2\,As^2\,Ga^+$	S	
70 – 78	Fine sand Ga^4	S	
78 – 120	Clay:silt with fine sand $Ag^2\,As^2\,Ga^+\,Sh^+$	G	5Y 6/2 to 5Y 4/2
120 - 124	Fine sand Ga^4	S	
124 – 135	Grey silty clay $As^3\,Ag^1\,Th^+$		
135 – 145	HIATUS		
145 – 178	Grey silty clay $As^3\,Ag^1\,Th^+$		
178 – 180	Fine sand Ga^4	G	
180 – 182	Silty clay $As^3\,Ag^1$	G	
182 – 184	Fine sand Ga^4	G	
184 – 217	Silty clay with fine sand $As^2\,Ag^1\,Ga^1\,Th^+$	G	2.5 Y 4/2
217 – 225	HIATUS		
225 – 386	Dark grey silty clay with organic and mollusc layers: Th^+ 234 – 268 cm, $As^2\,As^1\,Ld^1$ (moll.) 268 – 271 cm Th^+ 358 – 364cm	S	5Y 4/2 to 5Y 5/2

PTO

Table 5.4: Stratigraphic record for core LCPA (continued next page).

STRATIGRAPHY CONTINUED

Depth (cm)	Description and composition	Boundary	Colour
386 – 387	Fine sand Ga^4	S	5Y 5/2
387 – 591	Silty clay with fine sand As^3 Ag^1 Ga^+, with Tb^+ Th^+ 580 – 590 cm	D	5Y 5/2 to 5Y 5/2
591 – 629	Silty clay As^3 Ag^1	D	5Y 5/2
629 – 648	Silty clay with fine sand As^3 Ag^1 Ga^+		
648 – 650	HIATUS		
650 – 700	Clay silt As^1 Ag^2 Ld^1 (moll.) Ga^+ 650 - 660 cm As^1 Ag^2 Ld^1 (moll.) Ga^+ 675 – 700 cm Tb^+ Th^+ 675 – 700 cm	 G S S	2.5Y 4/2
700 – 710	HIATUS		
710 – 776	Silty clay As^3 Ag^1	G	(Mottles 10YR 4/2)
776 – 830	Silty clay with fine sand As^3 Ag^1 Ga^+		

Table 5.4: Stratigraphic record for core LCPA (boundary codes presented in Appendix IX).

5.3.4 Core LPCB stratigraphy

Core LPCB was extracted by the Montenegrin Geological service using a percussion corer enabling extraction of deeper sediments. This led to material being collected to a maximum depth of 27.4 m, but the core was retrieved in two units; the first being 0 - 7.5 m and the second section 25.5 - 27.4 m with an 18 m hiatus between the two sections. This may have been a result of: increased sediment wetness (lower siccitas), the inability to sample such unconsolidated sediments with a percussion corer, a variation in sediment composition rendering sampling difficult, or a combination of these factors. The full stratigraphic record is presented in Table 5.5, but only the upper section is discussed here, see Figure 5.18.

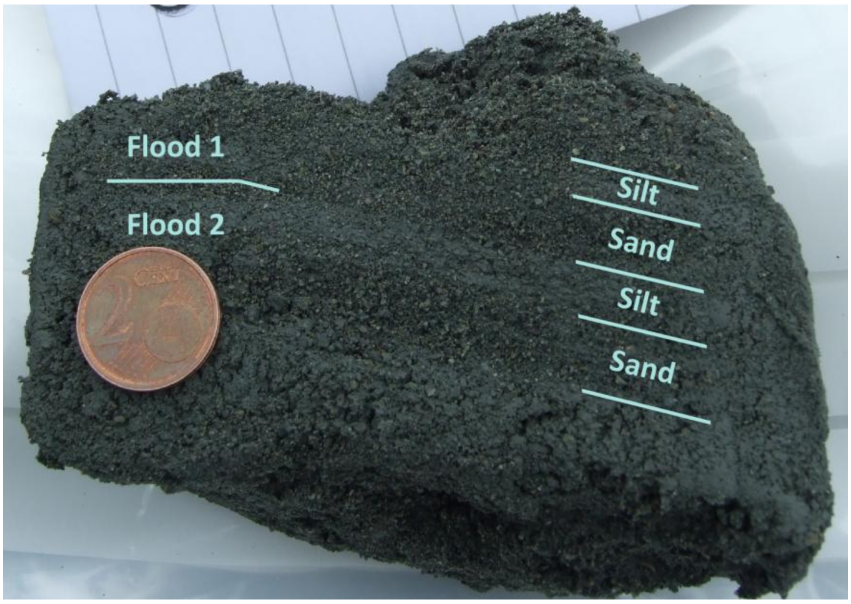


Figure 5.19: Couplets extracted from core LPCB at 740 – 745 cm (Photograph taken by Prof. Jamie Woodward).

Figure 5.20: Possible *Pinus* fragment extracted at 200 cm from core LPCB.

Core LPCB (0 – 7.5 m) was divided into 15 units and overall contrasted to cores C3 to C6 and LPCA, all of which were situated on the *Phragmites* dominated land, as seen in Figure 4.1 (Chapter 4). LPCBu[1] consisted of fine sand material with organic plant (possibly *Equisetum*) fragments and a rich organic layer between 721 – 714 cm that contained dark

(10GY 3/1) soil-like material. Between 745 - 740 cm distinctive flood couplets (seen in Figure 5.19) were identified as a dark sandy unit overlain by a light silty band (see Clague *et al.*, 2003; Chapron *et al.*, 2007). These couplets indicate a period of high energy overbank flow followed by period of settling out (e.g. Desloges and Gilbert, 1994; Wyrwoll and Miller 2001). At the top of LPCBu^{1} wood fragments (2 – 5 mm in length) were found before the sediment roughly changed to equal proportions of fine and medium sand, with a negligible silt fraction (Ag^{+}). LPCBu^{3} represented a return to predominantly fine sand sediments, with minor components of silt and medium sand. The following LPCBu^{4} and LPCBu^{5} contained organic fragments at the boundary of the two units, whilst the sediment composition mirrored that of the two previous units. The composition of LPCBu^{6} was the coarsest material throughout the core, with medium and fine sand dominating in a 1:1 ratio. Following this shift, fine sand is dominant for a relatively long period (see Table 5.5) with some organic fragments recorded at 425 – 420 cm. This suggests that the system returned to conditions seen during unit 5, dated to ***c.*** AD 1500 using an extrapolated radiocarbon date. LPCBu^{8} consisted of equal proportions of silt and fine sand. LPCBu^{9} was again dominated by fine sands, but a coarse sand band occurred within this unit between 310 – 305 cm. The sediments of LPCBu^{10} differ to those of LPCBu^{9}, being predominantly silty and fine sand material. Wood fragments around 2 mm in length were found at the boundary between 206 – 200 cm along with a large 11.6 cm wood fragment. The wood sample was extracted at 200 cm from a matrix of silt and sand and is thought to be *Pinus* (Figure 5.20; Dr Pete Ryan *pers. comm.*). The shifts between lower and higher depositional environments continued in the remainder of the core. Unit LPCBu^{11} was composed of fine sand material, while LPCBu^{12} was primarily composed of silty material. Then LPCBu^{13} and LPCBu^{14} marked a return to fine sand material and a fine to medium sand composition. The final and thickest unit of the core, LPCBu^{15}, was dominated by clay material in contrast to the previous units. Plant and woody fragments were also found throughout LPCBu^{15} which was mottled in appearance. Finally, core LPCB was the only core to be dominated by clay material rather than silt in the uppermost sediments.

Core LPCB stratigraphy

Depth (cm)	Description and composition	Colour
0 – 5	Clay with many rootlets and thick grass on upper surface $Th^1As^3Dh^+$	2.5Y 3/2
5 - 15	Rootlets continue throughout sediment As^3Th^1	↓
15 – 55	Mottling throughout the sediments dominated by clay and small sparse black oxidised wood fragments $As^4Tl^+Th^+$	5Y 4/2 to 7.5YR 4/4
55 – 113	No mottling, but dark woody fragments still occur As^4Tl^+	10BG 4/1
113 – 116	Fine sand Ga^4Tl^+	↓
116 – 134	Fine to medium sand Ga^2Gs^2	
134 – 136	Fine sand Ga^4	
136 – 140	Clay silt with some fine sand $Ag^3As^1Ga^+$	
140 – 150	Silt:Clay with some fine sand $Ag^2As^2Ga^+$	
150 – 189	Silty fine sand Ga^3Ag^1	
189 – 194	Silt:fine sand Ag^2Ga^2	
194 – 200	Sediments dominated by silt but a large wood/branch fragment, possibly *Pinus* recovered $Ag^2Ga^2Tl^+$	
200 – 206	Ligneous material continues but in smaller fragments and within more sandy sediments $Ga^3Ag^1Tl^+$	
206 – 211	Fine sand with some medium sand Ga^3Gs^1	
211 – 216	Silty fine sand Ga^3Ag^1	
216 – 221	Fine to medium sand Ga^3Gs^1	
221 – 226	Silty fine sand Ga^3Ag^1	
226 – 231	Silty fine to medium sand $Ga^3Ag^1Gs^+$	10BG 3/1

Table 5.5 Stratigraphic record for core LPCB (continued next page).

STRATIGRAPHY CONTINUED

Depth (cm)	Description and composition	Colour
231 – 250	Silty fine sand with greater medium sand content $Ga^2Ag^1Gs^1$	5BG 4/1
250 – 255	Silt:sand mix $Ag^2Ga^2Gs^+$	
255 – 260	Silty sand $Ga^3Ag^1Gs^+$	
260 – 265	Silt:sand mix $Ag^2Ga^2Gs^+$	10BG 4/1
265 – 280	Silty sand $Ga^3Ag^1Gs^+$	10BG 3/1
280 – 285	Silty sand $Ga^2Ag^1Gs^1$	
285 – 290	Silt:sand mix $Ag^2Ga^2Gs^+$	
290 – 305	Silt:sand mix Ag^2Ga^2	5G 3/1
305 – 310	Silty sand $Ga^2Ag^1Gs^1$	
310 – 325	Silt:sand mix Ag^2Ga^2	
325 – 335	Silt:sand mix $Ag^2Ga^2Gs^+$	
335 – 340	Silty sand with >2 mm plant fragments $Ga^3Ag^1Dh^+$	10BG 4/1
340 – 360	Silty sand $Ga^2Gs^1Ag^1$	
360 – 395	Silty sand Ga^3Ag^1	
395 – 420	Silty sand $Ga^3Ag^1Gs^+$	
420 – 425	Silty sand with >2mm plant fragments $Ga^3Ag^1Dh^+$	
425 – 430	Silty sand Ga^3Ag^1	
430 – 485	Medium to fine sand with finer fractions $Gs^2Ga^2Ag^+$	
485 – 535	Silty fine sand $Ga^3Gs^1Ag^+$	
535 – 550	Fine sand with some dark organic fragments $Ga^4Ag^+Th^+$	
550 – 565	Silty fine sand $Ga^3Ag^1Gs^+$	
565 – 600	Fine and medium sand Ga^2Gs^2	

PTO

Table 5.5 Stratigraphic record for core LPCB (continued next page).

STRATIGRAPHY CONTINUED

Depth (cm)	Description and composition	Colour
600 – 605	Silty fine sand Ga^3Ag^1	10BG 4/1
605 – 615	Silty sand $Ga^2Gs^1Ag^1$	
615 – 630	Medium to fine sand $Ga^3Gs^1Ag^+$	
630 – 660	Sand $Ga^2Gs^2Ag^+$	
660 – 705	Silty fine sand with limited wood fragments at 685 – 690 cm $Ga^3Ag^1Tl^+$	
705 – 721	Fine to medium sand with soily material at 714 – 721 cm $Ga^3Gs^1Ag^+$	10BG 3/1
721 – 750	Silty fine sand with plant fragments at 740 – 735 cm and couplets 745 – 740 cm $Ga^3Ag^1Th^+$	10BG 4/1
	HIATUS	
25.50 – 25.55 m	Silty fine sand Ga^3Ag^1	5G 3/1
25.55 – 25.60 m	Silty fine sand with some clay $Ga^3Ag^1As^+$	
25.60 – 25.65 m	Silty clay As^3Ag^1	5G 4/1
25.65 – 25.75 m	Silt:clay mix with some fine sand $Ag^2As^2Ga^+$	
25.75 – 25.90 m	Silty clay As^3Ag^1	
25.90 – 25.95 m	Silty sandy clay $As^2Ag^1Ga^1$	5B 5/1
25.95 – 26.00 m	Silty clay As^3Ag^1	
26.00 – 26.65 m	Clay with dark sediments speckled through and molluscs, particularly between 26.25 – 26.35 m $As^4Ag^+Dg^+$	5B 5/1 to 5B 2/1
26.65 – 27.20 m	Clay with organic fragments, molluscs at 26.70 – 26.75m and some mottling $As^3Th^1Dg^+$	5B 5/1 to 5B 2/1
27.20 – 27.40 m	Clay with some organic fragments and molluscs at 27.25 – 27.30 m $As^4Th^+Dg^+$	5B 5/1

Table 5.5 Stratigraphic record for core LPCB (boundary codes presented in Appendix IX).

5.4 CORE LPCA DATA

This section presents a multi-proxy data set for core LPCA, including particle size (PS), loss-on-ignition (LOI), magnetic susceptibility, palynological and ostracoda data.

5.4.1 Particle size analysis

Sediment grain sizes were classified following Wentworth (1922) and Kohnke (1968) cited in Håkanson and Jansson (1983; Table 5.6). This analysis was conducted to explore the significance of particle size variations in core LPCA, providing a high resolution record of PS. The technique aimed to capture changes that may have been overlooked during field observations (see Figure 5.21). Core LPCA was split into 6 PS phases, and the core was dominated by fine grained (< 63 µm), particularly silt material (Figure 5.21). Overall, greater variation in PS values were recorded between phases CAPS III to CAPS VI.

Throughout core LPCA no gravel or grit (> 2000 µm) was identified. CAPS I and II were relatively similar with maximum values of 82.0% and 83.6% silt, respectively (Table 5.7). Greater quantities of fine sand were identified in CAPS I, with between 6.7% and 23.1% recorded (see Table 5.7). Sediments at 695 - 678 cm (***c.*** AD 740 - 770) were characterised by a lack of grain sizes > 250 µm. The dominance of silt material and finer grade sands is indicative of a permanent water body and a site further from littoral or deltaic influences (Hakånson and Jansson, 1983; Digerfeldt, 1986; Stine, 1990). A brief increase in fine sands occurred at 658 - 652cm (***c.*** AD 810 - 820) with values reaching 18.0 – 23.1%, the maximum of CAPS I, before returning to values around the mean, 6.7% (see Figure 5.21, Table 5.7). Medium and coarse sand fractions increase towards the end of the phase, reaching maximum values (see Figure 5.21, Table 5.7). Overall, the phase is dominated by silt sized material with little variation in the quantity of clay, but greater change is visible in the sand records.

Fraction name	Fraction size (µm)
Clay	< 2
Silt	2 - 63
Fine sand	63 - 250
Medium sand	250 - 500
Coarse to very coarse sand	500 - 2000
Gravel/grit	>2000

Table 5.6: Fraction size classification used during particle size analysis, following Wentworth (1922) and Kohnke (1968) cited in Håkanson and Jansson (1983).

CAPS II is characterised by rising quantities of silt at the beginning of the phase (see Figure 5.21), with the mean value of silt material 4.1% greater than that of CAPS I. Quantities of clay during this phase are more variable, with a slightly greater range of values identified than the previous phase (see Table 5.7). Unlike CAPS I a more equal ratio of fine sand to sand > 250 µm was registered. Throughout the phase PS is comparatively stable, with the most obvious changes at the beginning and end of the phase captured in Table 5.7.

The third phase, CAPS III, marks the beginning of fluctuating PS ratios. This is reflected in the greater range of values, particularly in the silt and medium sand fractions (see Figure 5.21; Table 5.7). The silt fraction ranges by 33.5%, suggesting double the variability seen in previous phases. Medium sand values vary by 25.1%, nearly three times that of the previous phase (7.4% in CAPS II). Following the penultimate hiatus, the total sand fraction declined, before a pulse of sand sediments resulted in the silt fraction reaching a low of 46.7%. All sand fractions increased during this phase suggesting higher energy conditions, but still no particles > 2000 µm (2 mm) were identified. In total three pulses of sandy sediments characterised CAPS III at: ***c.*** AD 1550 (250 cm), ***c.*** AD 1630 (205 cm) and ***c.*** AD 1680 (175 cm), with the latter two being the largest.

Phase	Depth (cm)	Clay (< 2 µm) Minimum value (%)	Maximum value (%)	Mean value (%)	Silt (2 - 63 µm) Minimum value (%)	Maximum value (%)	Mean value (%)
CAPS VI	0 - 25	3.7	10.3	5.8	68.0	89.3	75.7
CAPS V	25 - 125	1.2	6.9	4.5	17.0	77.0	57.9
CAPS IV	125 - 170	8.0	10.6	9.1	83.8	88.9	85.6
CAPS III	170 - 265	2.8	7.8	5.0	46.7	80.2	65.2
CAPS II	265 - 390	4.8	8.4	6.7	69.5	83.6	79.5
CAPS I	390 - 830	5.1	7.8	6.9	66.4	82.0	75.4

Phase	Depth (cm)	Fine sand (63 - 250 µm) Minimum value (%)	Maximum value (%)	Mean value (%)	Medium sand (250 - 500 µm) Minimum value (%)	Maximum value (%)	Mean value (%)
CAPS VI	0 - 25	0.4	22.3	11.0	0.0	17.6	6.3
CAPS V	25 - 125	7.0	37.4	21.2	1.6	49.2	14.1
CAPS IV	125 - 170	0.5	4.6	2.8	0.8	1.9	0.9
CAPS III	170 - 265	5.7	20.5	13.0	3.2	28.3	11.2
CAPS II	265 - 390	4.3	13.9	7.7	1.8	9.2	4.0
CAPS I	390 - 830	6.7	23.1	12.7	0.0	7.4	3.7

Phase	Depth (cm)	Coarse sand* (500- 2000 µm) Minimum value (%)	Maximum value (%)	Mean value (%)	Gravel/Grit (>2000 µm) Minimum value (%)	Maximum value (%)	Mean value (%)
CAPS VI	0 - 25	0.0	3.2	1.2	0.0	0.0	0.0
CAPS V	25 - 125	0.0	6.0	2.3	0.0	0.0	0.0
CAPS IV	125 - 170	1.7	3.0	1.6	0.0	0.0	0.0
CAPS III	170 - 265	2.7	12.4	5.5	0.0	0.0	0.0
CAPS II	265 - 390	1.1	4.8	2.1	0.0	0.0	0.0
CAPS I	390 - 830	0.0	2.4	1.3	0.0	0.0	0.0

Table 5.7: Summarised particle size data for core LPCA, *Coarse to very coarse sand.

CAPS IV, beginning ***c.*** AD 1700, is characterised by the highest mean value for clay and silt in core LPCA. These finer fractions dominate the sediments of CAPS IV, with sand fractions registering a minor recovery through the remainder of the phase. All sands registered low values, but the mean values for fine and medium sand are the lowest seen in the record at 2.8% and 0.9% respectively (Table 5.7). The lack of substantial changes during this phase is reflected in the range of values for each fraction, for example, medium sand values vary by 1.1% during CAPS IV compared to 25.1% in CAPS III.

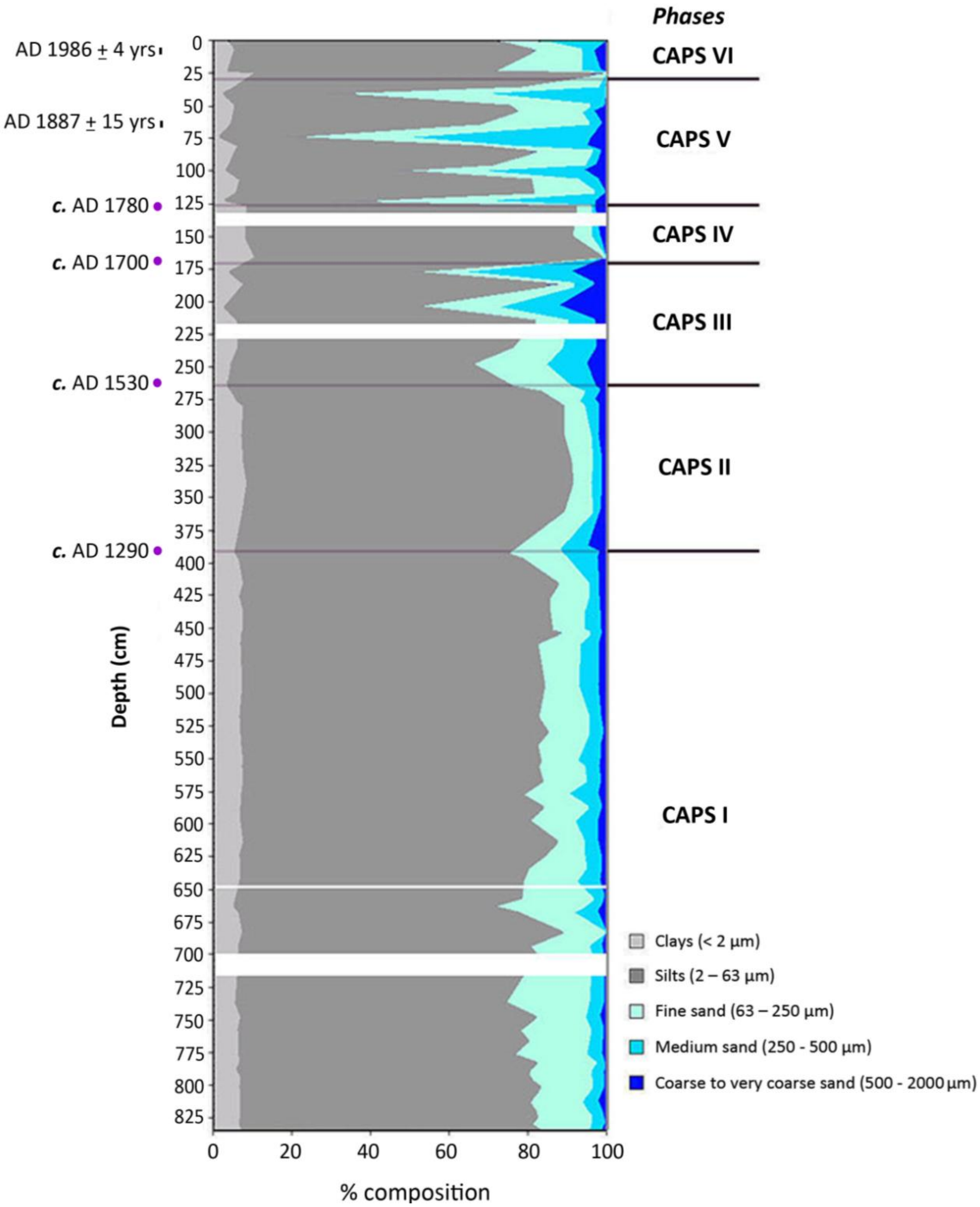


Figure 5.21: Particle size record for core LPCA (black blocks represent ^{137}Cs and ^{210}Pb dates, with purple circles providing extrapolated dates from ^{210}Pb data).

The greatest fluctuations in PS were seen in CAPS V, which is reflected by the minimum and maximum value data presented in Table 5.7. The silt fraction varies by 60%, the greatest range seen in the LPCA PS record. Fine and medium sands compose a greater proportion of the sediments overall, with the highest mean values for the profile (see Table 5.7). Four pulses of sandy sediments can be identified in CAPS V (see Figure 5.21) at: 122 cm (***c.*** AD 1780), 100 cm (***c.*** AD 1820), 81 cm – 73 cm (***c.*** AD 1860 - 1870) and 40cm (***c.*** AD 1930), with the largest pulse occurring ***c.*** AD 1820. Sands totalled 88.9% at this time with more than half being medium sand (250 – 500 µm; see Figure 5.21, Table 5.7). By the end of CAPS V (***c.*** AD 1950,) silts began to dominate the profile once more, reaching 89.3% by CAPS VI.

The final phase, CAPS VI, relates to the past 60 years. Mean values for this phase are most similar to those of CAPS I. However, the range of values recorded for each sediment fraction is greater than those seen in CAPS I. In total ranges within CAPS VI increased by 26% with the greatest increase in range registered within medium sand at 10.2%, but silts remained dominant.

Overall, the PS record is dominated by silt material and indicates little change during between ***c.*** AD 500 – 1530 (CAPS I and II). The latter four phases are punctuated by units with medium and fine sand, indicative of high energy pulses of sandy material. This suggests that site LPCA has been located between the central lake waters and littoral zone since ***c.*** AD 500 (Digerfeldt, 1986; Stine, 1990). The progressive variations in PS during CAPS III to VI (***c.*** AD 1530 - 2006) indicate that the system was influenced by fluvial inputs, as clay sized material (< 2 µm) was never dominant (Hakånson and Jansson, 1983). The sand units suggest the presence of higher energy conditions up core.

5.4.2 Magnetic susceptibility record

The Low frequency (LF) magnetic susceptibility for core LPCA is shown in Figure 5.22. During phase LPCA I the lowest LF magnetic susceptibility is registered, 0.0 $\times 10^{-9}$ m^3 kg^{-1} at 775 cm (***c.*** AD 600). Throughout this phase LF magnetic susceptibility rises, but values in LPCA II generally vary around the mean, 113.0 $\times 10^{-9}$ m^3 kg^{-1} (see Table 5.8). A depression occurs at 460 cm (***c.*** AD 1170) to the lowest value of LPCA II (see Table 5.8; Figure 5.22). LPCA III is characterised by LF magnetic susceptibilities varying by 25 x10^{-9} m^3 kg^{-1} around a mean of 130.8 x10^{-9} m^3 kg^{-1}. Throughout LPCA IV little change is evident, but LPCA IVb covers a hiatus making changes difficult to identify. Values are barely elevated by the end of LPCA IV with a mean of 139.0 x10^{-9} m^3 kg^{-1} for LPCA IVc, compared to 137.6 x10^{-9} m^3 kg^{-1} during LPCA IVa. The penultimate phase LPCA V is characterised by falling levels of LF magnetic susceptibility, with levels dropping to the lowest registered since LPCA II (see Table 5.8). Recent magnetic enhancement is evident during LPCA VI, where LF magnetic susceptibility rises by 85 x10^{-9} m^3 kg^{-1} reaching the highest levels seen throughout the record (Figure 5.22; Table 5.8). This is likely to be the result of recent atmospheric pollution and is a common feature of magnetic susceptibility records.

Magnetic susceptibility

Phase	Depth (cm)	Minimum value ($x10^{-9}$ m^3 kg^{-1})	Maximum value ($x10^{-9}$ m^3 kg^{-1})	Mean value ($x10^{-9}$ m^3 kg^{-1})
LPCA VI	0 - 25	109.5	194.9	135.8
LPCA V	25 - 176	84.6	161.6	109.1
LPCA IV c	176 - 204	129.1	147.2	139.0
LPCA IV b	204 - 235	124	127.9	126.1
LPCA IV a	235 - 265	130.6	143.3	137.6
LPCA III	265 - 374	109.8	155.8	130.8
LPCA II	374 - 616	60.4	129.9	113.0
LPCA I	616 - 830	0.0	143.6	95.5

Loss-on-ignition

Phase	Depth (cm)	Minimum value (%)	Maximum value (%)	Mean value (%)
LPCA VI	0 - 25	8.0	17.0	12.0
LPCA V	25 - 176	2.0	25.0	6.5
LPCA IV c	176 - 204	7.0	9.0	8.4
LPCA IV b	204 - 235	10.0	18.0	13.5
LPCA IV a	235 - 265	7.0	16.0	10.8
LPCA III	265 - 374	8.0	10.0	8.6
LPCA II	374 - 616	7.0	10.0	8.3
LPCA I	616 - 830	8.0	10.0	8.8

Table 5.8: Summarised magnetic susceptibility and LOI results for core LPCA.

In summary, the LF magnetic susceptibility of core LPCA is composed of regular minor fluctuations around the mean value for each phase. Particular declines were recorded ***c.*** AD 600 (775cm), ***c.*** AD 1170 (460cm) and ***c.*** AD 1720 (160cm). Without any significant fluctuations in the record compared to the PS records, the magnetic data do not mirror the variations evident within the PS record. A similar relationship has been observed by Parris *et al.* (2010) in their study of six lakes in the North-Eastern United States (see also section 6.1.4 Magnetics and loss-on ignition data, Chapter 6).

5.4.3 Organic content and loss-on-ignition record

The organic content of core LPCA is relatively constant throughout phases LPCA I to III (see Figure 5.22), with values residing between 6.5 and 13.5% (see Table 5.8). By LPCA IV, LOI becomes more variable, with values varying by 9% during LPCA IVa. LPCA IVb is characterised by an 8% rise in LOI following a hiatus, before levels of organic content fall to around 8.4% during LPCA IVc. Organic content was most variable during LPCA V with a 23% range of values (see Table 5.8). The 25% peak at 110 cm (***c.*** AD 1800) is likely to be the result of sampling the trace of *Substantia humosa* identified in the sediment core between 78 - 120 cm (see Table 5.4). Generally, there is a downward trend in organic content for phase LPCA V to 2%, the lowest value recorded. Organic content begins to rise again from 75 cm (***c.*** AD 1870), reaching 8% at the end of the phase. During LPCA VI LOI values rise to 17% at 20 cm, before a decline and subsequent recovery during the last 20 years. The increase is exemplified by the 15% LOI value for the uppermost sample. Overall, the organic content for core LPCA was around 9.6%, with peaks in LOI occurring during the 1500s - 1600s (LPCA IV), the late 1800s and throughout the 20th century especially in the last 20 years.

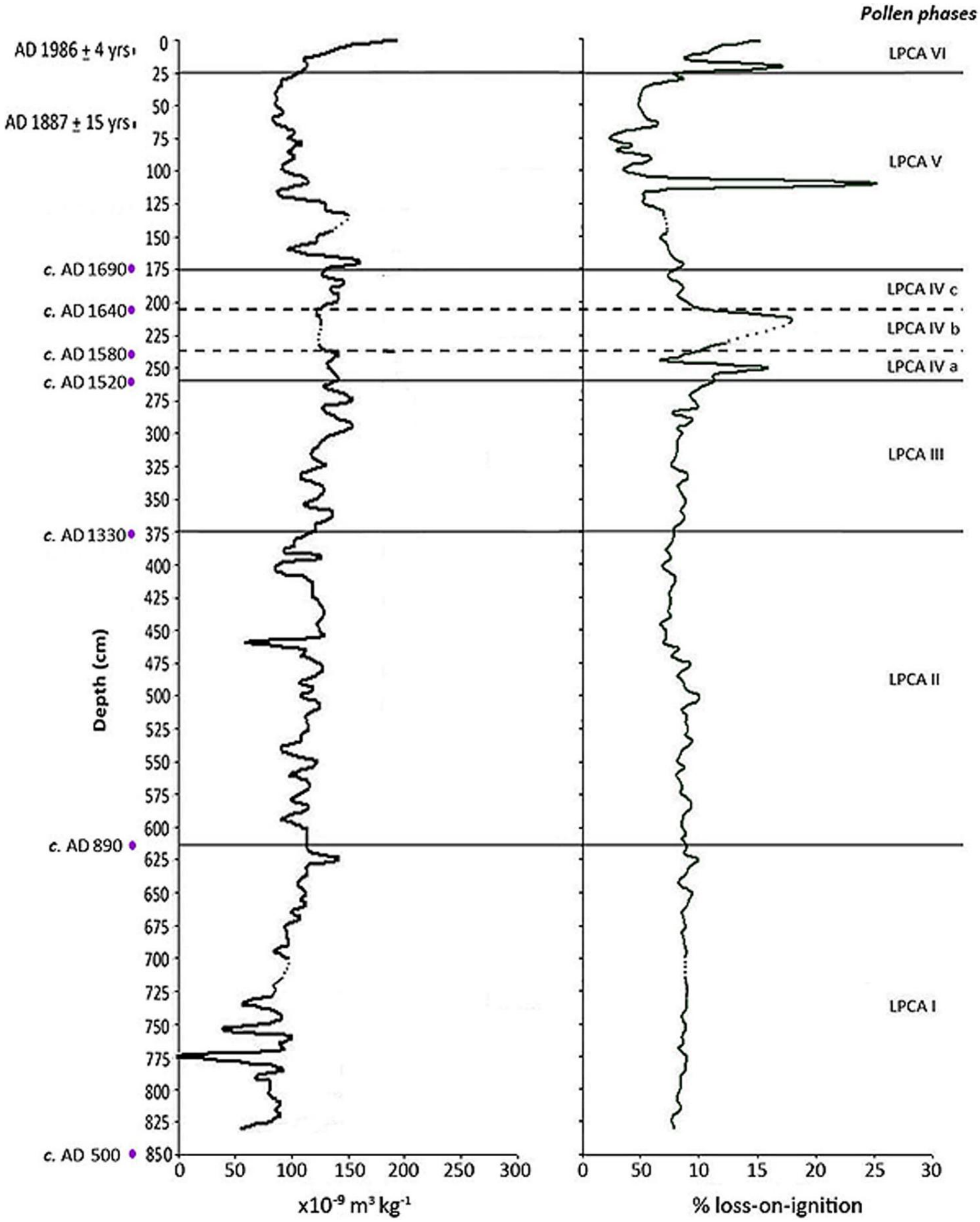


Figure 5.22: Loss-on-ignition and low frequency magnetic susceptibility for core LPCA, Lake Plav. The black blocks represent ^{137}Cs and ^{210}Pb dates, with purple circles showing extrapolated dates from ^{210}Pb data.

5.4.4 Palynological profile

The pollen profile presented in Figure 5.25 was plotted using TGview. The phases assigned to the profile were completed using a combination of visual examination of the profile and the stratigraphically constrained cluster analysis program CONISS (Grimm, 1987; as discussed in section 4.7.4, Chapter 4). A total of 37 dominant pollen taxa (> 2%) were used to produce the dendrogram in Figure 5.23. The pollen profile was divided following the first seven clusters and the dendrogram divisions are plotted against the final zonation in Figure 5.23. This demonstrates that clusters no. 1 and 3 (dotted line Figure 5.23) were used as sub-phases in the final zonation of the profile, dividing phase LPCA IV into three. These were not labelled as individual phases, because they were felt to represent the different initiations of an expansion registered in the aquatics and spores at different times. Furthermore, when these sub-phases were compared to the other proxy data from core LPCA trends were picked out well with the division into LPCA IV a, b and c. The remaining clusters were used to divide the profile into six phases. Overall, the division by CONISS corresponded well to the visual assessment of the profile, with CONISS dividing the initiation of aquatic and spore expansion which corresponds to expansion in herbs associated with lake edge environments i.e. *Equisetum*, *Poaceae*, and Cyperaceae.

84 pollen samples from core LPCA were prepared, but only 67 samples were suitable for palynological assessment with 61 of these providing raw counts >500 grains. Samples with 150 - 300 grain counts occurred between 40 - 180cm, whilst the remaining 17 samples were found to be non-polleniferous (see Table 5.9). A variety of explanations have been put forward in the literature for the occurrence of non-polleniferous or low concentration samples with many authors associating the problem with coarser grained sediments (e.g. Lowe and Watson, 1993; Velez *et al.*, 2001; Knipping, 2008; Li *et al.*, 2009). Two reasons have been put forward; (i) high energy events producing high rates of sedimentation 'dilute' the volume of pollen (Bruland *et al.*, 1975; Tipping, 1986, 1989) and (ii) the free draining nature of sandy material prevents preservation of pollen (Velez *et al.*, 2001). Middledorp (1982 cited in Bohncke and Wijmstra, 1988) suggests the pollen concentration of a sample is determined in one way by the number of years it represents.

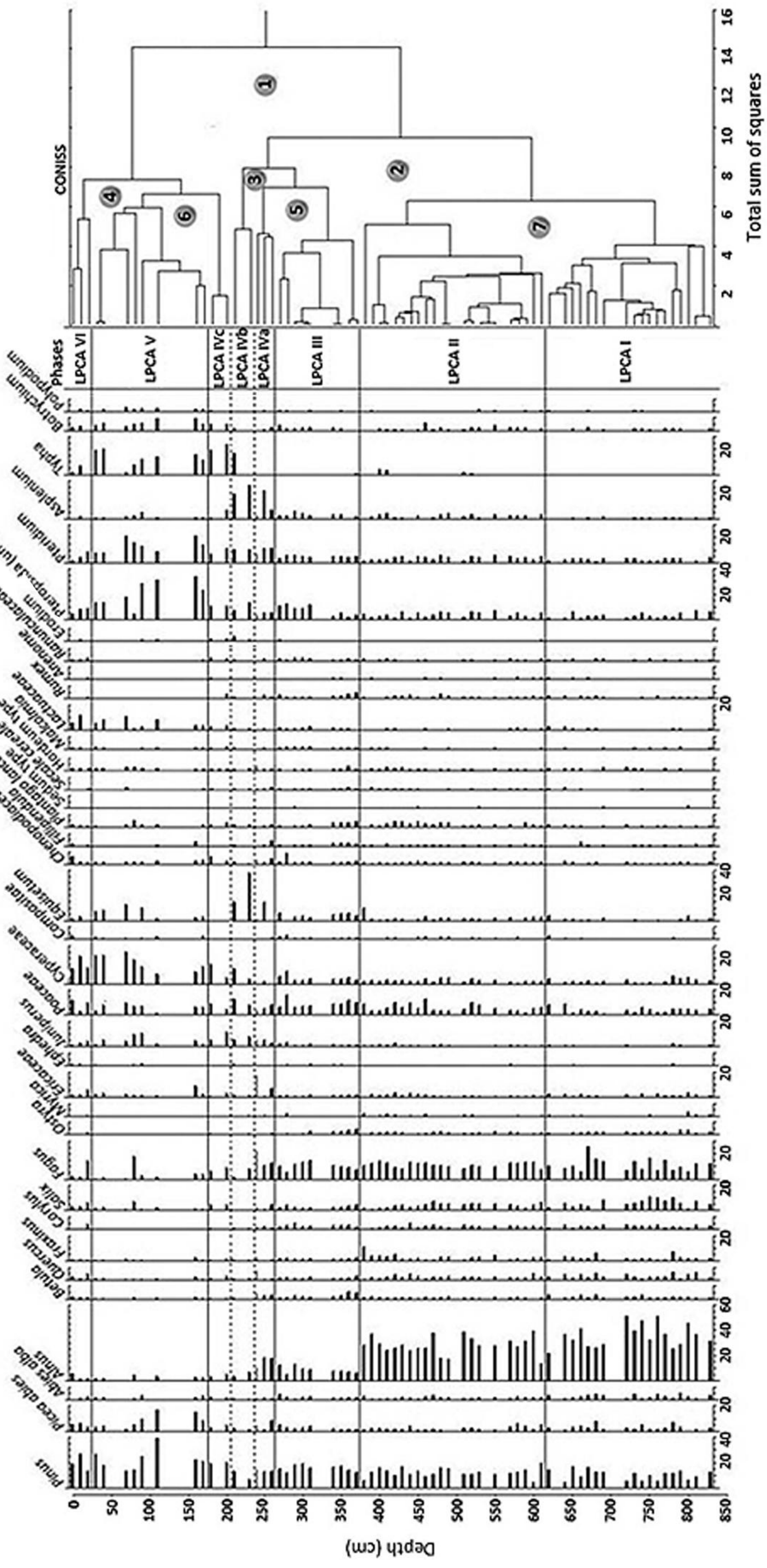


Figure 5.23: CONISS results for the pollen profile from Lake Plav (core LPCA), using only selected taxa with percentage >2%.

Palynolgical phase	Non-polleniferous sample depth (cm)	Pollen count 150 – 300 grains sample depth (cm)
LPCA VI	X	X
LPCA V	50; 60; 100; 120; 130; 140; 150	40; 70; 80; 90; 110
LPCA IV	220	180
LPCA III	320; 330	X
LPCA II	500; 540; 560	X
LPCA I	630; 700; 710; 820	X

Table 5.9: Non-polleniferous samples omitted from the LPCA pollen assemblage and low pollen count sample depths.

The samples from core LPCA were around 0.5 – 1 cm^3 and could represent as little as 3 months of sediment accumulation, which may go some way to indicate the absence of pollen in some units. As previously mentioned, low pollen concentrations have been linked to sandy material (> 25% fine sand; Velez *et al.*, 2001), which seems plausible as dry conditions may cause desiccation of pollen grains (Zetsche, 1932; Srivastava, 1994). Across the Mediterranean region dry conditions have been associated with poor pollen preservation as anoxic/waterlogged conditions are preferential for fossilisation (Cushing, 1967; Muller *et al.*, 2008). Leading on to the impact of climatic changes on pollen grain preservation, which has suggested that cool and dry periods may be associated with low pollen productivity (Andreev *et al.*, 2006). In contrast a sudden climatic deterioration may cause periods of inundation 'flushing out' the system (Atherden *et al.*, 1993). Greater quantities of coarse material suggest such inundation events which would increase sedimentation rates and thus sediment deposition. However, Digerfeldt's (19886) model suggests that greater coarse material indicates increased proximity to the shoreline were water levels are seasonally variable and may create the necessary conditions for pollen degradation.

Authors have also found fine clay/silt material producing samples with little to no pollen (e.g. Fritz *et al.*, 1987; Lowe *et al.*, 1988; Fredlund, 1995; Chambers *et al.*, 1996). Generally, finer material has been dated to the Late Devensian/Weichselian phase when cooler temperatures prevailed and vegetation is likely to have produced fewer pollen grains. Finally, the preservation of pollen grains can be affected by the time taken before deposition of a pollen grain into a suitable sediment environment i.e. anoxic or waterlogged. Early research at Lake Ontario indicated that as pollen was transferred to the lake decay would occur, as a result of the oxidising environment the grains were exposed to prior to deposition (McAndrews and Power, 1973).

Overall, the non-polleniferous samples of phase LPCA I to LPCA III were not clustered together and did not coincide particular changes in the PS or LOI values. Therefore, the poor pollen preservation may be the result of a number of situations: either slow pollen transfer from around the catchment to the site, microbial attack of the grains or the production of aerobic conditions at the site or possibly as a result of the MWP climatic amelioration. The non-polleniferous samples of LPCA IV and LPCA V (up to 130 cm) are suggested to be the result of the system being 'flushed out' by previous high energy events. In the case of sample P10 at 220 cm, the sand fraction is around 16% and had previously been above 25% so a combination of 'flushing out' and pollen concentration dilution may be applicable. Dilution of pollen in samples CA7, P8 and CA8 (130 – 150 cm) is unlikely as these samples are not subsequent to a high energy event and actually occur after a period of 99% clay and silt content. In a similar way to the samples discussed previously it is not obvious why these in particular provided no pollen. A possible suggestion is particularly harsh temperatures within this period affecting the pollen production in and around the catchment and/or possible aerobic conditions at the site causing the pollen grains to degrade. However, the remaining samples with little or no pollen are all associated directly with sediments of high sand content or follow a period when sands were deposited indicative of pollen 'dilution'. As sand does not fall below 19% during phase LPCA V it is suggested that this was a period of significant flooding/lake level variability, when sediments within the catchment may have been finite as large events flushed large quantities of sediment away.

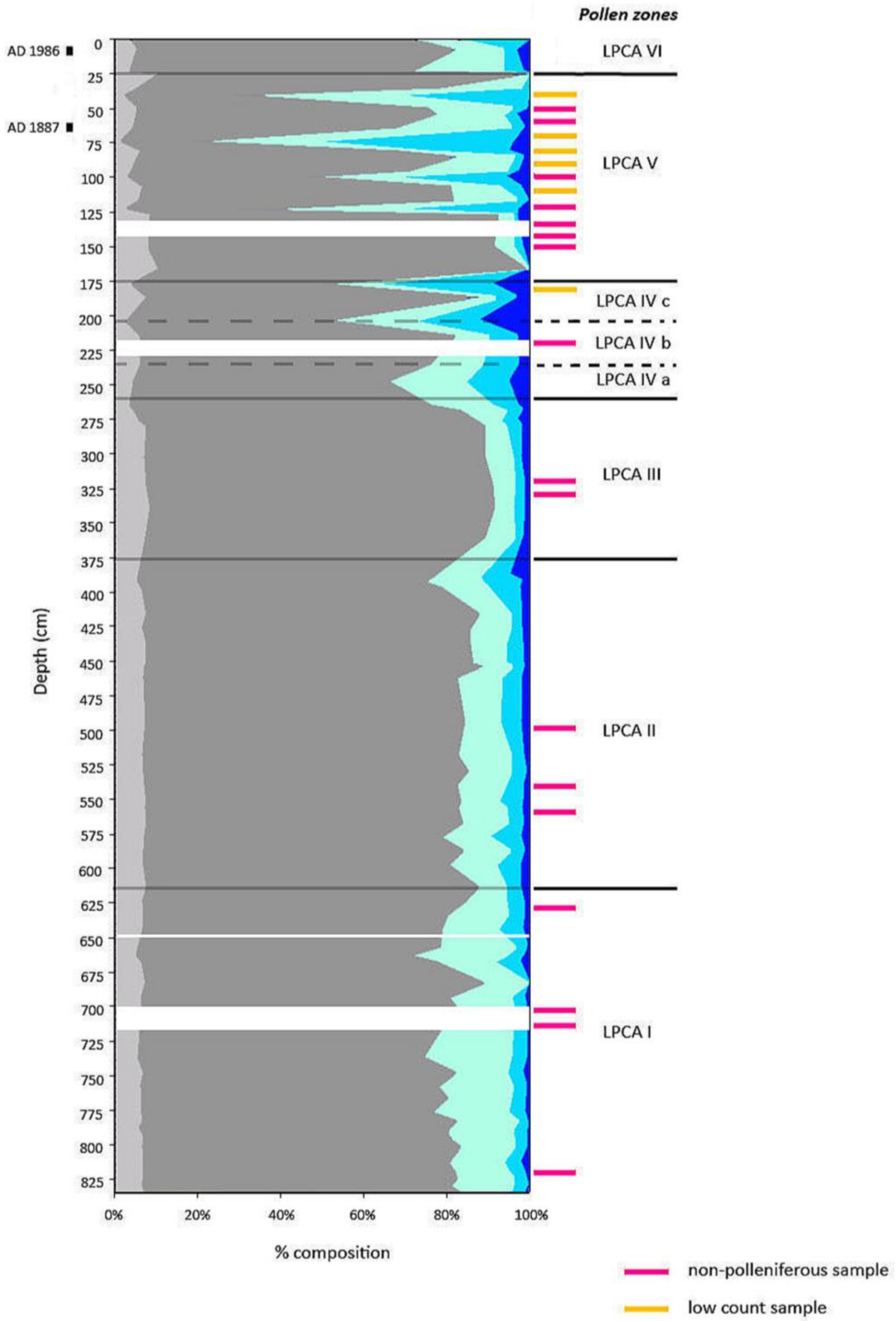


Figure 5.24: The LPCA particle size record and associated periods of little or no pollen preservation.

Palynological analysis of core LPCA was undertaken, but some samples were found to be non-polleniferous or contain low concentrations of pollen (see Table 5.9; Figure 5.24).

Overall, palynological analysis was completed at 10 cm intervals where possible producing a sampling interval of around 20 years (using the age depth model presented in Figure 5.12).

The LPCA pollen assemblage (Figure 5.25) was split into six phases and three sub-phases using the CONISS program, which after visual assessment of the profile were felt to adequately represent the shifts seen within the pollen profile. The following section has been divided into these phases for discussion.

Phase LPCA I (830 – 615 cm) c. AD 500 - 890

This first phase is dominated by arboreal types (>75% TLP for majority of phase). *Alnus* composes around 40% of the pollen assemblage during the first half of the phase, reducing slightly in the upper part at 690 – 640 cm (***c.*** AD 750 – 840). *Pinus*, *Fagus* and *Salix* compose a further 30%, with *Pinus* remaining at a comparatively continuous level during LPCA I. *Salix* registers a general decline through the phase, whilst *Fagus* peaks at 670 cm to 26%, the highest value registered for the species throughout the core. *Tilia* and *Ulmus* are both intermittently present during LPCA I at around 0.2 – 0.6%. Levels of *Fraxinus, Abies* and *Picea* remain below 13% throughout, with *Betula* limited to less than 3.3%. *Ostrya*/*Carpinus* provide a steady abundance at around 2.5%. The dominance of arboreal and herb types characterises LPCA I as there shrub types do not exceed 7%. *Poaceae* and Cyperaceae dominate the herbs, but there is generally a mixed assemblage with 44 herb types/species identified. A variety of high altitude and anthropogenic herbs are identified including Chenopodiaceae, *Campanula* and *Draba*, with pastoral and ruderal types dominating the anthropogenic indicators e.g. *Rumex*, Ranunculaceae, *Compositae*, *Plantago lanceolata* and Lactuceae (Dandelion family). Arable types presented within LPCA I include: *Juglans*, *Prunus*, *Secale cereale* and *Hordeum* - type, with the three former providing intermittent signals that did not exceed 0.7 - 0.5%. Coupled with the presence of potential anthropogenic weeds e.g. *Plantago lanceolata*, Chenopodiaceae, *Rumex* and *Pteridium* (woodland/grazed forest indicator) the pollen assemblage is indicative of local ground disturbance and anthropogenic activity in the

Lake Plav catchment. Within the aquatics and spore *Pteropsida* and *Pteridium* are dominant with *Nuphar* present (<0.4%).

Overall, LPCA I is relatively stable with arboreal types dominating the assemblage during the medieval age. *Alnus* dominates the whole phase and the signal is thought to indicate a dominance of the contemporary situations in which it is found i.e. valley bottom, lake/river edge environments. The arboreal assemblage indicates a 'closed deciduous woodland', suggested by the presence of *Quercus* – *Corylus* – *Alnus*, with *Tilia* and *Ulmus*. This view is supported by the limited *Fraxinus* and *Betula* present as these types enjoy an open forest canopy. Furthermore, both of these species are regarded as 'high to moderate' pollen producers, more likely to be overrepresented within the pollen profile. High altitude montane vegetation is evident with the presence of *Abies* and *Picea* alongside alpine meadow indicators such as Chenopodiaceae, *Campanula* and Cruciferae (exc. *Brassica*.). Woodland and ravine environments are apparent by the continued presence of *Ostrya*/*Carpinus* within this and subsequent phases. Throughout this first phase indication of anthropogenic activity is limited with the arable types *Juglans* and *Secale cereale* being intermittently present. It is likely that within the floodplain of Lake Plav wetlands existed, as indicated by the presence of Cyperaceae and to some degree *Pteropsida*, which can be indicative of both shaded woodland areas and wetlands. Finally, the earlier suggestion that LPCA I represents a period when core LPCA was submerged under water is further supported by the species diversity and also *Nuphar* (water-lily) present which prefers open waters (Polunin, 1980).

Phase LPCA II (615 – 375 cm) c. AD 890 - 1330

Phase LPCA II is similar to LPCA I, with *Alnus* dominating the pollen assemblage. The contribution of *Alnus* throughout this phase is slightly reduced compared to that of the previous phase, but it remains dominant at 13.8 - 41.3%. Declines in *Alnus* abundance coincide with similar trends in the *Pinus*, *Fagus* and *Picea* data. Minor increases in *Salix*, *Plantago lanceolata*, *Equisetum* and indicators of alpine meadows including Cruciferae (exc. *Brassica*.), *Rumex* and *Artemisia* occur, suggesting possible minor anthropogenic clearance. Throughout the phase *Salix* is present at a slightly lower level than previously seen, but both *Salix* and *Fagus* (16%) remain stable. Aquatic and spore species within LPCA II increase not only in number but variety, with the first indication of *Typha latifolia* (Great reedmace or Cats tail) coupled with a greater abundance of *Asplenium* fern type and *Nuphar* (water lily). Cyperaceae remains relatively consistent at ~ 3%, alongside *Equisetum* which is maintains a low presence. Changes in the catchment are further indicated by the persistence of *Secale cereale* alongside *Hordeum* - type, *Plantago lanceolata* and a slight increase in *Juglans* suggesting either a minor increase in anthropogenic activity or a change in source area.

Overall, this phase is similar to that of LPCA I, but it marks the initiation of *Alnus* decline within the catchment, coinciding with a decline in *Picea* and slight rises in both anthropogenic indicators and aquatics and spores. This pattern could suggest a variation in the source area or anthropogenically induced landscape changes at the catchment. As *Alnus* is the main source of decline within the arboreal types and is a type associated with marsh/waterside environment at Lake Plav, it is likely that the initial phase of *Alnus* decline could be attributed to variance in these habitats. Rising levels of meadow types and those indicative of anthropogenic activity coincide with the minor reductions in *Alnus.* Therefore, a change in waterside habitats, possibly via human manipulation of the local landscape, made the area less hospitable for the *Alnus* population. Once more the presence of some closed deciduous forest continues as similar levels of *Quercus* and *Corylus* occur, but a slight decline in the local open forest canopy is identified in the *Betula* record only. The taxonomic diversity of this and the previous phase supports the notion that the site was an 'open water' environment between ***c.*** AD 500 – 1330, with 47 herb types present within this phase and *Nuphar* intermittently present.

Figure 5.25a: Pollen profile for core LPCA, Lake Plav in the Prokletije Mountains Montenegro (arboreal to herb pollen).

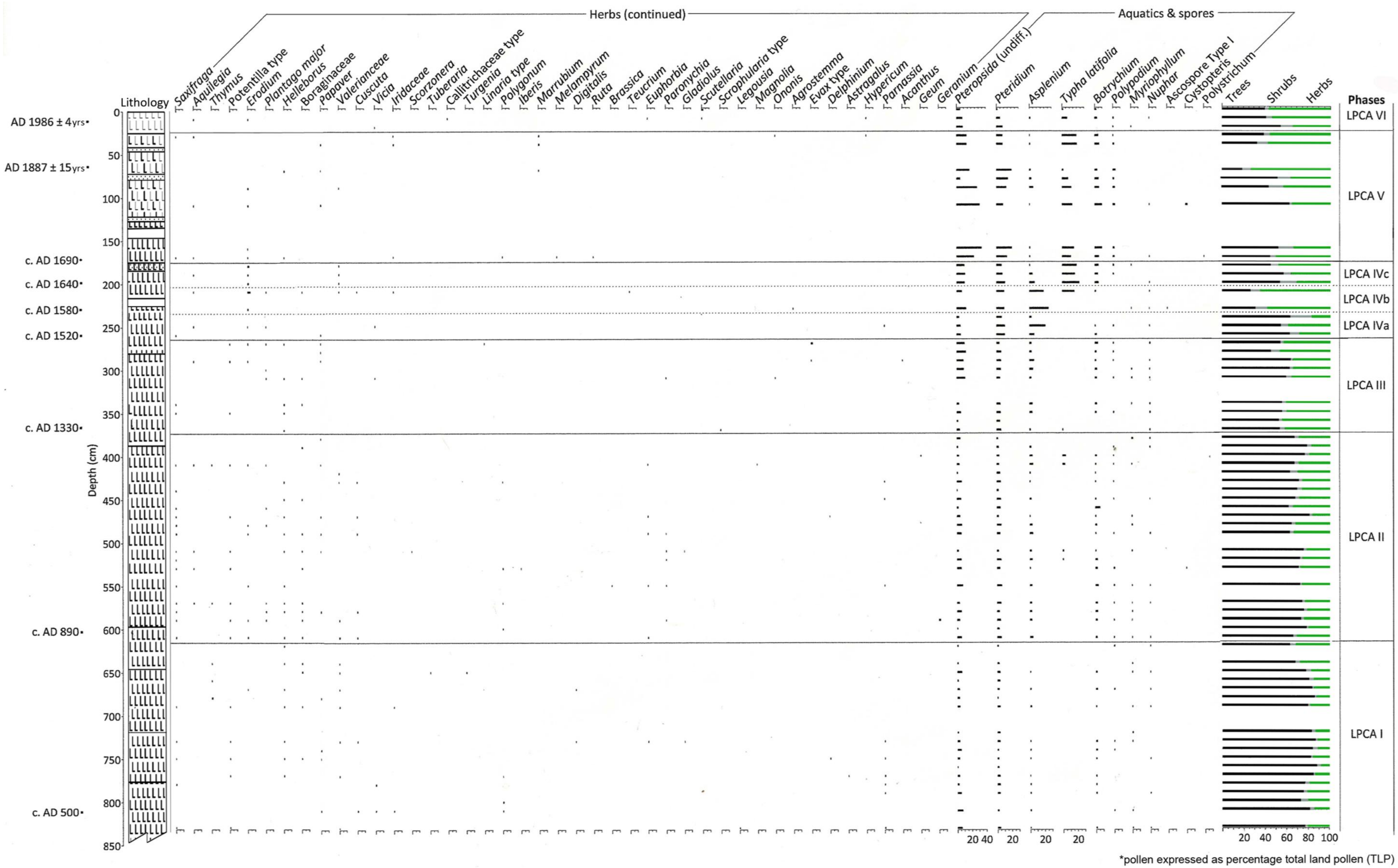


Figure 5.25b: Pollen profile for core LPCA, Lake Plav in the Prokletije Mountains Montenegro (remaining herb pollen with aquatics and spores).

Phase LPCA III (375 – 265 cm) c. AD 1330 - 1520

A marked decline in *Alnus* is the first significant shift seen throughout the LPCA pollen profile, with arboreal pollen falling from around 70% to 50% as a result. Throughout this period little change to the contribution of other arboreal types is visible, with only *Ostrya*/*Carpinus* declining from 5% to nearer 1% by the conclusion of the phase. The roughly 20% decline in *Alnus*, is absorbed by rises in the percentage of shrubs and herb types, with the most notable increases in *Equisetum* (which prefers grassy marshes) and further increase in weeds indicative of pastures or disturbed land e.g. Ranunculaceae, *Artemisia*, Lactuceae and initially *Plantago lanceolata*. The intermittent presence of *Prunus* alongside other arable types e.g. *Secale cereale*, *Hordeum* – type and *Juglans* suggests that between ***c.*** AD 1330 – 1520 the area continued to be used for farming and local food production. During the final half of LPCA III all aquatic types except *Typha latifolia* increased, coinciding with rises in Cyperaceae, *Poaceae* and a spike in Chenopodiaceae. As the taxonomic diversity of the pollen assemblage beings to wane during this phase (38 herb types identified) it is suggested that this may offer some indication of lake development at the site. However, the continued presence of *Nuphar* suggests the site was or local areas were still permanently inundated.

Overall, this phase is characterised by changeable conditions with *Alnus* decline and a continued increase in pastoral, anthropogenic and disturbed land indicators. The suggestion that the site was still under a permanent body of water at this time also indicates that the pollen assemblage produced provides indication of predominantly catchment scale vegetation variance.

Phase LPCA IV (265 – 175 cm) c. AD 1520 - 1690

This phase was divided into three sub-phases, LPCA IV a, b and c to highlight the variance seen throughout the phase. Increases of aquatic and spore types and *Typha latifolia* characterise this phase. Phase LPCA IVa begins with a short lived increase in *Alnus*, before declining to 9% at 260 cm. Coeval to this decline, *Fagus*, Asplenium, Ericaceae and

Juniperus increase, the latter two being indicative of open areas and at present more common in sites above 1900 m a.s.l.. Abrupt peaks in Ericaceae and *Equisetum* indicate a possible shift in source area, as they do not seem to fit the abundance recorded prior to or after their peaks. This first phase has reduced taxonomic diversity with 27 herb types identified and the anthropogenic indicators, *Juglans* and *Prunus*, are absence.

By LPCA IVb *Equisetum* records a peak of 37%, the highest input of a non-arboreal type throughout the core. This not only coincides with a peak in *Pteropsida* and Asplenium, but occurs prior to a rise in Cyperaceae and significant *Typha latifolia* input. The *Typha latifolia* record leaps from 0 to 16 and then 23 spores (raw count) in LPCA IVc. This species can tolerate up to 2 feet of standing water (Lieffers 1983; ISSG, 2006) and may offer the first indication of lake development and increased proximity to a lake edge or shallower environment. The continued presence of anthropogenic indicators indicates continued human activity within the catchment. This may be supported by the decline in all arboreal pollen, except *Pinus*. However, as the arboreal decline is coupled with reduced taxonomic diversity the changes are thought to indicate a more 'local' signal, which may account for the minor rise in anthropogenic indicators.

The final sub-phase, LPCA IVc is characterised by a recovery in many arboreal types with the presence once more of *Juglans* and *Prunus*, and waning levels of *Hordeum* – type, *Secale cereale* and *Rumex*. A slight rise in Lactuceae, Ranunculaceae and Valerianceae could provide indication that meadows or fallow lands were expanding or that they were beginning to encroach on the River Ljuča floodplain. The final sample of the phase at 180 cm has previously been noted as a 'low count' sample, suggesting that this period was characterised by change.

Generally, LPCA IV is characterised by variable presence in pollen types coupled with steep rises in types associated with lake periphery/wetland sites. Therefore, the record suggests that during ***c.*** AD 1520 – 1690 the LPCA core site began to change and was most likely under water <2 feet deep for all or most of the year from ***c.*** AD 1620 - 1690.

Phase LPCA V (175 – 25 cm) c. AD 1690 – 1960

This phase is dominated by increasing levels of Cyperaceae (Sedge) which is indicative of lake margins. The latter stages of LPCA V are dated to AD 1887 and are characterised by a variable arboreal content and increased herb, aquatic and spores abundances. However, within this phase five 'low count' samples were identified. The first coincides with core peaks in *Pinus* and *Picea* and this may be indicative of changing source area, as both *Pinus* and *Picea* are indicative of high altitude montane forest. A rise in *Juniperus* coincides with the peaks in *Pinus* and *Picea*, which may be a further indication of variation in source area. Similar peaks in types associated with areas away from the floodplain such as Lactuceae, *Dianthus* – type and *Papaver* are evident at this time. During the latter stages of this phase, anthropogenic indicators decline, but wetland indicators rise (Cyperaceae and *Typha latifolia*). This suggests that pollen assemblage is more representative of the locality than the catchment which is further supported by low taxonomic diversity of this phase. The continued presence of Chenopodiaceae and increased Lactuceae contribution suggests that land disturbance within the catchment continued as both types prefer grassy open areas. Other anthropogenic indicators such as *Rumex* and *Plantago lanceolata* are absent or intermittent in abundance, which may be the result of pollen source area. A similar pattern is seen within the *Ostrya*/*Carpinus* and *Fagus* record, which are absent for the majority of the phase, before a peak in *Fagus* coincides with rises in Cyperaceae, Asplenium and *Equisetum*. Caution must be taken with the interpretation of these variations as a number of samples throughout LPCA V were found to be non-polleniferous or had poor pollen preservation which could influence the assemblage recorded.

Generally, LPCA V indicates shallow waters or seasonal terrestrialisation of the LPCA core site. The site may have been permanently under a low level of water (< 2 feet) or seasonally flooded, with the latter being more likely as the stage progressed coinciding with increased dominance of Cyperaceae.

Phase LPCA VI (25 – 0 cm) c. AD 1960 - 2006

The initiation of the final phase is marked by a continued reduction of arboreal pollen (< 50%) content, increased herbs types and low taxonomic diversity. *Pinus* dominates the arboreal types, with fluctuating *Fagus* and recovered *Quercus* inputs. Shrub types are dominated by *Ericaceae* and *Juniperus*, alongside increased levels of *Dianthus* type and Cruciferae (exc. *Brassica*) which are indicative of alpine meadows and rocky open areas. These changes suggest an expansion of open landscapes in the higher altitudes of the catchment. Coupled with a decline in arboreal types this may point to anthropogenic pressure upon the Lake Plav catchment causing pastoral areas in the upper and lower catchment to expand, contracting the woodlands that were evident in previous phases. Furthermore, the suggestion that a more open landscape dominates the upper catchment is compounded by the disappearance of *Abies* by the end of LPCA VI.

Aquatics decline during LPCA VI, with a significant decline evident in the *Typha latifolia* record. The presence of more marsh loving types (Cyperaceae; *Equisetum*) and types found in the contemporary meadow environment close to the LPCA core site (e.g. *Lactuceae*, *Poaceae, Compositae*) infers that during the 1900s the site became what it is today i.e. a terrestrial and marshy habitat. Anthropogenic indicators are variable during LPCA VI with only *Hordeum* – type and *Plantago lanceolata* maintaining a presence throughout.

Overall the pollen assemblage of core LPCA is characterised by a general reduction in arboreal pollen and increase in wetland indicators ***c.*** AD 1330 – 1520. Finally, the present day assemblage discovered at Lake Plav, is one dominated by marsh-land types with some anthropogenic indicators and high *Pinus* input.

5.4.5 Ostracoda results

Ostracoda were recovered from core LPCA below 200 cm, but were sparse in core LPCB.

Those extracted from core LPCB were broken not permitting identification. In total six samples from core LPCA provided good preservation of ostracoda valves, although the quality and quantity of preservation was variable. Positive identification to at least the family type was possible in the core LPCA samples and Table 5.10 highlights the number of valves retrieved and level to which they were identified. The number of fossils retrieved was lower than expected, as a predominantly karst catchment would be expected to provide excellent conditions for the preservation of such crustaceans, due to the high $CaCO_3$ content (discussed further in section 6.1.5, Chapter 6).

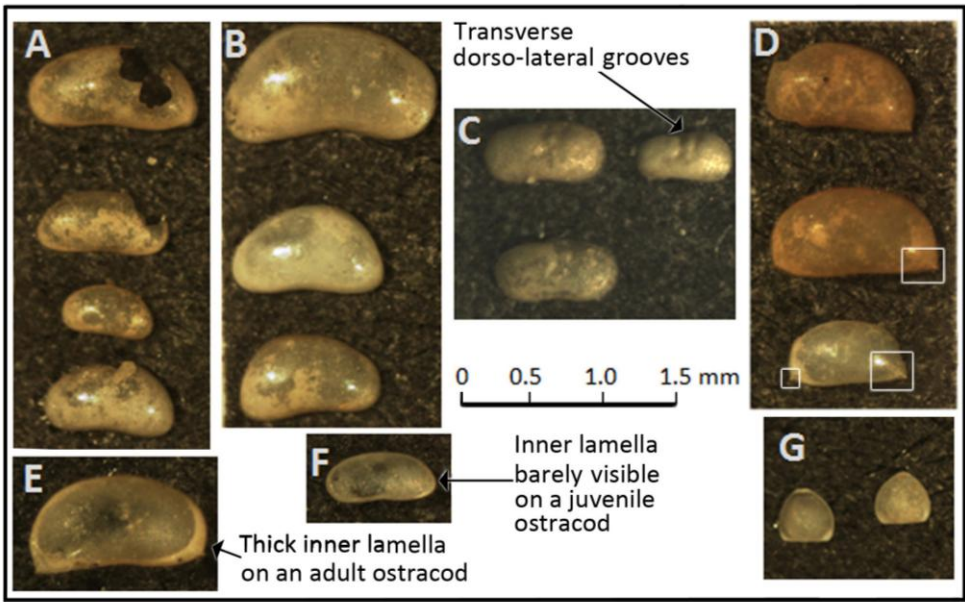


Figure 5.26: A selection of ostracoda found in core LPCA sediments. (A) Elongate Candona – type, (B) wedge shaped Candona – type, (C) Ilyocypris, (D) Candona type with posterior and anterior spine like features, boxed areas, (E, F) inner lamella of and adult and juvenile ostracod and (G) glochidium bivalve unionoindean larvae.

The final ostracoda data are presented as 'Total valves' and thus one carapace is counted as 2 valves, but the samples in which full carapaces were found are indicated within Table 5.10. Full family and sub-family information is presented in Table 5.11 for each genus identified within core LPCA. Ostracoda were only present in sediments below 200 cm, dated to pre AD 1640. Good taxonomic diversity existed in all but Po540, where no ostracoda could be positively identified. In total 8 different genera were identified within

LPCA, with 13 varieties of ostracoda represented. Three *Candona* types were identified to genus level, and were split into three types: (1) more elongate (see Figure 5.26 A), (2) a wedge shape and (3) flatter wedge shape (see Figure 5.26 B). *Psuecandona* were identified by the fine surface pitting of the valve, whilst *Fabaeformiscandona* are characterised by a slight anterior protrusion. The carapace of *Ilyocypris* was defined by the small pits across the valve and transverse dorso-lateral grooves; sulci (Figure 5.26 C). Finally, the *Cypria* and *Cyclocypris* genera were differentiated by the surface shine and globular nature of the valve (*Cypria* being flatter with less shine due to a slightly pitted surface). The three varieties of ostracoda that could not be identified to genus level are thought to be of the Candonidae or Cyprididae family (Dr David Horne *pers. comm.*). A 'spiny' ostracoda of the Candonidae family, possibly *Candona* type, was identified with the assistance of Dr David Horne and it was characterised by spine-like features on the posterior and anterior of the valve (see Figure 5.26 D). Juvenile and adult ostracoda were distinguished by the thickness of the inner lamella, with adults having a thicker inner lamella (see Figure 5.26 E, F), and females by an arching 'shadow' at the posterior of the valve, remnants of the ovaries (Horne *et al.*, 2002). The preserved state of the valves gave some indication of local variance in the chemical composition of sediments at the site or iron pan/soil formation with valves in Po 360 stained an orangey red colour (Dr David Horne *pers. comm.*). The core stratigraphy does not indicate iron pan/soil formation at this point and thus it is more likely that the chemical composition of sediments changed. Indication of this would be possible by Redox analysis throughout the core.

The general ostracoda assemblage contained a variety of adults and juveniles with complete carapaces found in samples Po200, Po440 and Po660, with juveniles indicating the first moult was not reached before death (Waas, 1995). The percentage of complete carapaces found in the core LPCA samples was 9.5%. Over 90% of the assemblage was composed of disarticulated valves, but even through some ostracoda are more prone to disarticulation the high percentage of separated valves is indicative of post-mortem reworking (Griffiths and Holmes, 2000). Lake conditions (water chemistry, pH, temperatures etc) at the time of deposition cannot be reconstructed as none of the valves were identified to species level. If SEM photography was available or appendages in types such as *Ilyocypris* had been preserved, identification to species level may have been possible. Two glochidium bivalve unionoindean larvae were found (see Figure 5.26 G) in Po360, which are parasitic on fish gills and therefore indicate the presence of fish within the lake at AD 1350 (Bogan, 1993; Thorp and Covich, 2009; Kováts *et al.*, 2010). A final note is the lack of any *Darwinlodea* or *Cytheroidea* types as the Lake Plav assemblage was dominated by, on average, 63 % Candonids (max. 74% at Po360; min. 46% at Po800), with greater quantities of Candonids seen in the lower assemblages; Po440 – Po800.

Depth (cm) → → *Sample code → →*	**200** *Po200*	**360** *Po360*	**440** *Po440*	**540** *Po540*	**660** *Po660*	**800** *Po800*
Candona sp. 1	1	32	17			
Candona sp. 2	20	9	25			
Candona sp. 3		7				1
Fabaeformiscandona			19			
Psuedocandona	1		12		10	11
*Cyclocypris**	2		30		14	1
Cypria*	2	19	33		10	3
Spiney poss. *Candona*		3 (CC/IP 2)	6		1	2
Ilyocypris	4		3			
Potamocypris		20	22			1
Cypridopsis		30 (CC/IP 6)				
Cyprididae juvenile					4	4
Candonid				12		2
No ID adult					1	1
No ID juvenile		26			12	1
No ID mix	18	7	7	5		
Total valves	**48**	**153**	**204**	**17**	**52**	**27**
COMMENTS	*2 full carapaces		*20 full carapaces	Poor preservation	*2 full carapaces	

Table 5.10: Ostracoda found within core LPCA. CC/IP indicates orangey red stained ostracoda thought to be a product of the chemical composition of sediments.

Overall, a mix of adult and juvenile valves/carapaces were identified which is indicative of local deposition. Central lake sediments would be characterised by a concentration of juvenile carapaces as a result of wave action suspending these smaller individuals and re-depositing them in more central lake positions (e.g. Whatley, 1988; Waas, 1995; Frenzel and Boomer, 2005; Pérez *et al.*, 2010). The presence of allochthonous valves such as *Potamocypris*, which are associated with slow flowing water, is a possible indicator of deltaic influences at the site ***c.*** AD 1206 and ***c.*** AD 1351.

Family	Sub-family	Genus	Comments
Candonidae	Candoninae Cyclocypridinae	- *Candona* sp. type 1 elongate - *Candona* sp. type 2 short - *Candona* sp. type 3 flat - *Fabaeformiscandona* - *Psuedocandona* - *Cyclocypris* - *Cypria*	- Cosmopolitan family - Generally indicative of a permanent water body
Ilyocyprididae	Ilyocypridinae	- *Ilyocypris*	- Generally indicative of a permanent water body - Species ID requires preservation of appendages
Cyprididae	Cypridopsinae	- *Cypridopsis* - *Potamocypris*	- Cosmopolitan genus - Cosmopolitan genus, but normally associate with flowing water

Table 5.11: Family, sub-family and genus documentation for the ostracoda found in core LPCA (following Meisch, 2000).

5.5 CORE LPCB DATA

Three records were produced for core LPCB including detailed particle size analysis, loss-on-ignition and low frequency magnetic susceptibility.

5.5.1 Particle size analysis

Figure 5.27 provides the record of sediment particle size (PS) for core LPCB, which was divided into four phases. This followed the dominance of coarser versus finer fractions. Throughout the high resolution PS record continual variations are evident (see Figure 5.27). The greatest quantities of coarse grained material, >250 µm, were recorded in LPCB I (see Table 5.12; Figure 5.27). A 4-fold decrease of silts to medium sands occurred to 710 cm (***c.*** AD 1450) when coarse to very coarse sands peaked. Finer materials then increase to 80% at 675 cm (***c.*** AD 1270) followed by another period of dominance by coarse to very coarse material. This pattern was replicated across the phase, with medium sand shifting between 10% and 31%. The percentage of sand grains coarser than 2000 µm (2 mm) was low throughout the phase at < 0.4% (see Table 5.12), but coincided with peaks in coarse to very coarse sands (Figure 5.27). This suggests that the advent of larger flood events or greater fluvial inputs, and it is therefore unsurprising that clay sized particles remained < 4%. Overall sand fractions larger than 250 µm dominated phase LPCB I (***c.*** AD 1100 - 1660) with periods of sand fractions larger than 500 µm indicative of increasing turbulence.

During LPCB II (***c.*** AD 1660 - 1760) there is a visible increase in clays and silts (Figure 5.27), but when quantified the mean values for both rise by only 0.9% and 4.7% respectively (Table 5.12). Medium and coarse to very coarse sands reach some of their lowest levels in the record during this phase, comprising less than 10% of the sediment at 285 cm. The general trend by the end of LPCB II is rising levels of silt, but by 275 cm silt falls to its lowest value during this phase, to 13.2% (see Figure 5.27; Table 5.12).

By ***c.*** AD 1760, phase LPCB III, the sediments were dominated by coarse to very coarse sands, in relatively equal ratios. The peaks in silts versus sands were more prolonged during this phase and less abrupt than previous phases. In contrast to LPCB I the proportion of fine sands increased from 194 cm (***c.*** AD 1820) onwards peaking at 44%, as the proportion of coarse to very coarse sands fell. The dominance of silts and fine sand between 194 - 176 cm decreases to 150 cm (***c.*** AD 1820 - 1850), a pattern mirrored by clay material. Throughout LPCB III fine sands dominate the fractions >63 µm, with medium and coarse to very coarse sands registering their lowest values so far (see Table 5.12). During the latter stages of LPCB III the proportion of clays and silts rise, with the highest maximum value of silt for LPCB I to III registered (see Table 5.12).

Phase	Depth (cm)	Clay (< 2 µm)			Silt (2 - 63 µm)		
		Minimum value (%)	Maximum value (%)	Mean value (%)	Minimum value (%)	Maximum value (%)	Mean value (%)
LPCB IV	0 - 110	3.6	9.7	7.2	57.2	82.0	74.7
LPCB III	110 - 275	0.8	4.6	2.4	10	64.5	29.5
LPCB II	275 - 425	1.1	5.5	3.0	13.2	53.7	27.9
LPCB I	425 - 750	0.7	5.1	2.1	9.8	60.1	23.2

Phase	Depth (cm)	Fine sand (63 - 250 µm)			Medium sand (250 - 500 µm)		
		Minimum value (%)	Maximum value (%)	Mean value (%)	Minimum value (%)	Maximum value (%)	Mean value (%)
LPCB IV	0 - 110	2.3	23.8	12.3	0.8	6.0	2.2
LPCB III	110 - 275	8.9	44.4	22.8	3.5	37.9	24.6
LPCB II	275 - 425	6.3	19.8	13.6	14.6	36.2	25.7
LPCB I	425 - 750	5.9	25.9	12.7	9.8	41.2	26.9

Phase	Depth (cm)	Coarse sand* (500 - 2000 µm)			Gravel/Grit (>2000 µm)		
		Minimum value (%)	Maximum value (%)	Mean value (%)	Minimum value (%)	Maximum value (%)	Mean value (%)
LPCB IV	0 - 110	1.4	10.8	3.6	0.0	0.1	0.0
LPCB III	110 - 275	1.9	47.6	20.7	0.0	0.2	0.0
LPCB II	275 - 425	7.3	56.9	29.7	0.0	0.4	0.1
LPCB I	425 - 750	3.4	57.6	34.3	0.0	0.4	0.8

Table 5.12: Summarised particle size results for core LPCB, *coarse to very coarse sand.

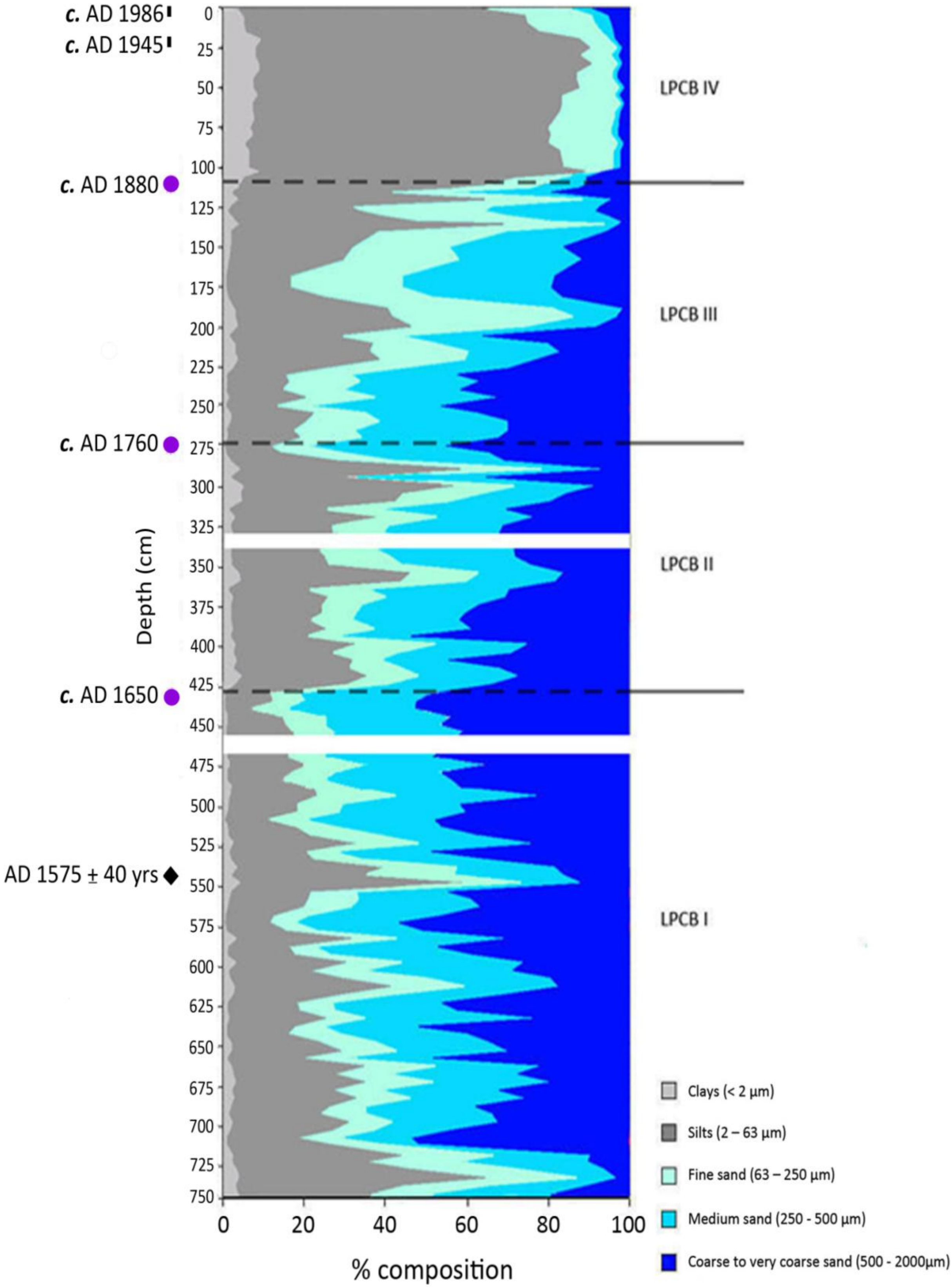


Figure 5.27: Particle size record for core LPCB. The square blocks represent ^{137}Cs dates, the diamond indicates a radiocarbon date and purple stars provide extrapolated radiocarbon dates.

The beginning of the final phase, LPCB IV (***c.*** AD 1880 - 2008), coincided with the end of the LIA across the Mediterranean region and Europe (e.g. Grove, 1988; Briffa *et al.*, 2001; Matthews and Briffa, 2005; González-Trueba, 2008). The particle size data for LPCB IV was dominated by fractions < 250µm in contrast to the previous phases. This is evident in the summarised data of core LPCB (Table 5.12), which indicates that maximum values and the highest mean values for clays and silts were recorded during this phase. Clay material returned to levels seen during LPCB II by the end of LPCB IV. Coupled with the high percentage of silts suggests that the depositional setting since ***c.*** 1880 has been less turbulent than previous times at Lake Plav. The lowest values for all sand fractions were recorded during LPCB IV, but during the uppermost 15 cm a minor rise in the percentage of sands occurred as the proportion of clay and silt sized particles fell.

Overall, the particle size distribution of LPCB is dissimilar to LPCA as coarser fractions (> 250 µm) dominate the record. This suggests that the location of LPCB was dominated by a more turbulent depositional environment with low energy events being atypical. The sediment composition continually changes, with periods of greater instability seen between AD 1100 - 1880, when coarser fractions dominated. This contrasts with the recent and present day environment, based upon the sediment composition of the uppermost samples, where silts dominate.

5.5.2 Magnetic susceptibility record

The low frequency (LF) magnetic susceptibility of core LPCB (Figure 5.28) bore some similarity to the LOI record (Figure 5.28), in that it did not vary significantly until the upper part of the record, ***c.*** AD 1880 - 2008. This is evident in Table 5.13 as the mean values for LF magnetic susceptibility differ by just 38.2 $x10^{-9}$ m^3 kg^{-1} and the minimum values by just 13 $x10^{-9}$ m^3 kg^{-1}. The brief hiatuses recorded within the stratigraphic record are represented by dotted lines within the graphs and the PS data phases are applied.

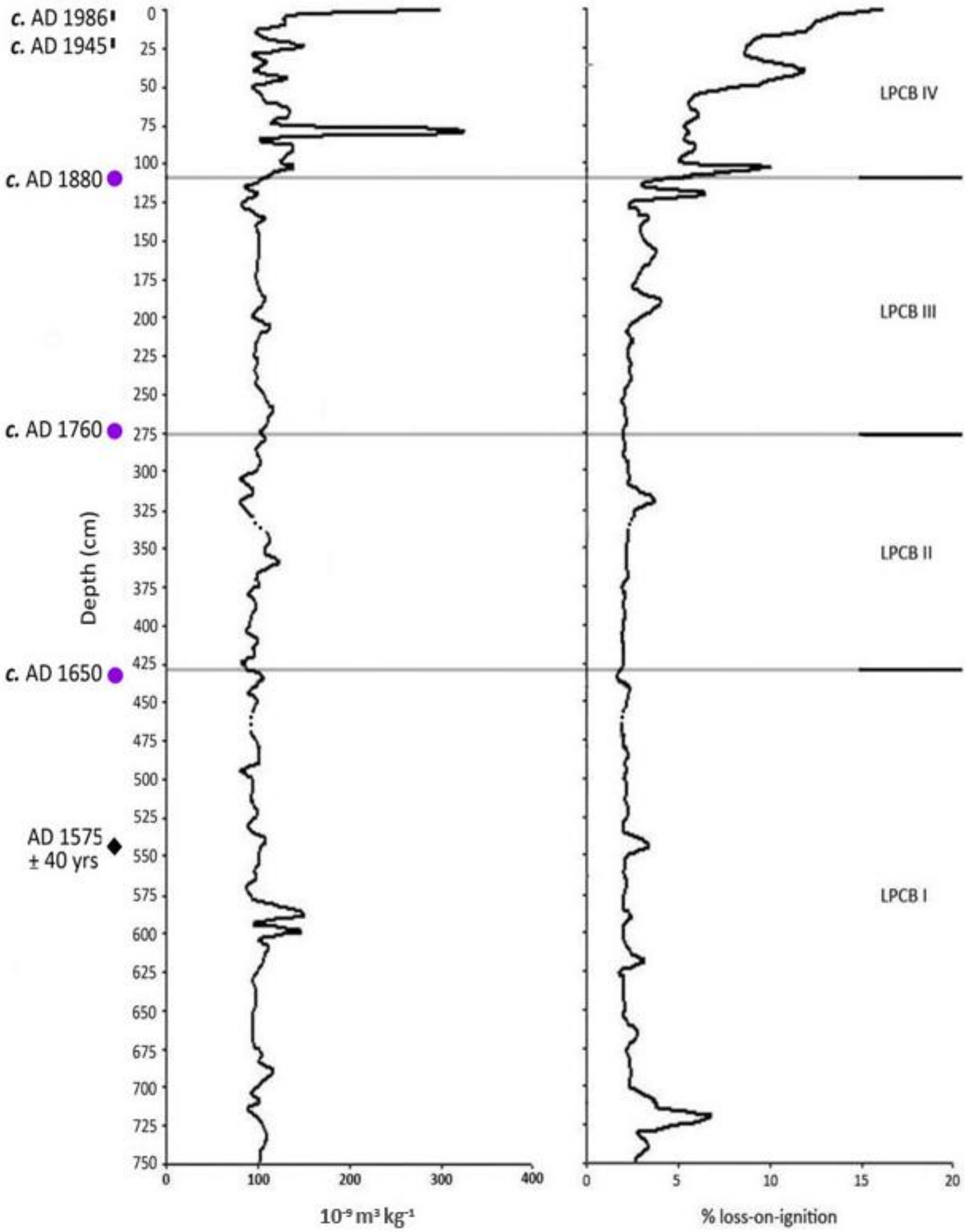


Figure 5.28: Low frequency magnetic susceptibility and Loss-on-ignition for core LPCB, phases based upon particle size data. The square blocks represent ^{137}Cs dates, the diamond indicates a radiocarbon date and purple stars provide extrapolated radiocarbon dates.

Magnetic susceptibility

Phase	Depth (cm)	Minimum value ($x10^{-9}$ m^3 kg^{-1})	Maximum value ($x10^{-9}$ m^3 kg^{-1})	Mean value ($x10^{-9}$ m^3 kg^{-1})
LPCB IV	0 - 110	95.0	327.0	137.2
LPCB III	110 - 275	85.0	117.0	101.7
LPCB II	275 - 425	82.0	125.0	99.0
LPCB I	425 - 750	83.0	150.0	102.2

Loss-on-ignition

Phase	Depth (cm)	Minimum value (%)	Maximum value (%)	Mean value (%)
LPCB IV	0 - 110	5.1	16.2	8.7
LPCB III	110 - 275	2.0	6.5	2.9
LPCB II	275 - 425	1.9	3.7	2.3
LPCB I	425 - 750	1.7	6.8	2.5

Table 5.13: Summarised magnetic susceptibility and loss-on-ignition results for core LPCB.

LF magnetic susceptibility remained below 150 $x10^{-9}$ m^3 kg^{-1} throughout phases LPCB I to LPCB III. Theses phases are characterised by minor fluctuations around a mean of between 99 – 102.2 $x10^{-9}$ m^3 kg^{-1}. A brief increase in LF magnetic susceptibility during LPCB I between 605 – 580 cm (***c.*** AD 1430 – 1480) to values of 150 $x10^{-9}$ m^3 kg^{-1} are evident in Figure 5.28. Two pronounced peaks are apparent in the upper core i.e. phase LPCB IV. The highest LF magnetic susceptibility value was recorded at 80 cm (***c.*** AD 1900), but this was the result of a single sample. Samples prior to and after were around the mean 137.2 $x10^{-9}$ m^3 kg^{-1}. The final uppermost samples produced rising LF magnetic susceptibility, which is likely to be a product of recent magnetic enhancement as a result of anthropogenic activity.

5.5.3 Organic content and loss-on-ignition record

Core LPCB was characterised by low organic content, visible in both the stratigraphic record and quantified by the LOI record presented in Figure 5.28. Table 5.13 highlights the low organic content as mean values remain below 8.7%. LOI figures for LPCB I began comparatively high at 6.8%, before falling to around 2.5% with only minor increases (~ 1%) through the remainder of this phase. LPCB II and III continued in a similar fashion until ***c.*** AD 1830 (180 cm; LPCB III), when LOI began to rise slowly reaching 6.5% by the end of the phase. This rising trend was continued during LPCB IV, where a peak to 10% organic content ***c.*** AD 1890 occurred. Levels then plateau just below the mean at 6% between ***c.*** AD 1890 – 1920. Organic content then begins to rise, doubling at 40 cm (***c.*** AD 1930) and rising to 16.2% at the uppermost sample of the record.

In summary, the organic content of LPCB is suppressed below 6.8% for all but the final phase of the core, where the transition between LPCB III and IV marks a rise in organic content. Organic content does not steadily rise during the final phase but occurs in peaks and troughs.

5.6 CORE BJC1 FROM LAKE C, BUNI I JEZERCES

A single core (BJC1) was extracted to a depth of 1.45 m from Lake C within the Buni i Jezerces lake complex (see Figure 3.5, Chapter 3). Dating (see section 5.2.3), sedimentary and palynological assessment of core BJC1 was completed. LOI and particle size analysis were not completed on the sediments as the primary aim for extracting this core was to produce a primarily climate driven pollen record. Furthermore, palynological indicators such as *Pediastrum* can provide a reliable proxy for both lake level and lake productivity. Low organic content was apparent within the predominantly silt sediments.

5.6.1 The stratigraphic record of core BJC1

Core BJC1 was dominated by silt sediments, as seen in Table 5.14 and Figure 5.29. The slight variations through the core are captured by the five units identified in Figure 5.29. A limited number of small organic, possibly herbaceous, remains and some clay were evident in the first unit, but the sharp boundary to unit two represented a change to mottled silt. Unit three had a more diffuse boundary and sediments remained a mottled clay silt. The fourth unit was mottled silt sandwiching a sharply bounded clay silt, free from mottles. The final unit was dominated by silt sediments, a heterogeneous colour and a minor fraction of herbaceous fragments.

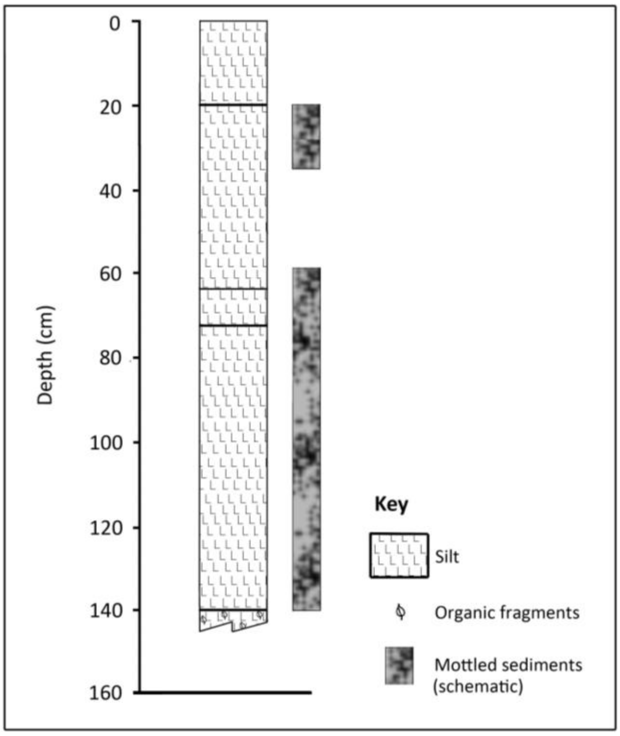


Figure 5.29: Core BJC1 stratigraphic record.

Depth (cm)	Description & composition	Boundary	Colour
0 – 8	Ag^4 Th^+	G	5Y 4/1
8 – 12	Ag^4 Th^+	G	5Y 3/1
12 – 17.5	Ag^4 Th^+	D	5Y 4/1
17.5 – 20	Ag^4 Th^+	S	5Y 3/1
20 – 36	Ag^4	S	2.5Y 5/2; mottles 2.5Y 6/2
36 – 59	Ag^3 As^1	S	2.5Y 4/2
59 – 63	Ag^4	D	2.5Y 5/2; mottles 2.5Y 6/2
63 – 73	Ag^3 As^1	D	2.5Y 5/2; mottles 2.5Y 3/1
73 – 117	Ag^4	D	2.5Y 5/2; mottles 2.5Y 7/1 &7 /2
117 – 140	Ag^4	S	2.5Y 5/2; mottles 2.5Y 3/1, 2.5Y 6/2, GLEY1 4/5G
140 – 145	Ag^3 As^2 Th^1		2.5Y 4/1 to 4/2

Table 5.14: Stratigraphic record for core BJC1, (boundary code, see Appendix IX).

5.6.2 Magnetic susceptibility record of core BJC1

The most prominent change in LF magnetic susceptibility through core BJC1 is an abrupt rise to 487 x10^{-9} m^3 kg^{-1} at 110 cm (***c.*** 1540 BC), which occurred between smaller 19 x10^{-9} m^3kg^{-1} to 58 x10^{-9} m^3 kg^{-1} fluctuations (see Figure 5.30; Table 5.15). This was followed by a similarly steep decline to 83 x10^{-9} m^3 kg^{-1} by 105 cm (***c.*** 1380 BC; Figure 5.30). The lower level of magnetic susceptibility is then sustained into BJMS II where it reaches a minimum 14 x10^{-9} m^3 kg^{-1} at 90 cm rising slowly to around 155 x10^{-9} m^3 kg^{-1} at 40 – 45 cm (***c.*** AD 560 - 720). Between ***c.*** AD 560 – 1040 magnetic susceptibility declines to BJMS III, where a relative plateau in susceptibility occurs (mean 38.4 x10^{-9} m^3 kg^{-1} see Table 5.15), before at 20 cm (***c.*** AD 1360) susceptibility increases. Throughout phases BJMS I to III the minimum LF magnetic susceptibility values are relatively similar (see Table 5.15). During BJMS IV a rising trend to 10 cm (***c.*** 1690 - 1850) is evident, before a second plateau is apparent. Magnetic susceptibility rises in BJMS IV until 2.5 cm where the record wanes to 106 x10^{-9} m^3 kg^{-1} at 0 cm. Unlike the lower catchment records there is no evidence of magnetic enhancement in the uppermost records.

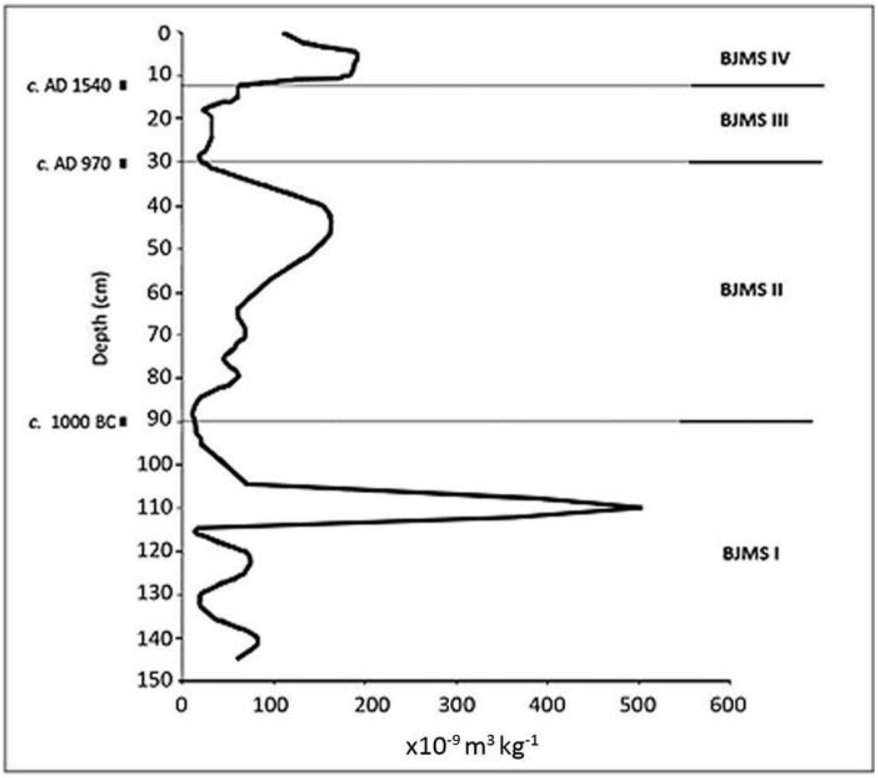


Figure 5.30: Magnetic susceptibility record for core BJC 1.

Overall, the record is characterised by three periods of elevated magnetic susceptibility at: ***c.*** AD 560 – 720, ***c.*** AD 1540 and ***c.*** AD 1690 – 1850, with particularly low susceptibility evident between ***c.*** BC 2670 – 730 and ***c.*** AD 1040 – 1440.

Phase	Depth (cm)	Minimum value ($x10^{-9}$ m^3 kg^{-1})	Maximum value ($x10^{-9}$ m^3 kg^{-1})	Mean value ($x10^{-9}$ m^3 kg^{-1})
BJMS IV	0 - 11	106.2	186.1	156.9
BJMS III	11. - 30	21.1	62.1	38.4
BJMS II	30 - 90	13.9	158.6	81.4
BJMS I	90 - 145	19.4	487.0	89.0

Table 5.15: Summarised LF magnetic susceptibility results for core BJC1.

5.6.3 Core BJC1 palynological profile

To objectively identify the points of greatest change, a dendrogram was produced (method presented in section 4.6.4, Chapter 4) and to ensure changes in the contiguous samples were not overlooked this section (0 – 30 cm) was assessed separately to the 'full record' (0 – 145 cm).

In the contiguous sections the greatest similarities were identified at 18 – 30 cm (see Figure 5.31 A). Three groups were single samples characterised by abrupt shifts, the *Pinus* collapse (25 cm) and *Picea* expansion (19 cm). Analysis of the full profile suggested seven divisions, three of which overlapped the contiguous sample results, see groups 1, 2 and 5 Figure 5.31 A and 1, 3 and 4 Figure 5.31 B. Of the four remaining divisions (Figure 5.31 B) only group 2 was used as the non-polleniferous samples at 75 cm and 80 cm provided a visual division. A further three divisions were made at 29.5, 25.5 and 18 cm. The 29.5 cm division was implemented as a sub-division as it identified a single sample at the base of the contiguous profile (Group 5, Figure 5.31 A).

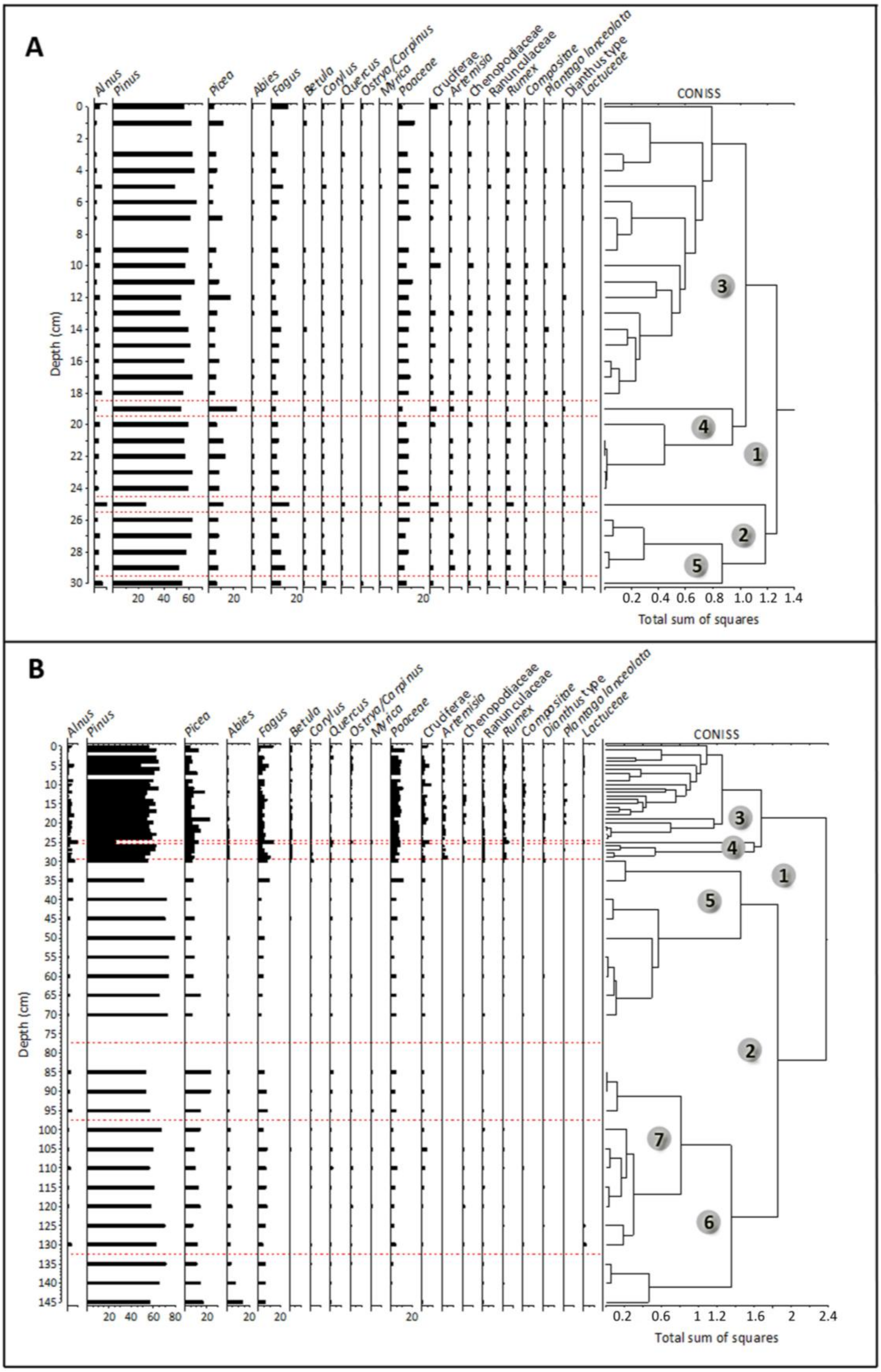


Figure 5.31: CONISS results for core BJC1, analysis results for (A) contiguous samples only and (B) full pollen profile, red dotted line indicates CONISS zone.

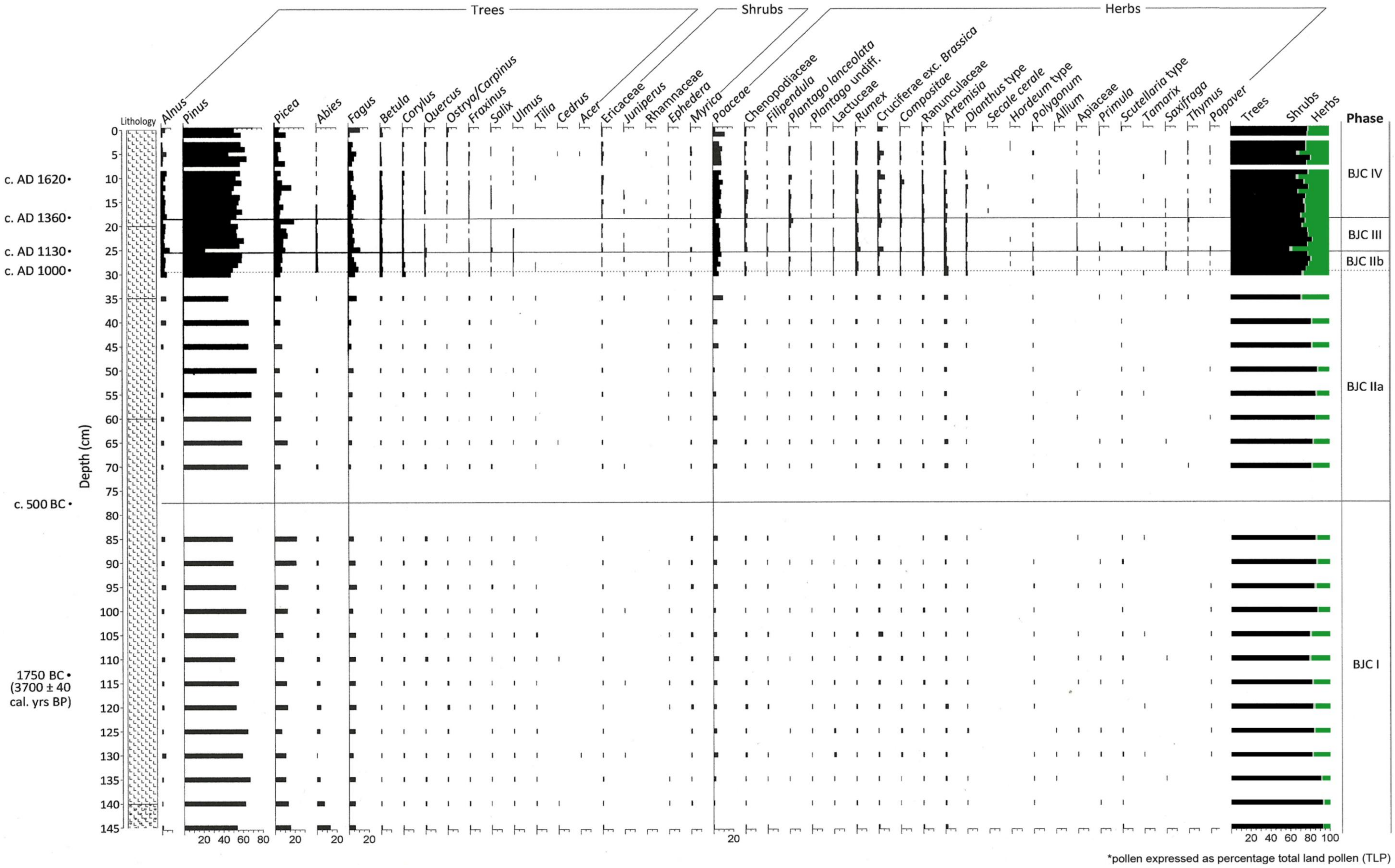


Figure 5.32a: Pollen profile for core BJC1 extracted from Lake C, Buni i Jezerces (arboreal types to herbs).

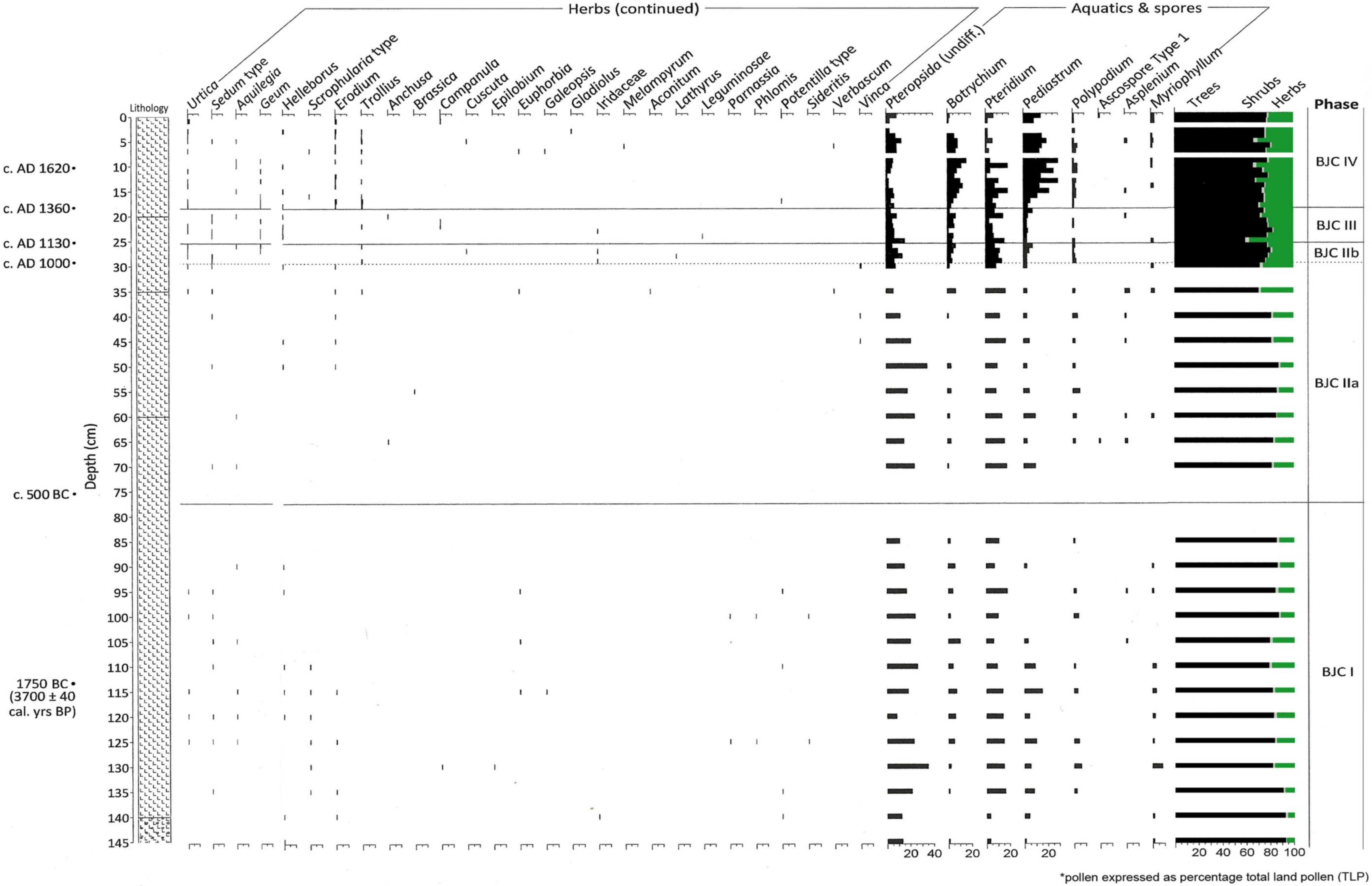


Figure 5.32b: Pollen profile for core BJC1 extracted from Lake C, Buni i Jezerces (herbs continued with aquatics and spores).

Arboreal pollen, and in particular coniferous types, dominated the palynological profile of core BJC1 at 64 – 95%, see Figure 5.32.

Phase BJC I (145 – 77.5 cm) c. 2700 – 590 BC

Arboreal types are particularly dominant at the beginning of this phase (***c.*** 2670 BC) at 95% being primarily composed of *Pinus*, *Picea* and *Fagus*. *Pinus* is renowned for its high pollen production and thus, the 49 - 64% *Pinus* abundance in this phase is unlikely to represent such dominance of *Pinus* within the local vegetation. This is because large quantities of *Pinus* pollen grains are produced and have a long dispersal range, causing over-representation within pollen diagrams (Briks, 2003). *Picea* was the second most abundant type (<22%) and in this region indicates high-altitude montane forest, alongside *Abies*. Any such montane forest in the area is likely to have been dominated by *Picea* as *Abies* reached <1.8% throughout this phase. The remaining arboreal types, except *Alnus* and *Fagus*, maintain a relatively constant abundance ~0.8%, with a sporadic presence of *Cedrus* and *Acer*. *Alnus* prefers wetland floodplain type areas, but its presence remained <4% most likely due to the altitudinal situation of Lake C. Deciduous woodland is indicated by the presence of *Fagus*, whilst types such as *Ulmus*, *Tilia* and *Quercus* suggest the possible presence of 'closed deciduous woodland' in the region. A minor shrub component is apparent with *Myrica* most dominant, but *Juniperus* is rarely present and *Rhamnaceae* is absent. The taxonomic diversity seen within the herbs suggests a variable alpine pasture throughout the area dominated by *Poaceae*. At approximately 2265 BC herbs double in abundance, coincident with a slight waning of *Pinus*. A limited recovery of *Pinus* occurred but herbs remain around 18%. Lactuceae forms a minor component of the pollen sum at <2%, unsurprising given the dominance of arboreal types and *Lactuceae* being an indicator of a 'largely tree free environment' (Collins *et al.*, 2005). Anthropogenic activity within the locality is limited with *Plantago lanceolata* intermittently present (~0.2%) and no cereal types registered. Aquatics and spores were dominated by *Pteropsida* (undiff.), *Pteridium* and *Pediastrum*. A number of studies have used *Pediastrum*, which inhabits lakes and ponds, as an indicator of water depth, temperature, water pH and nutrient levels (Veski, 1994; Edwards *et al.*, 2000; Weckström *et al.*, 2010). The end of BJC I was defined by rising levels of *Picea* and a slight waning of *Pinus*.

Phase BJC II (77 – 25 cm) c. 590 BC – AD 1130

At the end of BJC I and beginning of BJC IIa two samples encountered similar problems to those of the lower catchment and after prolonged HF treatment samples CA 40 (80 cm; ***c.*** 570 BC) and CA 39 (75 cm; ***c.*** 410 BC) were found to be non-polleniferous. This may have been the result of the lake drying out causing conditions to become aerobic leading to grain degradation. A suggestion corroborated by the absence of *Pediastrum* prior to ***c.*** 570 – 410 BC, which following previous research suggests low lake level (e.g. Harrison and Digerfeldt, 1993; Edwards *et al.*, 2000; Sarmaja-Korjonen *et al.*, 2006). LF magnetic susceptibility at this time fell to between 58.1 - 46.1 $x10^{-9}$ m^3 kg^{-1}, suggesting reduced erosion and thus a drier climate which may result in lake evaporation.

Phase two is similar to BJC I with arboreal pollen > 80% and composed primarily of *Pinus*. More equal abundances of *Picea* and *Fagus* exist and the peak in *Picea* at 65 cm (***c.*** 88 BC) may be the result of a fall in pollen concentration to 34,426 g cm^3. The presence of *Abies*, *Salix*, *Tilia* and *Ulmus* becomes intermittent and for the latter two types this is indicative of climatic deterioration. Ericaceae and *Myrica* remain stable at <1%, with *Juniperus* and *Rhamnaceae* remaining sparse. A slight reduction in the diversity of herb pollen is apparent, but generally levels remain similar to BJC I and increase at the end of BJC IIa as a result of rising *Poaceae* abundance. The first evidence of arable farming is indicated by the presence of *Brassica* (<0.2%), but due to the very low and sporadic abundance this is likely to be from a remote source. There is continued evidence of pastoral activity, with *Plantago lanceolata* present throughout the phase. The addition of Ascospore type 1 to aquatics and spores, coincided with the sporadic abundance of *Myriophyllum*. Levels of *Pediastrum* fell slightly, but the taxon was present throughout this phase. Overall, BJC IIa was characterised by reduced taxonomic diversity even though total pollen concentrations remained relatively stable (***c.*** 50,000 g cm^3).

Pollen concentrations collapsed at the beginning of BJC IIb to 3,500 g cm^3 ***c.*** AD 1000 before recovering to previous levels by ***c.*** AD 1030. This may have been the result of climatic downturn or increased sedimentation rates, but the sedimentary data available do not shed any light on the cause of this collapse. Around AD 1000 the pollen assemblage changes little, with similar levels of both arboreal and herb types maintained. This is surprising as significant reductions in pollen concentrations generally have the effect of reducing taxonomic diversity. Therefore, it is difficult to determine why the pollen concentration fell so abruptly. During BJC IIb levels of *Artemisia* rose as *Fagus* declined into BJC III. The diversity of aquatics and spores was lost as Ascospore Type 1, *Asplenium* and *Myriophyllum* were absent, indicative of a possible reduction in lake productivity. Arable farming maintained a sparse presence with minor abundances of *Hordeum*-type registered twice in BJC IIb (<1%).

Phase BJC III (25 – 19 cm) c. AD 1130 – 1330

This penultimate phase represents ***c.*** AD 1130 – 1330 and began with a brief *Pinus* 'collapse' to 22.1%, similar to the collapse of pollen concentrations previously. Given the nature of this shift, flanked by stable *Pinus* levels, it is not thought to be the result of deforestation. Rather this may be a product of short lived period unfavourable conditions for *Pinus* trees in the area or a change in dominant sediment source area. *Alnus*, *Fagus*, *Artemisia* and Cruciferae (exc. *Brassica*) all increase alongside *Pteropsida* (undiff.) and *Pteridium* at this time, suggesting that a possibly wetter environment was conducive to expansion of both deciduous woodland and alpine meadow. The remainder of BJC III is characterised by a near absence of *Ostrya/Carpinus* and *Tilia* is absent from the profile for the first time. *Picea* peaked at the end of BJC III coinciding with a *Poaceae* reduction (***c.*** AD 1330), which following the earlier collapse of *Pinus* suggests that conditions were variable, but sometimes favoured coniferous types.

Phase BJC IV (19 – 0 cm) c. AD 1330 – present

The final phase (***c.*** AD 1400 – present) remains dominated by *Pinus*, *Picea* and *Fagus*, but levels of *Quercus*, *Ostrya*/*Carpinus*, *Tilia* and *Ulmus* all decline with some becoming intermittent. This indicates a change in the local environment and most likely climate as *Quercus*, *Ostrya*/*Carpinus*, *Tilia* and *Ulmus* are described as thermophilous types in Table 6.4 (Chapter 6). Levels of *Poaceae* remain relatively stable, but anthropogenic indicators become more varied with small quantities (<2%) of both *Secale cereale* and *Hordeum*-type identified. Types such as *Plantago lanceolata*, *Rumex* and *Chenopodiaceae* indicate anthropogenic activity, which increases throughout this final phase. In contrast to the lower catchment record, anthropogenic indicators in core BJC1 compose <7% TLP. Taxonomic diversity within the herb types returns somewhere closer to that of BJC I and within the aquatics and spores evidence for increasingly open water and possibly greater precipitation was evident from the much expanded levels of *Pediastrum* (16 – 0 cm; ***c.*** AD 1430 - present). Overall, the landscape presented by the pollen assemblage of BJC IV indicates limited anthropogenic modification and arboreal types continue to dominate the assemblage.

5.7 RECONSTRUCTION OF THE MAJA E KOLJAET GLACIER

This section is divided into three sub-sections, with the first aiming to describe the current Maja e Koljaet glacier (Figure 5.33) using field description and geomorphological mapping. Section two discusses past glacial activity at the site, the lichenometric dating used to date past glacial extents and past ELAs of the Maja e Koljaet glacier. Finally, palaeoclimate modelling results are discussed in section three to provide an independent climate record for the Maja e Koljaet glacier and Lake Plav catchment.

5.7.1 The Maja e Koljaet glacier and glacial geomorphological mapping

A large snow and ice body was identified in September 2009 at the head of Buni i Jezerces (Valley of Lakes) in a north east facing cirque below the Maja e Koljaet peak (2490m a.s.l.; Figure 5.33). Inspection of perimeter exposures and surface pits confirmed that the core of the body was composed of ice and that is was not a perennial snow patch (Figure 5.34).

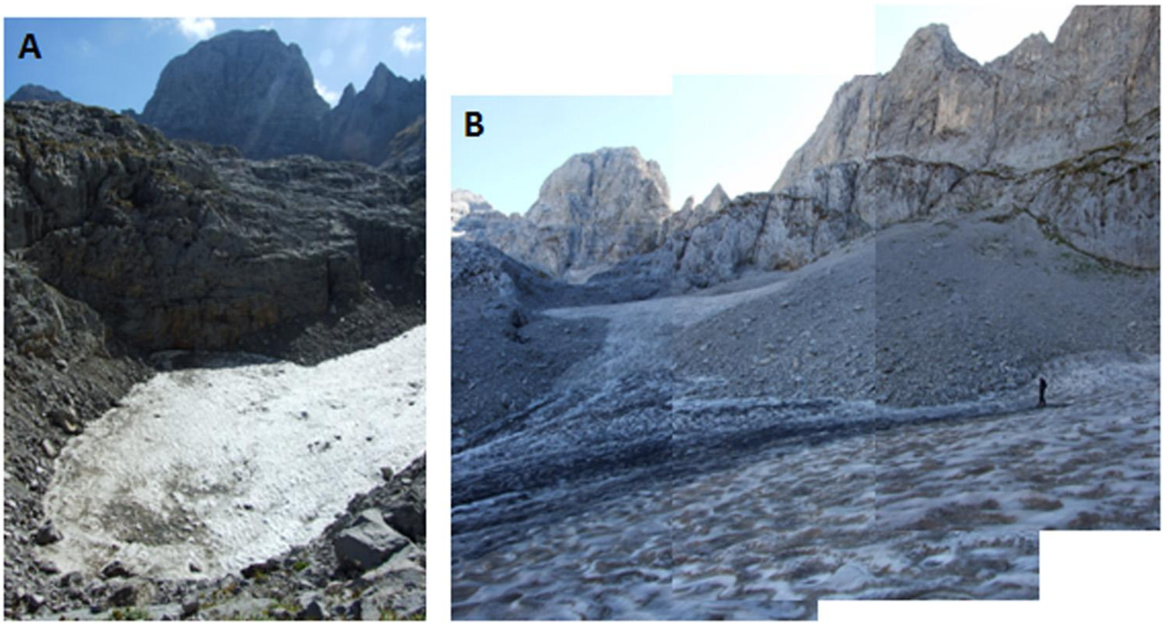


Figure 5.33: (A) Intermittent deformed debris bands visible along the snout of the Maja e Koljaet glacier, (B) Maja e Koljaet glacier looking towards Maja e Jezerces (October 2009).

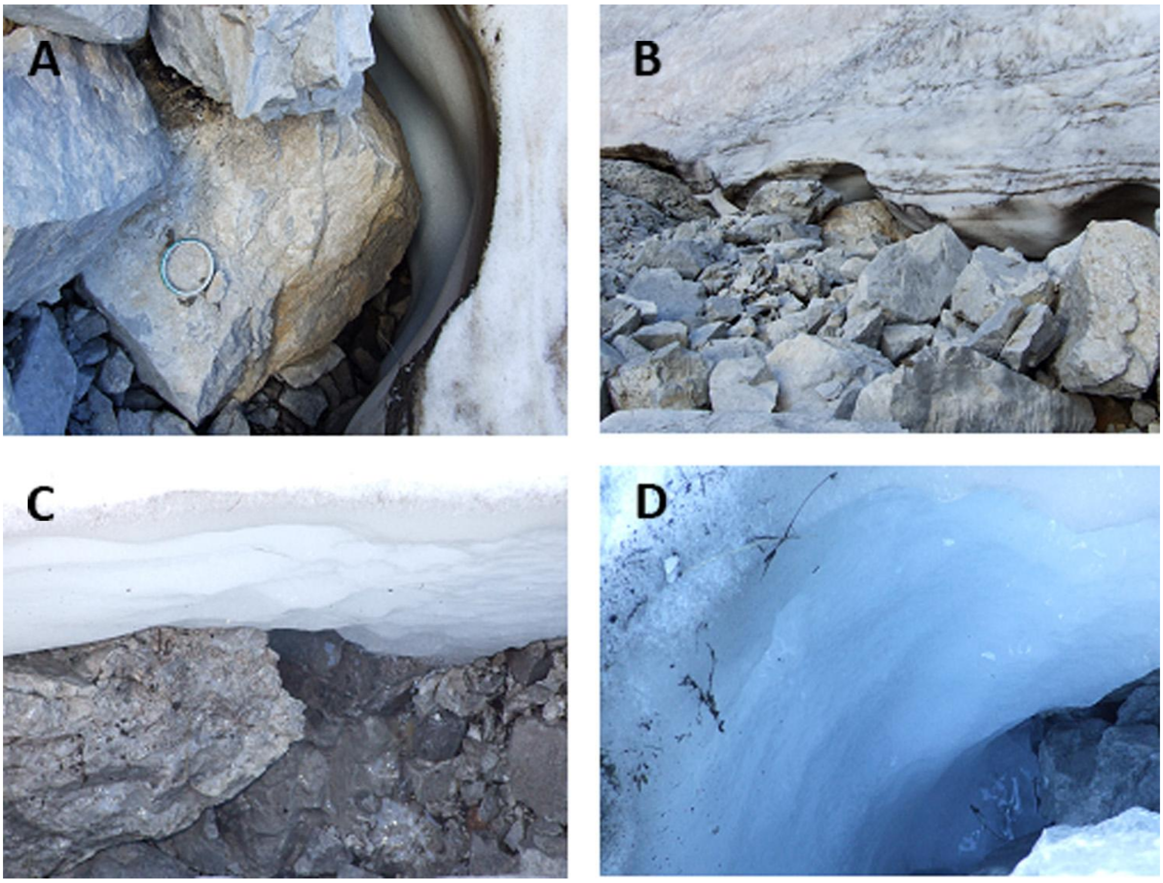


Figure 5.34: Photographs taken along the back wall edge (A, B) and at the current snout of the glacier (C, D October 2009).

The glacier had been recognised by Milivojević *et al.* (2008) who suggested the presence of an active glacier at the site in 2007. During 2009 the glacier lacked crevassing and intermittent debris bands visible at the snout of the glacier suggested that the ice body was inactive (Figure 5.33). However, moraine crests at the snout of the glacier were not colonised by lichen, confirming that the glacier had recently been active in agreement with the observations of Milivojević *et al.* (2008; Figure 5.35). The glacier is situated in a north to north east facing cirque between 1993 m and 2029 m a.s.l. (42^{o} 27’ 12.5”N, 19^{o} 47’ 52.8”) well below the current snow line, covering an area of *c.* 8.4 ha.

Geomorphological mapping of Maja e Koljaet (Figure 5.36) identified four clear frontal moraine ridge crests ahead of the glacier, with the two bordering the lichen free snout. Trim-lines either side of the glacier revealed the maximum vertical extent of glacial expansion at the site, while two lateral moraine ridges provided evidence of the course taken during past glacial expansions (Figure 5.35). Towards the base of the cirque a doline depression exists and at the valley floor, ‘hummocky’ moraines were identified (Figures 5.35; 5.37). Overall, the glacier had a northerly orientation, but progressing down valley this shifted to a more north easterly direction.

Vegetation around the glacier in 2009 was photographed (Figure 5.38) to provide some idea of the recent extent of glacial expansion with the line of vegetation following that of the lichen-free moraine (dark blue moraine in Figure 5.36) and along to a lateral moraine (no. 4 Figure 5.36).

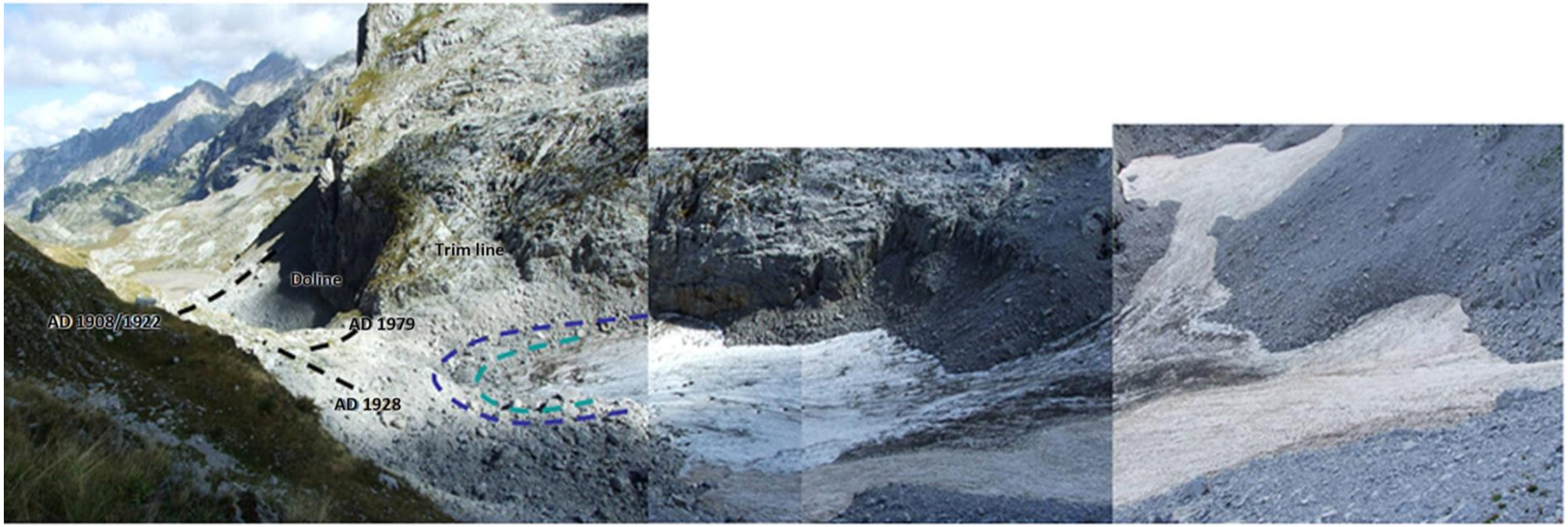


Figure 5.35: Past extents of the Maja e Koljaet glacier, looking down valley to Buni i Jezerces and the Ropojana Valley (October, 2009). The blue and turquoise dashed lines represent the lichen free moraines identified in Figure 5.36.

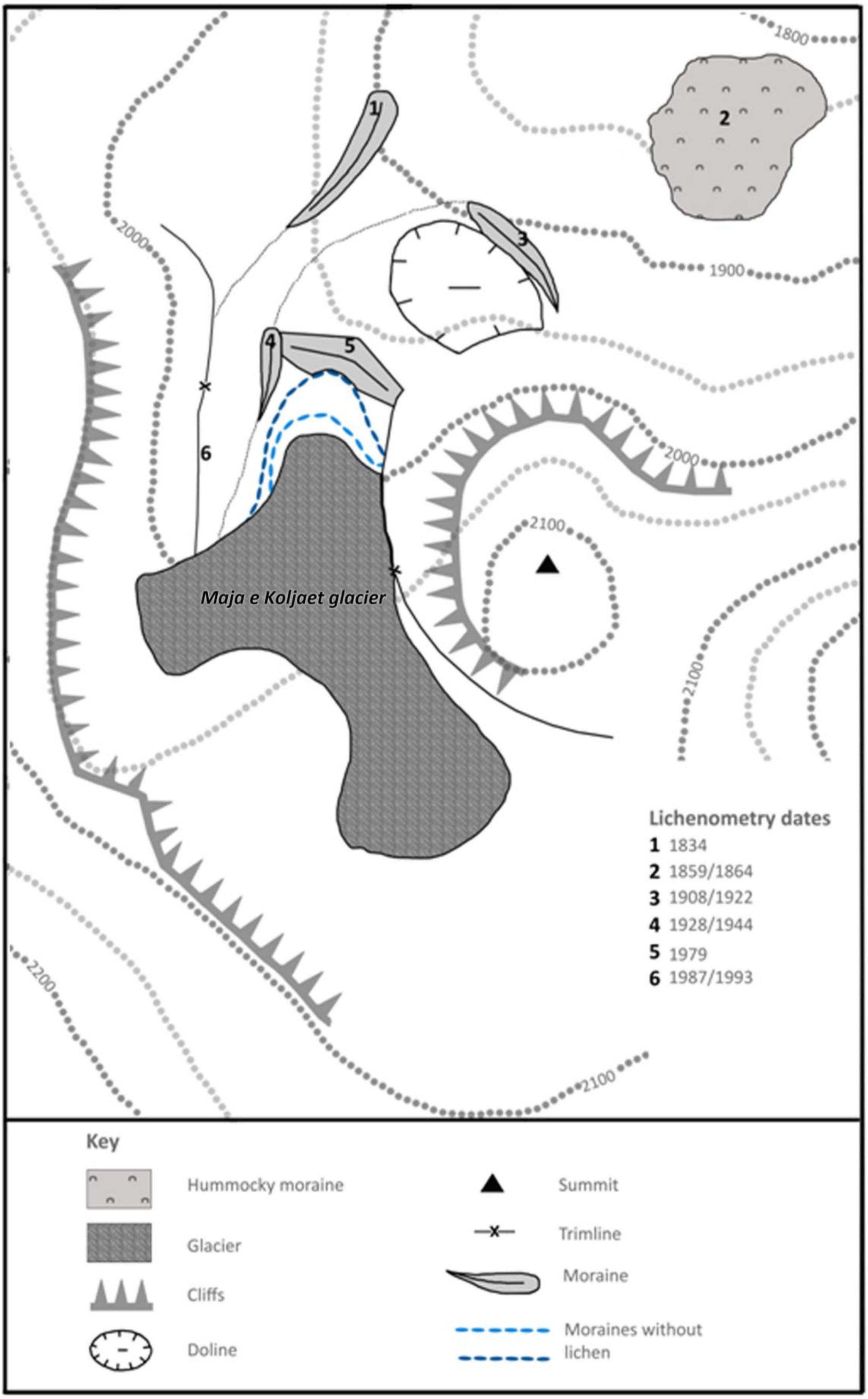


Figure 5.36: Geomorphological map of Maja e Koljaet glacier.

Figure 5.37: Maja e Koljaet glacier from Buni i Jezerces, hummocky moraines visible in the valley below (October 2009).

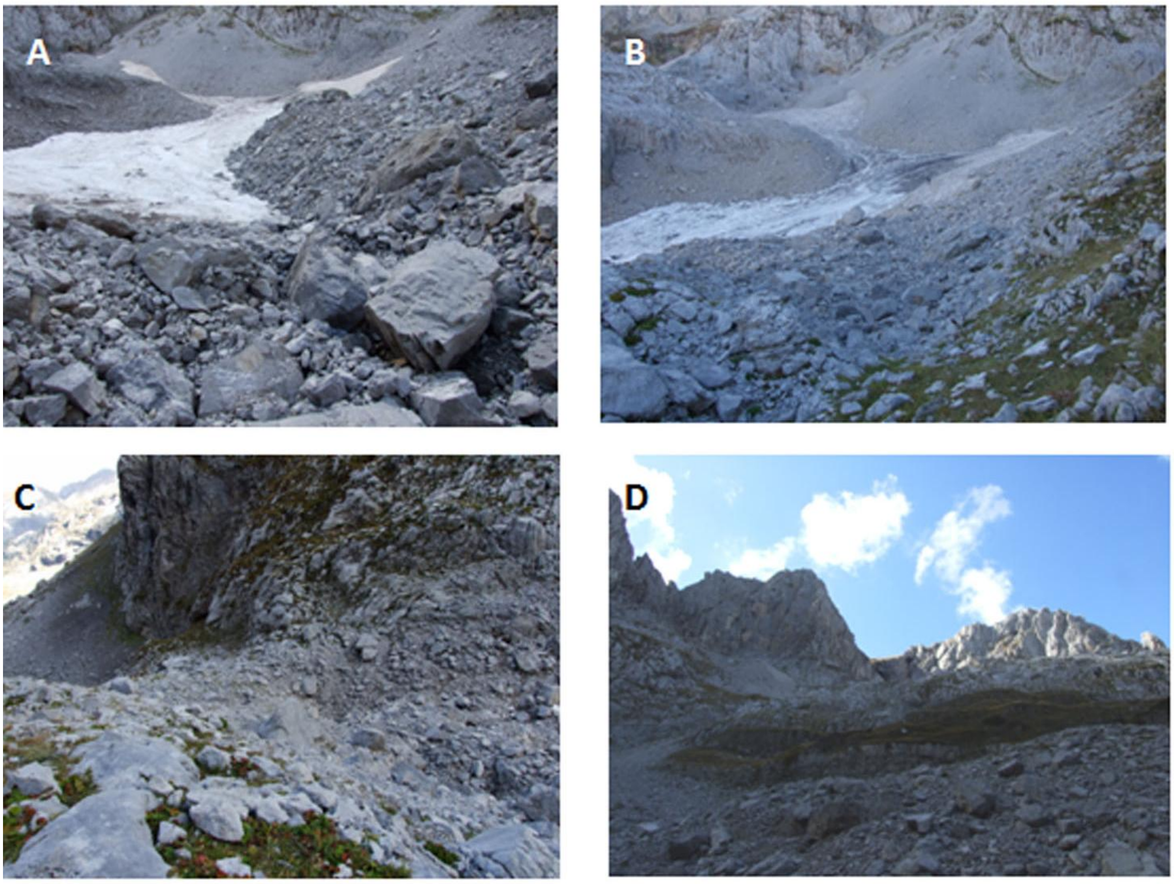


Figure 5.38: Vegetated debris around Maja e Koljaet glacier, 'A' represents close proximity to the glacier, but B to D illustrate clear lines of vegetated land above the glacier and deglaciated debris.

5.7.2 Evidence of past glacial activity

The glacial history of the site was identified with the use of geomorphological mapping and lichenometry to calculate the age of any glacial features recorded. Lichenometric dating was undertaken on six glacial landforms and the diameter of 76 lichens documented, with the intention of calculating the mean of the five largest lichens per landform. However, the colonisation of lichen was not as prolific as expected, with many boulders being free of lichens, or lichens had merged making measurement unreliable (see Figure 5.36). In total, lichens on 11 boulders were satisfactory for lichen analysis to assess the mean of the five largest lichens, but for glacial landform no. 1 (Figure 5.36), a long north-facing lateral moraine, the mean of the three largest lichens were completed due to sparse lichen colonisation. Table 5.16 presents the lichen data collected for each glacial landform examined, and the bracketed letter indicates the boulder used, from a ridge location, for lichen measurement. Although the number of boulders analysed was low, the data presented does use the standard 'mean of the five largest lichens' for all but one landform and the sparse lichen colonisation could not have been foreseen. The lichen data were then collated and inputted into the lichen growth curve developed for Durmitor Massif 100 km north-west, in Northern Montenegro, by Hughes (2009b): Y = -8.26 + 0.74X (presented in Chapter 4). This enabled the age of the lichen measured to be deduced and the age was then taken from AD 2009, the year the samples were measured (Table 5.16).

Location (number relates to Figure 5.26)	Average lichen size (mm)	Age of lichen (years)	AD 2009 minus lichen age
6 (A)	3.88	16.40	1993
6 (B)	7.83	21.75	1987
5	13.60	29.54	1979
4 (A)	52.00	81.43	1928
4 (B)	41.00	66.57	1942
4 (C)	40.00	65.22	1944
3 (A)	66.80	101.43	1908
3 (B)	55.80	86.57	1922
2 (A)	98.80	144.68	1864
2 (B)	102.50	149.68	1859
1	121.00	174.68	1834

Table 5.16: Lichen records from Maja e Koljaet glacier.

Figure 5.39: Past glacial extents of the Maja e Koljaet glacier.

Past glacial extents of Maja e Koljaet glacier were then constructed following the position of the mapped moraines and the surface ages of the moraine boulders across the site (see Figure 5.39). Four discrete glacier extents can be identified. The first and largest glacier phase covered an area of 53.8 ha at the site and extended down into the Buni i Jezerces valley floor, over the steep rock step visible in Figure 5.37. This would have been characterised by an ice fall, or heavily crevassed glacier surface, at that time. The second and third extents lay above the steep rock step and either side of the doline depression, with the final, fourth, extent representing the glacier position in 2009. The glacial history developed for the Maja e Koljaet glacier based on the geomorphology and lichenometry showed that the largest reduction in glacier area was in the 19th century (between ***c.*** AD 1800 - 1900). During this time the glacier retreated up valley to the currently visible doline depression reducing in size by around 35.5 ha. This was possibly caused by its more exposed situation at the valley floor and a lack of localised shading and avalanching, unlike the later position of the glacier, closer to the cliff back-wall (see below). In the subsequent 75 years, through the 20th century, the glacier receded close to its current position, before re-treating a final time around the late 1980s/early 1990s to its current situation. However, between the early 1900s and 2009 the glacier surface area diminished from 18.3 ha to 8.4 ha (-9.9 ha). This was a significantly smaller and less abrupt transformation than the changes from the 1800s to 1900s, although the glacier surface area still decreased by more than 50%. The doline depression may have slowed the AD 1908 - 1979 retreat of the glacier as it would have been conducive to the accumulation and preservation of ice. Furthermore, the close proximity of the back-wall cliffs below Maja e Koljaet peak would have sheltered the site from solar radiation and avalanching would have made a major contribution to annual accumulation.

The median elevation of the glacier at each extent was used to estimate the ELA at the time of each different glacial extent (see Chapter 4). The median glacier elevation divides the glacier surface into two equal parts and reflects the statistical median of the glacier, surface area-altitude distribution (see Figure 5.41). The latter was achieved by extrapolating surface contours onto the different ice extents from base maps (1:50,000) and calculating areas between the different contours on the glacier surface using ImageJ

(Image Processing and Analysis in Java) as seen in Figure 5.40. The lichen dates associated with a particular landform were assigned to the corresponding median and frontal moraine elevation to allow palaeoclimatic modelling of the site (Table 5.17).

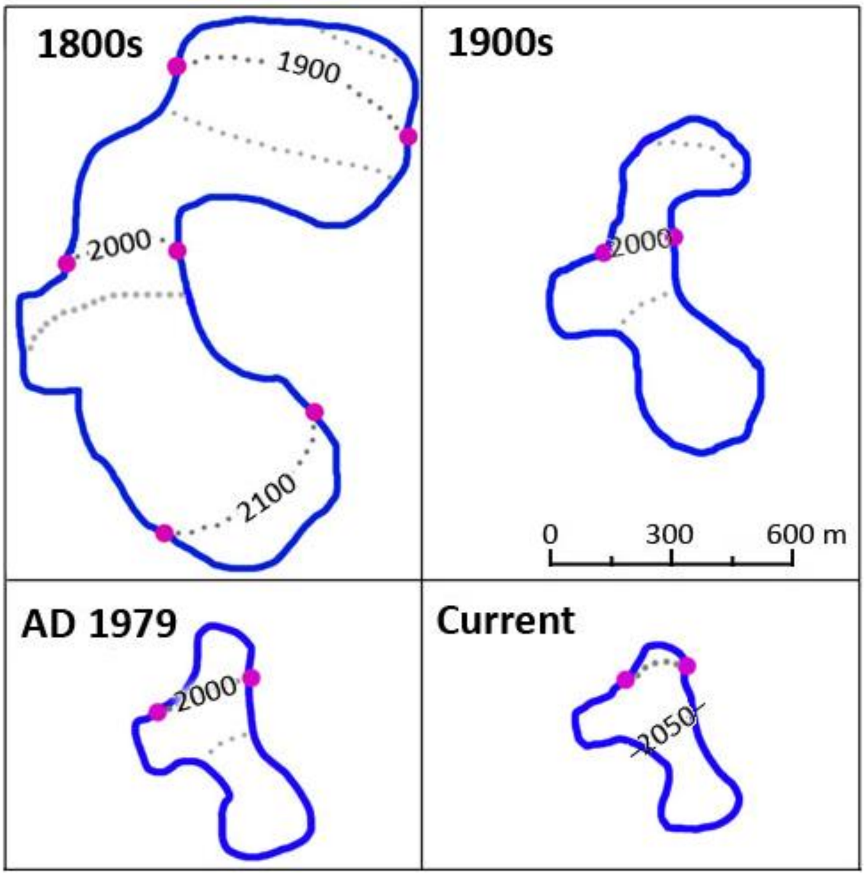
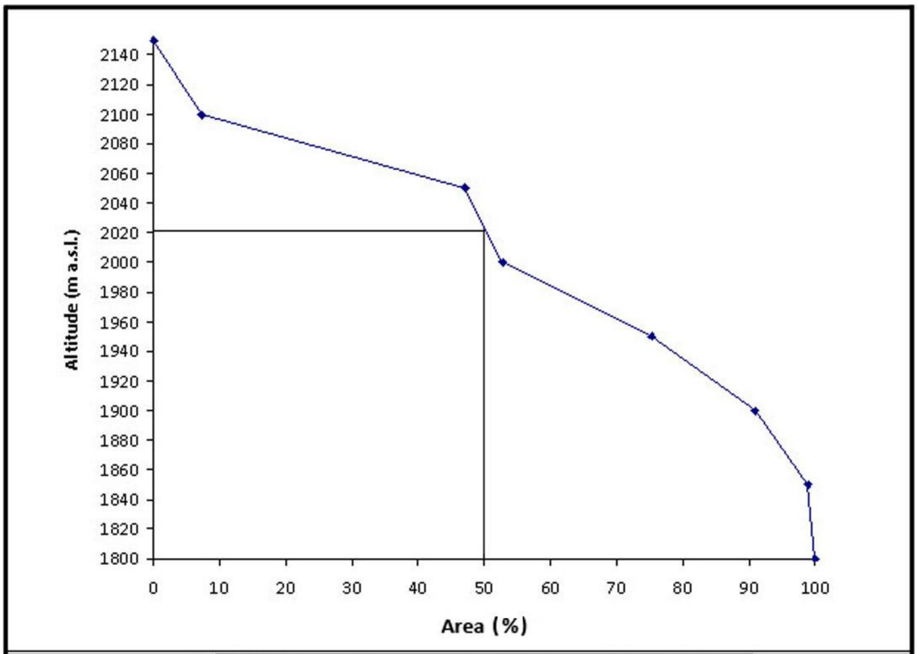


Figure 5.40: Glacier outlines used in ImageJ to calculate the areas and thus ELA of each glacial extent, A: 1800s, B: 1900s, C: 1979; D: current and altitudes on dotted lines (m a.s.l.)

AD 2009 minus lichen age	Median elevation (m)	Front elevation (m)
1859	2025	1800
1908	2045	1900
1979	2045	1950
2009	2040	1950

Table 5.17: Median and frontal elevations of the Maja e Koljaet glacier, reconstructed using lichenometically dated moraines.



Altitude (m a.s.l.)	Area	% area	Cumulative %
1800			
1850	0.006	1.1	100.0
1900	0.043	8.0	98.9
1950	0.084	15.6	90.9
2000	0.121	22.5	75.3
2050	0.031	5.8	52.8
2100	0.214	39.8	47.0
2150	0.039	7.2	7.2
TOTAL	0.538	100.0	0.0

Median elevation	Front elevation
2025	1800

Figure 5.41: Example of median elevation calculation for the 1800s glacial extent using the estimated area covered by each glacial portion.

5.8 PALAEOCLIMATE MODELLING OF THE LAKE PLAV CATCHMENT

Following the geomorphological data presented for the Maja e Koljaet glacier palaeoclimatic reconstructions can be made for the different ELA positions during each ice extent. These can then used to understand climatic fluctuations in the Lake Plav catchment. However, as discussed in section 3.3.8 (Chapter 3), identifying the ELA of small high mountain glaciers can be problematic. Table 5.17 highlights the minor variation in median elevations, which differed by only 15 m and was unrepresentative of the glacial variations reconstructed using geomorphological mapping (with the larger AD

1908 and AD 1979 glaciers appearing to be higher than the smaller AD 2009 glacier). The reason for this is that, for small glaciers with limited area-altitude distributions, any significant change in the shape of the glacier surface will affect the median elevation of the glacier. This results in the unusual situation whereby a larger glacier with lower fronts has a higher median altitude than a smaller glacier with higher fronts. Therefore, the front elevations for each past extent were used during modelling as they best characterised the glacial variations since the 1800s at Maja e Koljaet. As previously discussed, climate data from Kolašin was used during calculations (Table 4.5, Chapter 4) and this section discusses the results produced using a Degree-day Model (Brugger 1996; Hughes, 2008, 2009, 2010) and the Ohmura *et al.* (1992) regression to reconstruct past temperatures at the site.

5.8.1 Degree day model results

The annual temperature range for Kolašin between AD 1973 - 2008 was 19.4^{o}C and the mean annual temperature data for the varying heights are presented in Table 5.18 alongside the frontal moraine elevations of the Maja e Koljaet glacier.

This data was then processed using the equation of Brugger (2006) at each required elevation and the mean annual temperature distributed over a sine curve (Figure 5.42) and a degree day factor of 4mm day^{-1} K^{-1} was applied following Braithwaite *et al.* (2006; full data presented in Appendix VII). Following the completion of the degree day model the annual accumulation required to balance melting at the Maja e Koljaet glacier at its AD 2009 position 1950 m a.s.l. was **6504mm** water equivalent (w.e.). However, the frontal moraine during the LIA was reconstructed to be at 1800 m a.s.l., therefore, assuming that average precipitation has remained the same throughout the LIA to present the contemporary frontal moraine conditions calculated at 1950 m a.s.l., would have occurred at 1800 m a.s.l. during the LIA. Under current conditions a temperature depression of ***c.*** 0.9^{o}C (with mean annual melt/accumulation of **6504 mm** w.e.) would be required for the Maja e Koljaet glacier to extend down valley by 150 m to AD 1832 levels (Table 5.18).

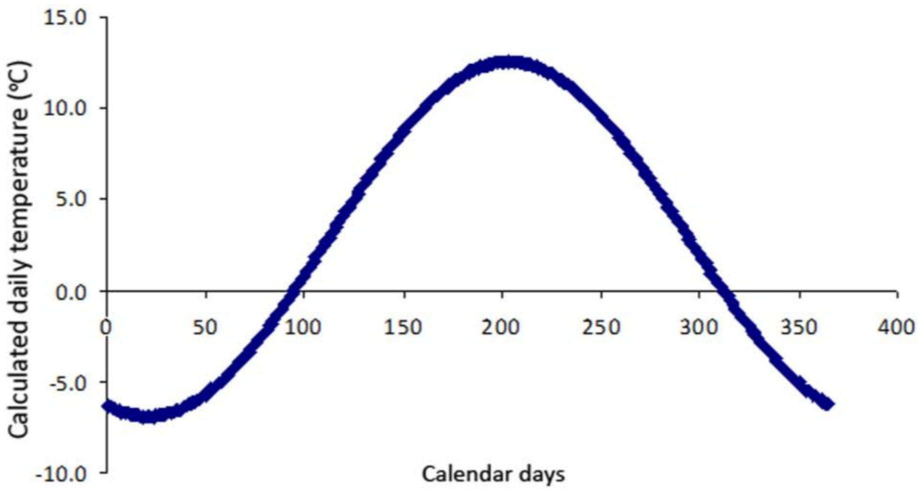


Figure 5.42: Sine curve produced following the processing of a mean annual temperature of 2.82°C relating to AD 1908 moraine.

AD 2009 minus lichen age	1859	1908	1979	2009
Reconstructed frontal moraine altitude (m a.s.l.)	1800	1900	1950	1950
Modern mean annual temp. at this altitude (°C)	3.42	2.82	2.52	2.52
Mean annual temperature at 2009 glacier front (°C)	2.52	2.52	2.52	2.52
Temperature depression (compared with modern data) required to enable glacial advance to frontal moraine altitude (°C)	-0.9	-0.3	0	0

Table 5.18: Degree day model reconstruction for Maja e Koljaet.

5.8.2 Ohmura *et al.* (1992) palaeoclimatic model

The mean June/July/August (J/J/A) temperature for Kolašin between AD 1973 and AD 2008 was 17.2°C and the extrapolated J/J/A temperatures for the varying front elevations are listed in Table 5.19.

These data were then processed following the method of Ohmura *et al.* (1992) and the results suggested that the annual accumulation required to maintain the glacier at its AD 2009 position, with a front elevation 1950 m a.s.l. would be **5101 mm** w.e. (precipitation/accumulation difference of **1403 mm** w.e.) with a mean J/J/A temperature of 11.22^{o}C. As noted for the degree-day model a frontal depression of 150 m would require a temperature decrease of 0.9^{o}C, matching the figures suggested by the degree day model (Table 5.20).

Estimated moraine age (AD)	Front elevation (m)	Mean J/J/A temperature (°C)
1859	1800	12.12
1908	1900	11.52
1979	1950	11.22
2009	1950	11.22

Table 5.19: Current extrapolated June/July/August temperatures for past front elevation heights.

Date (AD)	1859	1908	1979	2009
Reconstructed frontal moraine altitude (m a.s.l.)	1800	1900	1950	1950
Mean J/J/A temp. extrapolated to altitude (°C)	12.12	11.52	11.22	11.22
Mean J/J/A temperature at 2009 glacier front (°C)	11.22	11.22	11.22	11.22
Temperature depression required to allow glacial expansion (°C)	-0.9	-0.3	0	0

Table 5.20: Ohmura *et al.* (1992) model reconstruction for Maja e Koljaet.

This chapter aims to decipher the Lake Plav catchment records to enable discussion of when, how and to what degree the Little Ice Age affected this region of Montenegro/Albania compared to other Mediterranean regions.

6.1 THE LAKE PLAV LOCALITY

Documentary records for the Lake Plav catchment, suggest that the lake has reduced in size since the early 1900s, when Cvijić (1913) mapped the area. A comparison between the Cvijić (1913) map and the most recent Huber Verlag/Lokalna turistička organizacija Plav (2008) map of Lake Plav suggests that the lake was 42% larger in AD 1913 than AD 2008 (Figure 6.1). Although it is not known in which season the two maps were created, the larger extent mapped by Cvijić (1913) is not thought to represent a flood (such as that of 2010 discussed in Chapter 3), but rather the 'normal' lake extent at that time. This is because the River Ljuča maintains the same width as the present river and there is no evidence that the river had overtopped its banks. A remnant of the past extent of Lake Plav could be Malo Bato (Figure 6.1), a small water body now cut off from the main waters of Lake Plav by a seasonally waterlogged area dominated by *Phragmites,* Cyperaceae and *Typha latifolia*. Overall, this information suggests that the lake once covered a far greater area than its present size and progradation of the River Ljuča floodplain led to a reduction in the size of the water body. This is supported by the stratigraphic and palynological records at the site. The infilling of the lake is the product of fluvial sediment supply controlled by climate, catchment erosion and land use which, together, can determine the flux of water and sediment to the lake (Baster *et al.*, 2003). Climatic variance was identified as one of the primary drivers of rising sediment yields in Europe by Esteve *et al.* (2004). Anthropogenic impacts upon a catchment are likely to contribute, as variance in land use types was shown by Poesen and Hooke (1997) to affect the annual soil loss within different Mediterranean environments. Within the Lake Plav catchment sedimentation rates and river sediment loads during the winter months and spring melt season transport large quantities of sediment to the lake. The pastoral farming that dominates the landscape could increase the quantity of sediment available for transportation due to reduced land stability. This was shown in the 2010 study by Ausseil and Dymond, which suggested sediment yields were 50% higher in pasture land (defined as 'erosion prone land') than forested land.

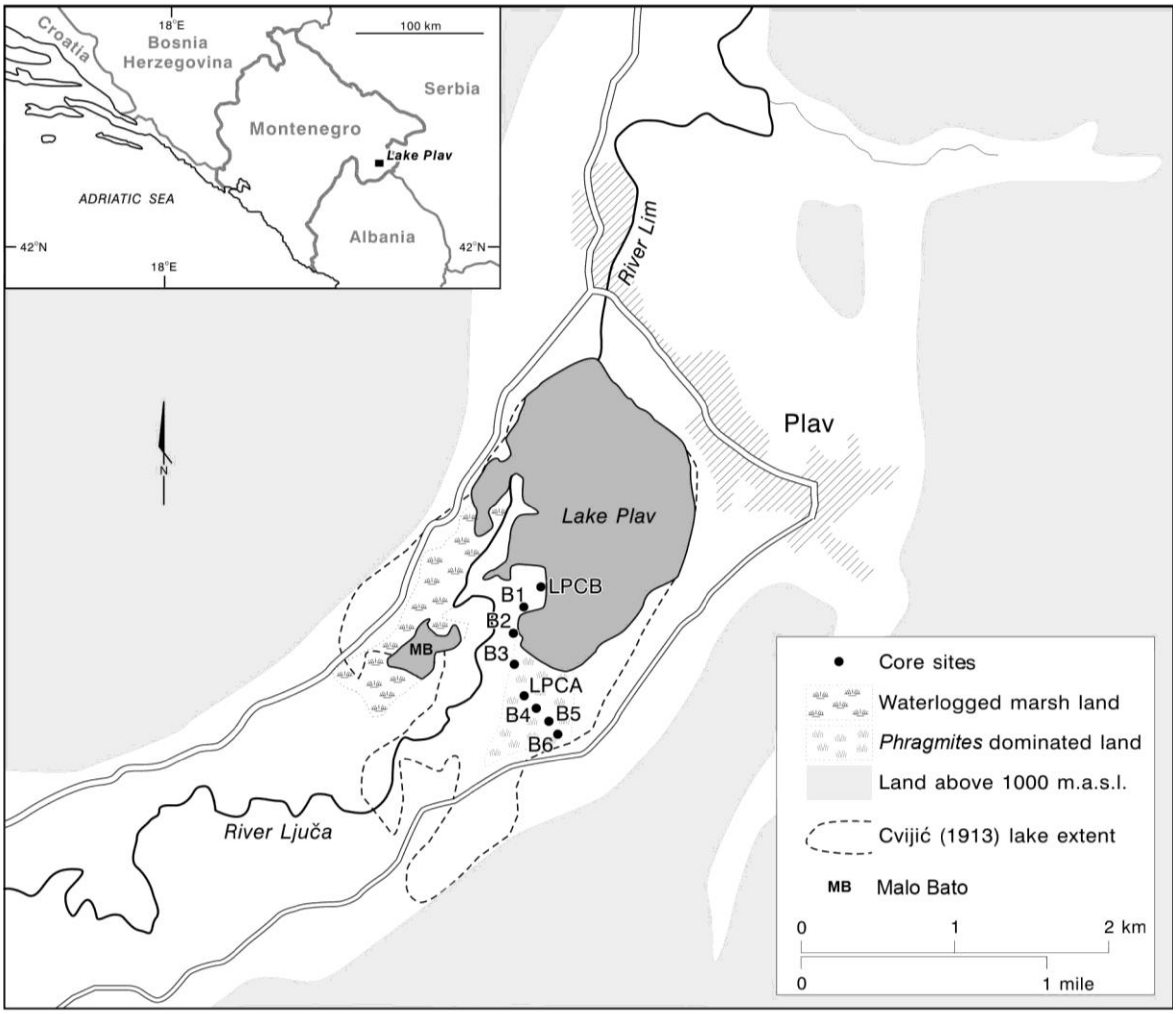


Figure 6.1: Core locations in relation to the past lake extent following Cvijić (1913).

6.1.1 Stratigraphic record

This discussion is presented in two sections, the former discussing Group 1 (cores LPCB, B1, B2) and the latter, Group 2 (cores B3, LPCA, B4, B5, B6; see Figure 5.18, Chapter 5).

The Group 1 cores are dominated by sands and gravels indicative of a high energy environment from ***c.*** AD 1100 to 1880, and are likely to be the result of fluvio-deltaic activity. Research across Europe has suggested that rising lake levels were coeval to the LIA (e.g. Harrison *et al.*, 1993, 1996; Magny, 2004; Holzhauser *et al.*, 2005; Chapron *et al.*, 2007a). Figure 6.1 indicates that the Group 1 cores would have been located further from the lake edge during periods of increased lake level. However, coarser material is associated with the littoral or deltaic areas of a lake and fine grained material with still or profundal lake waters (Hakånson and Jansson, 1983; Kostic and Parker, 2003). Coupled with the high reconstructed sedimentation rates for core LPCB (see section 5.2.2, Chapter 5) it is suggested that this area was under the influence of fluvial inputs and would have greater rates of sedimentation than open water sites (Moore, 1966). Using Figure 6.1, the River Ljuča inflow would have been located at least 400 m from the Group 1 cores and closest to B2, which has the coarsest stratigraphy of all the 8 cores extracted. Alçiçek *et al.* (2006) listed similar lithological characteristics of lacustrine sediments in the Değne member of Çameli Basin, South-Western Anatolia in Turkey, with those of deltaic or fan-deltaic origin being more coarse in nature compared to the open lake and ephemeral lake sediments that were dominated by clay. It is suggested that the Group 1 data provide three cores through deltaic sediments. The model implied is visually presented in Figure 6.2 and suggests that as the lake infilled, the influence of the River Ljuča fell and the build up of deltaic sediments waned, causing progressively finer sediments to be registered. Between AD 1575 and AD 1945 sedimentation rates declined from 1.41 $\pm$ 0.17 cm yr^{-1} to 0.277 $\pm$ 0.03 cm yr^{-1}. The five fold decline in sedimentation rate coupled with change to finer sediments in upper core units, indicates a change from deltaic depositional environment to lake edge environment (Hakånson and Jansson, 1983; Kostic and Parker, 2003), thus corroborating the model implied in Figure 6.2. The delta would have been up to 800 m long and supplied coarse grained material to what would have been central lake sites, following the Cvijić 1913 map (Figure 6.1). Sparse plant and wood remains were found in the cores of Group 1, indicative of allochthonous inputs i.e. organic fragments transported by the River Ljuča and then deposited as part of deltaic sediment. Hutchinson (1957) indicated that in wooded areas, fallen leaves are an important source of allochthonous organic inputs, entering a lake via streams and rivers. An exclusive

Substania humosa layer would have represented periods of stabilisation and soil development (Digerfeldt, 1986). However, particles of *Substania humosa* and organics found in cores B1 and LPCB were within a coarser matrix of silty fine sand which are more indicative of flood deposits (Williams and Roberts, 1990; Collins *et al.*, 2005) and thus transportation via the River Ljuča. This is not unlikely considering research from around the Mediterranean reported periods of increased fluvial activity throughout the LIA (Holzhauser *et al.*, 2005; Macklin and Woodward, 2009).

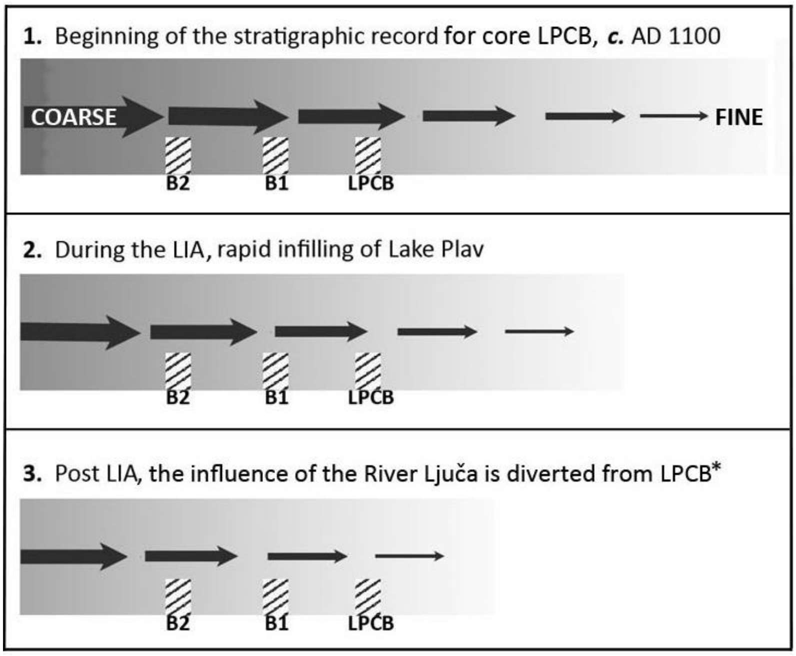


Figure 6.2: Schematic illustration of the waning River Ljuča influence and its associated delta upon the cores of Group 1. *Decline inferred from age-depth model of core LPCB.

From the generalised Group 1 data a possible past route of influence of the River Ljuča can be deduced (see Figure 6.3) and highlights the river progradation and lake infilling that has occurred in line with other river delta contexts across the Mediterranean (Arnaud-Fassetta and Provansal, 1999). This is corroborated by pollen analysis of core LPCB which was found to be non-polleniferous, a possible result of high sedimentation rates (Middledorp, 1982 cited in Bohncke and Wijmstra, 1988). The particle size data of core LPCB shows a progressive fining of material before the sediment composition becomes coarser (Figure 6.4) thought to infer a continual waning/increasing river flow and such inputs are consistent with the heterogeneous nature of the LIA period. Particular events throughout each of the cores within Group 1 can be identified, with approximate dates applied using the LPCB age-depth model. Coarse material dominates Group 1, but around AD 1820 core LPCB registered rising quantities of silt and clay material and the lowest levels of coarse to very coarse sand to date. This is also a turning point for B2 as no material coarser than silt is registered again. Fluctuations between coarse and finer material continue in core LPCB and B1 until ~1.15 m, ***c.*** AD 1870, this may indicate the end of the LIA which would have provided larger floods to allow the rapid infilling of Lake Plav. Alternatively, at this point the River Ljuča ceased to flow towards the centre of Lake Plav and was diverted to the north-northwest course it now takes, following the sedimentation in the Group 1 area. The low organic contents within core LPCB and B2 also suggest an allochthonous sediment origin rather than local macrophyte colonisation.

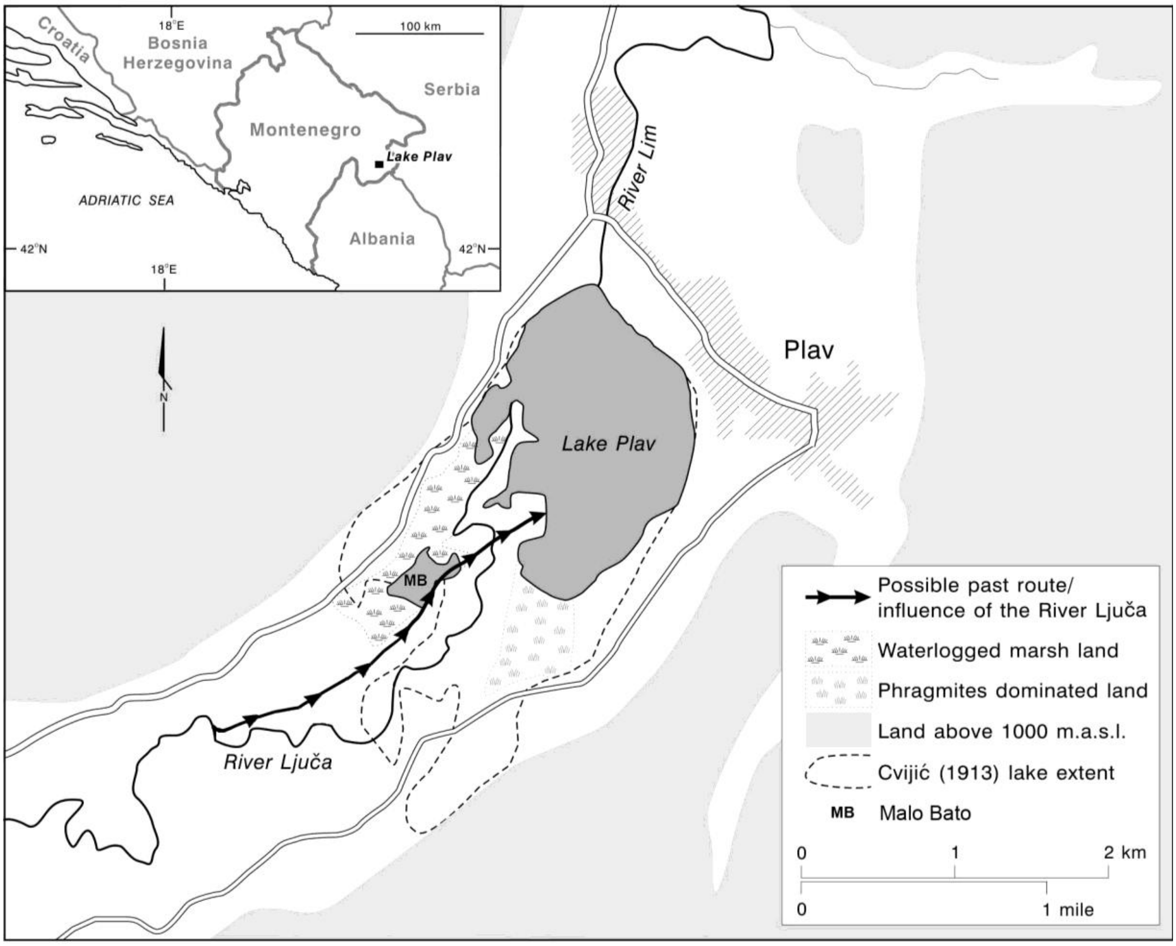


Figure 6.3: Possible past route of the River Ljuča into Lake Plav.

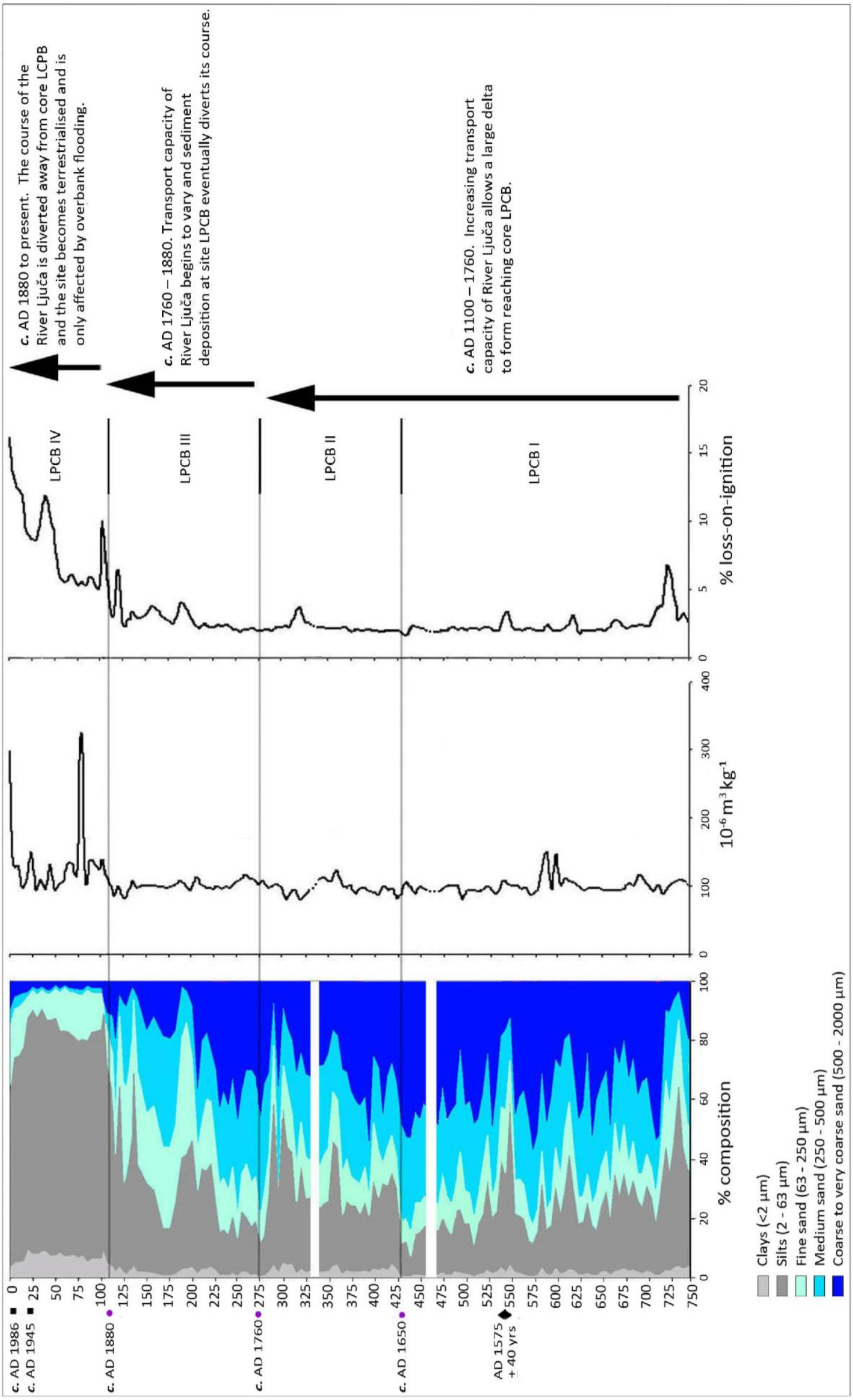


Figure 6.4: Synthesis of the core LPCB data sets.

In contrast the lower units of the Group 2 cores are indicative of a central lake and low energy (possibly deeper lake) environment being dominated by clays (Hakånson and Jansson, 1983; Kostic and Parker, 2003). Following Benvenuti's 2003 study of the Mugello Basin in central Italy, stratified clays were said to be indicative of a waning high energy environment. This would fit in well with the suggestion that the Group 1 cores record deltaic sedimentation and the cores of Group 2 would be located at the edge of such a formation in a waning high energy environment.

Figure 6.1 indicates that the Group 2 cores were located at sites increasingly distant from the central basin area and the AD 1913 course of the River Ljuča. This would account for the generally finer sediments identified within the Group 2 cores. The soil development noted in B4 to B6 would be consistent with the previous indication of lake infilling and shrinking, as these sites were located on the outer lake limits where infilling would, over time, cause terrestrialisation and thus soil development. Overall the Group 2 cores (LPCA; B3; B4; B5; B6) are all located at sites that would have been immersed in AD 1913, following Cvijić, whereas, today the sites are only seasonally flooded. A reverse pattern of coarsening upwards within the Group 2 cores (in contrast to Group 1) is indicative of increasing proximity to the lake periphery and inward displacement of the lake shoreline, following the Digerfeldt (1986) model of lake sedimentation. Although plant fragments were found throughout the Group 2 cores, they were sparse and generally <2 mm in size. This suggests that, rather than being locally sourced, they were brought to the site via fluvial inputs i.e. the River Ljuča. Detrital peat and silty loam identified in B4, B5 and B6 are indicative of possible vegetation colonisation and surface stabilisation. Although no such sediments were found within core LPCA there is palynological evidence to suggest wetland expansion during the 1500s – 1800s. This fits the B4, B5 and B6 trend, with tentative application of the LPCA age depth model to these cores. Following the period of possible vegetation colonisation, B3 and LPCA register periods of fine to medium sand between ***c.*** AD 1780 - 1880s for LPCA and ***c.*** AD 1770 - 1860 for B3 (using the LPCA age depth model). B3 registers a thicker unit of sand compared to LPCA, where sand fractions are interspersed with 17 cm and 5 cm silty clay units. This is thought to be the result of the LPCA location further from the deltaic area created by the River Ljuča. Although B3 is dominated by medium sand between 100 – 124 cm, thin <3.5 cm silty loam organic horizons were identified at four points (see Appendix IX) within this medium sand unit, suggesting possible local erosion as a result of land disturbance. Both the Group 1 and 2 data provide evidence for the changing fluvial inputs to Lake Plav at the end of the 1800s, with the River Ljuča course shifting towards its present day route at this time. This is corroborated by the AD 1939 map of the area (Figure 6.5) which indicates a similar location of the River Ljuča compared to the present day. The variability in the records particularly around the late 1800s coincides with the end of the LIA across the Mediterranean region and Europe, which has been attributed to supplying increased seasonal melt water and sediment load to terminal areas, such as Lake Plav, enabling infilling.

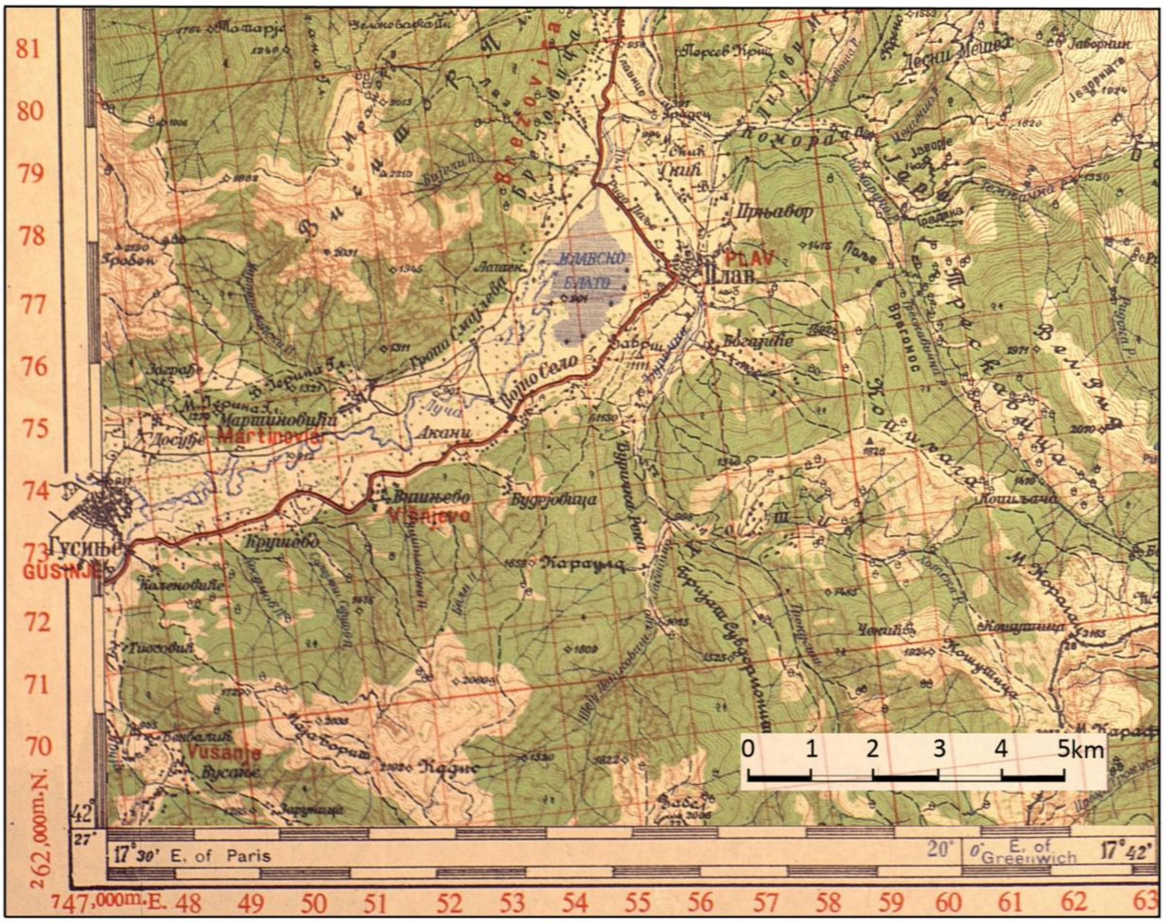


Figure 6.5: The area around Lake Plav, AD 1939. Printed by the Geographical section general staff (no. 4396) War office 1943, copied from a Yugoslavia map dated 1939, heliographed by OS 1943. *ПЛАВСКО БЛАТО (Lake Plav).

The upper sediments throughout Group 2 are indicative of a lower energy environment, similar to those of Group 1, as silts dominate. This supports the previous suggestion that the influence of the River Ljuča has been diverted towards Visitor Mountain by ***c.*** AD 1910 and in the years following the Cvijić (1913) map the lake level fell revealing the diversion (Figures 6.3, 6.5). Finally, the dominance of relatively inorganic silt in the uppermost sediments of 6 (of 8) cores extracted, is indicative of a floodplain site with some seasonal overbank flooding (**cf.** Sturm, 1975 section 6.1.3).

Sedimentation rates across the Lake Plav transect varied (using the LPCA and LPCB age-depth models); with the LPCB core indicating that rates of deposition at ~1.41 cm yr^{-1} during AD 1575 ± 40 yrs. The age depth model created for core LPCB indicate that sedimentation rates were high until the late 1800s/early 1900s, a pattern seen across Southern Europe (e.g. Macklin *et al.*, 1995; Devillers and Provansal, 2003) and supported by the core LPCB age depth model. At Lake Plav this change is seen as evidence for the gradual infilling of the lake and thus position of the inflow delta throughout the study period. Higher sedimentation rates indicate greater fluvial inputs to the site, which caused lake infilling and possibly the avulsion of the River Ljuča inflow around the lake to its present location (see Figures 6.1, 6.3, 6.5). As the lake has infilled, the core sites have been periodically inundated by shallower water, with cores on the very edge of the lake (B5 and B6) becoming terrestrial. Sediment supply to these sites has been reduced as the River Ljuča transports sediment away from these sites and towards Visitor Mountain during the 20th and 21st centuries.

6.1.2 Particle size data for core LPCA

The LPCA particle size record is dominated by silt, indicative of a location between central lake waters and the littoral zone, where exclusively fine grained material dominates the former and coarse material characterises the latter (Digerfeldt, 1986; Stine, 1990; Figure 6.6). The small component of clay material (<10%) throughout the core suggests that at no point was core LPCA beneath a significant volume of standing water i.e. a profundal or central lake situation. The constant supply of coarse minerogenic material is consistent with the suggestion that the site has been situated close to deltaic sedimentation throughout its history. This is because deposition close to the river mouth is represented by coarser sediment types where hydraulic conditions transform abruptly (Hakånson and Jansson, 1983). As discussed in the stratigraphic data, core LPCA would have been located on the periphery of the fluvio-lacustrine environment represented by the Group 1 cores, where coarse material was present, but not dominant. Therefore, core LPCA would not always be influenced by the fluvio-lacustrine environment close by. Rather core LPCA would be affected by the dominant processes of the time, either the deltaic sedimentation of the Group 1 cores or the littoral lake zone. Throughout CAPS I, the levels of silt indicate that the site was part of a permanent lake body between ***c.*** AD 500 - 1290. This is supported by the dominance of silt material, which is representative of this environment following Digerfeldt (1986) and Campbell (1998). The particle size record of core LPCB (beginning ***c.*** AD 1100) indicates large fluvial inputs and delta development. Core LPCA was located only 500 m from the suggested AD 1913 River Ljuča course (Figure 6.3) and would most likely be affected by the high energy environment affecting core LPCB. Therefore, both littoral lake and fluvio-lacustrine sedimentation is thought to have influenced sediment deposition at core LPCA at this time. Between ***c.*** AD 1290 - 1520 the slight rise in silt content is once more indicative of a permanent water body, but the lake level may have risen at this time as a result of an increase in discharge from the River Ljuča. This is in agreement with the coarse particle size record from core LPCB and the grit and coarse sand identified in B2. In contrast to Digerfeldt's (1986) model of lake sedimentation the increases in coarse grained material during this period are not interpreted as outward shore displacement i.e. falling lake level. Instead, periods of coarse sedimentation registered in the cores LPCA and LPCB particle size records are suggested to highlight an expansion in the influence of the River Ljuča delta. These are dated to ***c.*** AD 810 - 860 and ***c.*** AD 1280 – 1320. The pollen and LOI records corroborate this suggestion, as the pollen assemblage is varied with 44 – 47 herb types identified and few littoral macrophytes or organic sediments between ***c.***AD 500 - 1520 (LPCA I to III) which would be expected if the coarse material represented periods of outward shore displacement (Digerfeldt, 1986).

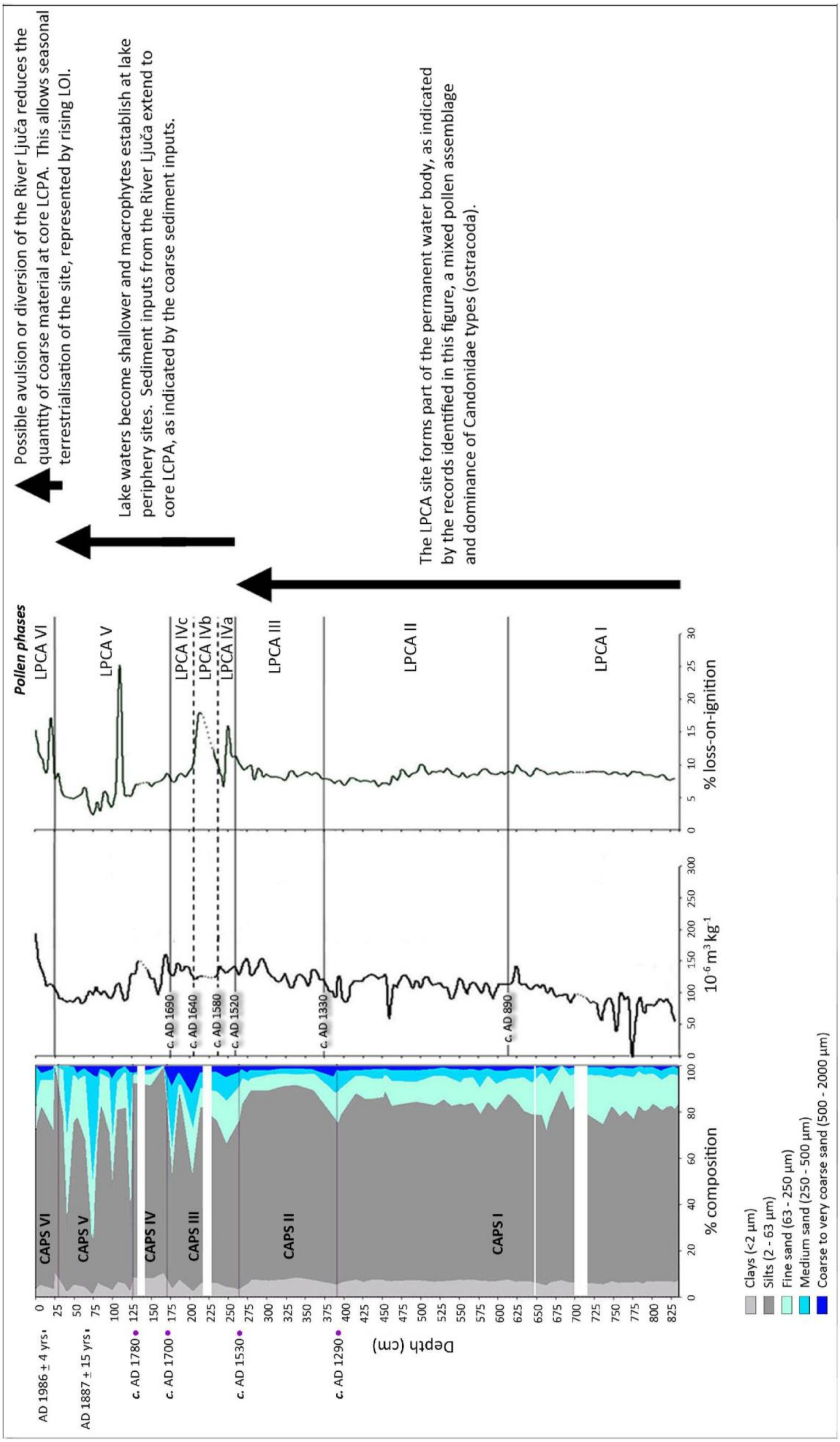


Figure 6.6: Comparison of the core LPCA particle size, magnetic susceptibility and LOI data sets.

Throughout CAPS II silt remains dominant, increasing slightly suggesting that fluvio-lacustrine influences may not have reached the site. This contrasts with the LPCB data, which indicates river activity and delta development were as strong as ever. As lakes act as a site for river input entrapment (Hakånson and Jansson, 1983), rising precipitation would be likely to increase the lake level rendering the LPCA site one with a higher water level, decreasing the volume of coarse sediment accumulation, but maintaining the supply of coarse material to core LPCB. Overall, the particle size record of CAPS II is thought to represent a time of increased lake level, likely to be the result of climatic deterioration.

Depth (cm)	Approximate date (AD)	PS Phase
50 – 35	1915 – 1940	**CAPS V**
81 – 64	1860 – 1890	**CAPS V**
107 – 90	1810 – 1840	**CAPS V**
128 – 120	1770 – 1780	**CAPS IV**
183 – 174	1670 – 1690	**CAPS III**
217 – 190	1610 – 1660	**CAPS III**
265 – 230	1520 – 1590	**CAPS II/ CAPS III**

Table 6.1: List of coarse (>63µm) grained events identified within the LPCA particle size record.

CAPS III to V, present seven shifts (see Table 6.1) towards a sediment composition dominated by coarse grained fractions between ***c.*** AD 1520 and 1940. Using the previous interpretation, these shifts could easily be explained by rising water levels and fluvial inputs. However, the core LPCA palynological data indicates that around AD 1640 an expansion of macrophytes occurred within the catchment. Beyond ***c.*** AD 1520 whereby the site is becoming shallower allowing macrophytes to establish close to the lake periphery and coarse shifts indicate sediment inputs from the River Ljuča. This change can be seen when the particle size data of ***c.*** AD 1910 are compared to the Lake Plav extent drawn by Cvijić in AD 1913. A shift towards a silt dominated system occurs around this time, when the LPCA site would have been inundated (see Figure 6.6). By the final particle size phase (CAPS VI), silt reaches its second highest level throughout the core at ***c.*** AD 1960 before returning to levels more comparable to the earlier CAPS I. Climatic influences are then thought to have caused lake levels to drop exposing the floodplain seen today.

6.1.3 Particle size data for core LPCB

Across the Mediterranean and Europe the LIA not only caused lake levels to increase (Magny, 2004; Holzhauser *et al.*, 2005), but the climatic deterioration created an abundance of clastic material which could be transported by river systems (Revel-Rolland *et al.*, 2005). The model of deltaic sedimentation applied to the core LPCB stratigraphy was due to its relatively central lake location (Figure 6.1, 6.4), but dominance of coarse grained material with around 60% sediments >250 µm. Such coarse material has been associated with deltaic sediments (Sturm, 1975; Table 6.2).

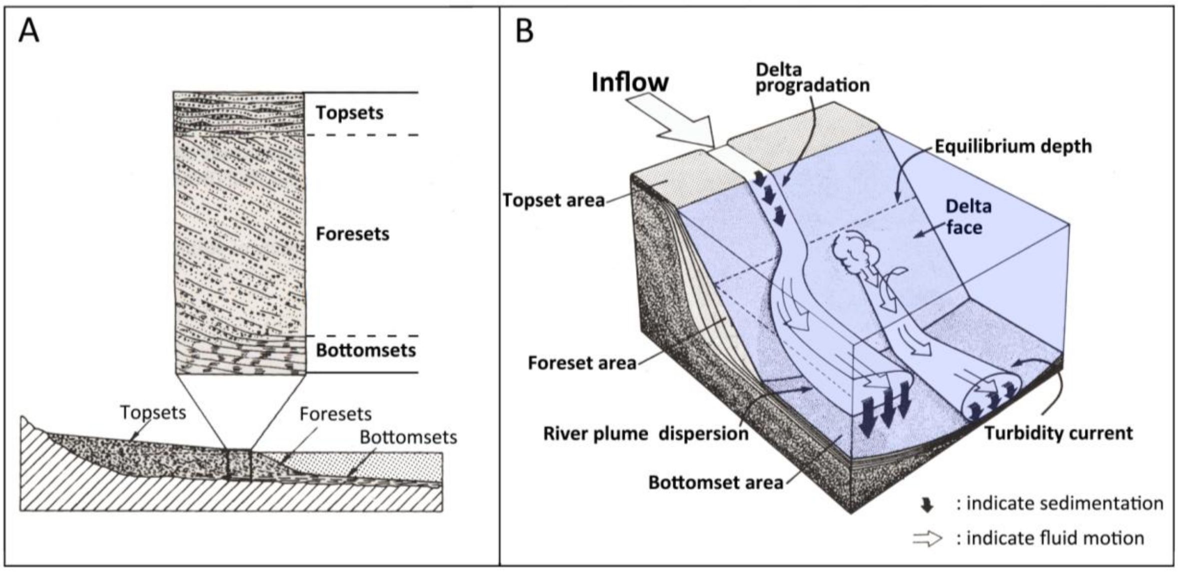


Figure 6.7: (A) Idealised profile and section through a delta. (B) Sedimentation mechanisms associated with an inflow entering a lake (adapted from Friedman and Sanders, 1978; Pharo and Carmack, 1979 both cited in Håkanson and Jansson, 1983).

Figure 6.7 provides a schematic illustration of deltaic progradation and identifies the sediment fractions expected to dominate particular areas during sediment deposition. This deposition is concentrated where the jet flow begins, with sands characterising river-mouth areas, whilst prodelta sediments consist of finer sediments (Moore, 1966; described at Lake Brienz in Table 6.2).

	Delta area	**Central basin plain**	**Lateral slope**
Energy level	High	Low, occasionally high	Low
Environment	Erosional = depositional	Erosional < depositional	Depositional
Channels	Common	Not distinct	Absent
Sedimentation rate (1900-1983)	40 – 50 mm yr^{-1}	5.5 – 6.8 mm yr^{-1}	2.5 – 3.3 mm yr^{-1}
Dominant sediment type	Sand	Mix sand, silt, clay	Mix silt and clay

Table 6.2: Summary of the various depositional environments associated with Lake Brienz (Sturm, 1975), adapted from Hakånson and Jansson (1983).

Using the core LPCB age depth model, periods of greater fluvial inputs are suggested to have occurred between; AD 1180 - 1530, AD 1575 - 1700, AD 1710 - 1740, AD 1760 - 1820, AD 1830 - 1860 and ***c.*** AD 1870 (see Figure 6.4). Following research by Hutchinson (1957) and, Hakånson and Jansson (1983) the coarse sands sediments of core LPCB are associated with the river mouth – lake transition i.e. delta area (Table 6.2). This follows the interpretation of the stratigraphic records for the Group 1 cores discussed earlier. The continual increase and decline in coarse grained material throughout the record, points to periods of greater fluvial activity. This currently occurs during spring melt and periods of torrential precipitation (Autumn, Llasat and Rodriguez, 1997), which are both associated with greater fluvial sediment inputs (Hakånson and Jansson, 1983; Leeman and Niessen, 1994). At the beginning of LPCB couplets were identified (***c.*** AD 1110 –

1120, see Figure 5.19, Chapter 5), which are indicative of a flooding event. However, the relatively low organic content (<7%) suggests sub-aqueous deposition and thus a flood extending into the lake; as floodplain sediments are organically rich with silt-sized particles (Koutsios *et al.*, 2010) and overbank flood deposits are sandy in nature (e.g. Williams and Roberts, 1990; Faust *et al.*, 2004; Collins *et al.*, 2005). Throughout LPCB I there is a progressive decline in the volume of silt and clay material which may represent increased precipitation associated with climatic deterioration. The record is punctuated by discrete coarse grained units produced by larger flows (see Figure 6.3). By the end of LPCB I the velocity and turbulence of the River Ljuča waters increased as >80% of sediment were medium and coarse to very coarse sand, ***c.*** AD 1650. The fluctuating grain size during LPCB I and subsequent LPCB II and III phases would indicate climatic variability, which affects the flood regime of rivers (Noren *et al.*, 2002; Parris *et al.*, 2010). This is consistent with the idea that the LIA was not a period of homogeneous climatic deterioration (e.g. Briffa *et al.*, 1988; Büntgen *et al.*, 2005; Johnston *et al.*, 2010).

The second phase (***c.*** AD 1660 - 1760) registers greater quantities of silt and clay material (minimum ~23%), the result of a lower energy environment. A relative plateau in silts and clays occurs in the first half of LPCB II, but these fractions increase towards the phase end. This suggests that the competence of the River Ljuča may have reduced allowing greater quantities of finer fractions to accumulate at the site. The record remains indicative of deltaic sediments, with coarse fractions still dominating the sediment composition for the majority of LPCB II. At the end of this phase sandy material reaches peak values comparable to those of LPCB I.

During the penultimate phase (***c.*** AD 1760 - 1880), larger floods return for a time. However, a step-wise fall in the quantity of sands, interspersed with two peaks in silts and clays (***c.*** AD 1800 - 1835; 227 – 175 cm), suggests a reduction in fluvial inputs. LPCB III is seen as the transitional phase, where the influence and course of the River Ljuča may be changing to that of the present day (Figures 6.1, 6.3). During the last two decades of LPCB III (135 – 113 cm), silts and clays twice rise above 50% indicating a lower energy environment before silts and clays remain above 80% for the majority of LPCB IV.

River progradation is evident from comparison of the current and past extents of Lake Plav (see Figure 6.1). Prior to the end of the 1800s, any large inputs of coarse grained material seem to have been followed by a short unit of finer material followed by another coarse grained input. It was not until the end of the 1800s that the final deluge of coarse grained sediments (***c.*** AD 1830 - 1870; 175 – 130 cm) may have caused an avulsion, forcing the River Ljuča around Lake Plav to its present day course. As a result the supply of coarse grained material ceased and silts and clays became the dominant sediment type into the final phase. The site is not thought to have become terrestrial at this time, due to the still relatively low organic content (Figure 6.4).

Finally, LPCB IV contrasts with the entire particle size record, with no more than 20% sand until the uppermost 10 cm, when this increases to nearly 40%. The composition of these sands is different to previous phases, as they are dominated by fine sands rather than medium and coarse to very coarse sands. The higher percentage of clay sized particles suggests the site is far less turbulent than previous times and thus the fine sand particles introduced are most likely the result of overbank floods rather than deltaic deposition as in previous phases (an observation made at other sites: Williams and Roberts, 1990; Faust *et al.*, 2004; Collins *et al.*, 2005). The sediment composition in the surface layers of core LPCB were indicative of a vegetated floodplain environment, as woody fragments and rootlets were present in the top 15 cm and LOI increases throughout.

6.1.4 Magnetics and loss-on ignition data for cores LPCA and LPCB

Overall, the magnetic susceptibility record of LPCA remained relatively stable with the mean values varying by just 43.5 $x10^{-9}$ m^3 kg^{-1} (see Table 5.8, Chapter 5), with indication of possible magnetic enhancement in the latter stages of the record.

Magnetic susceptibility and LOI were particularly suppressed during phase LPCA I (***c.*** AD 500 - 890) which may be indicative of drier conditions and thus, lower levels of eroded material entering the lake at this time. However, these conditions would also infer a lake level drop consistent with MWP reconstructions throughout the Mediterranean (Hughes and Diaz, 1994; Bradley *et al.*, 2003), but this does not fit in with the lake development model, which suggest the lake was a permanent lake body. This pattern is continued until ***c.*** AD 1520 (the end of LPCA III) and is broadly consistent with the particle size record which registers little change until ***c.*** AD 1520 (see Table 6.1). Therefore, the LOI and magnetic susceptibility records are thought to suggest environmental stability at the site until ***c.*** AD 1520.

During LPCA IVa and LPCA IVb, LOI and 'total sands' rise suggesting that sediment inputs increased (Figure 6.6; Digerfeldt, 1986). The poor correlation of particle size and LOI records to the magnetic susceptibility record suggests that, at the LPCA site, different sources and variations were affecting these variables throughout the record (Table 6.3). Through LPCA IVc (***c.*** AD 1580 - 1640) a possible change in source area is indicated by a rise in magnetic susceptibility, but falling LOI (Dearing *et al.*, 1981; Digerfeldt, 1986; O'Sullivan, 1994). This is consistent with the particle size record of the time which records a spike in coarse grained material indicating a source less dominated by limestone. However, without a record of sediment provenance, this cannot be confirmed.

Magnetic susceptibility continues to rise by ***c.*** AD 1690 (LPCA V), but finer clay and silt particles dominate the particle size record, suggesting a different dominant source. By ***c.*** AD 1820 (105 cm) a spike in LOI coincides with a brief period of coarser material suggesting the deposition of allochthonous material from a more organic rich source. Although the reason for these changes may also be coincidence; as inferred by the low correlation between LOI and particle size (see Table 6.3). Through the latter half of LPCA V magnetic susceptibility falls, but 'total sands' rise suggesting that any flood inputs were dominated by diamagnetic material e.g. pure limestone or quartz (Hrouda *et al.*, 2009). Research by Parris *et al.* (2010) supports this suggestion, as they found magnetic susceptibility and LOI may not always be as sensitive to palaeostorm/increased sediment transport events. This may account for the lack of compatibility between the three datasets throughout the record.

The LPCA stratigraphy suggests that during the 20th Century the lake level began to fall, as silts dominated, indicative of floodplain sedimentation (Koutsios *et al.*, 2010). The final LOI spike at ***c.*** AD 1960 indicates terrestrialisation of the site, supported by the intertwined rootlets recorded within the stratigraphy at this point. Overbank flooding still affected the area, as sands remain present (Figure 6.6; e.g. Williams and Roberts, 1990; Faust *et al.*, 2004; Collins *et al.*, 2005). Enhanced magnetic susceptibility at the end of the record, may be the result of rainfall events bringing magnetic material to Lake Plav, as seen at Shkodra Lake from AD 1964 (Welden *et al.,* 2008). As the record did not seem to conclusively highlight periods of possible erosion it is thought to be the result of anthropogenic pollution in the area.

In conclusion, the LOI records seem to provide the greatest information and support the suggestion that site LPCA remained part of the lake until the 20th century, when seasonal terrestrialisation of the area occurred. The magnetic susceptibility record is less conclusive with little change seen throughout the record and periods of apparent increased fluvial inputs not identified within the record.

Variables	LPCA, correlation coefficient	LPCB, correlation coefficient
Mag. susc. & LOI	0.30	0.45
Mag. susc. & particle size	-0.10	-0.35
LOI & particle size	-0.29	-0.755

Table 6.3: Correlation coefficients for various variables from core LPCA and LPCB.

The core LPCB record provides a similarly stable low frequency (LF) magnetic susceptibility and organic content throughout. However, magnetic susceptibility values are higher within core LPCB than that of core LPCA, which indicates greater concentration of magnetic materials. This may be the result of the sites greater concentration of flood deposits at the site, which may contain more magnetic material.

LOI remains relatively low until the end of LPCB III, suggesting that the biological productivity within the lake was low. Elevated LOI is thought to provide a record of terrestrialisation at the site. This is supported by the strong negative correlation between particle size and LOI (Table 6.3). Suppressed LF magnetic susceptibility readings suggest that any allochthonous inputs to the site are relatively consistent until ***c.*** AD 1440 – 1500, when two small peaks in activity occur, coinciding with rising levels of coarse grained material (Figure 6.3). Between ***c.*** AD 1500 - 1870, both the LF magnetic susceptibility records and LOI records register relatively stable values. A slight increase in LOI between ***c.*** AD 1730 - 1733 is coeval to a minor reduction of LF magnetic susceptibility and reduced levels of sand, suggesting flood activity may have diminished slightly. By ***c.*** AD 1870 rising LOI is indicative of site stabilisation with reduced fluvial inputs, corroborated by the particle size record (Figure 6.3). The particle size record indicates lake infilling before seasonal terrestrialisation with overbank flooding, represented by greater silt content, becomes dominant.

Finally, the current, primarily terrestrial, state of the LPCB site is suggested to have taken hold by ***c.*** AD 1920 when LOI begins to steadily rise (from 6 – 16%). Again the LF magnetic susceptibility record does not seem to provide much indication of sediment deposition or depositional environment as it remains in a suppressed but variable state at around 130 $\times 10^{-9}\ m^3kg^{-1}$ (possibly variable due to seasonal inputs of allochthonous material and suppressed due to the dominance of diamagnetic materials). Particle size data at this time suggests that between ***c.*** AD 1880 the River Ljuča may have been diverted around the lake, allowing sediment to build up and terrestrialisation of the site to occur. Elevated magnetic susceptibility values during the last ~20 years may be the result of material transported from different sources entering the site, but most likely anthropogenic activity as in core LPCA.

6.1.5 Interpreting the ostracoda record

Ostracoda are present in core LPCA below 200 cm. As the individuals found were intermittently present, the record suggests that either (1) the lake could not sustain ostracoda communities, or (2) ostracoda populations that did exist were affected by deltaic influences – which crushed the fragile valves or re-deposited them in a more central lake location, away from their original habitat. The low number of individuals throughout the core was unexpected, as limestone-dominated catchments are ideal for ostracoda preservation. Belmecherie *et al.* (2010) suggested sediments from Lake Ohrid in Albania-Macedonia without ostracoda represented Quaternary glacial periods when the site lacked active groundwater. This explanation seems inappropriate for the Lake Plav data, as the LIA was not characterised by the extensive glaciations seen during the Quaternary. Furthermore, recent climatic amelioration has not resulted in recovery of the ostracoda population.

The influence of the River Ljuča upon core LPCA is thought to have resulted in the destruction and/or flushing away of individuals. The variety of species identified is indicative of a permanent water body over the 1500 year period core LPCA represents, whilst identification of *Potamocypris* and *Cypridopsis* (types associated with flowing water) indicates that some ostracoda were transported to the lake from fluvial environments/habitats, consistent with a fluvio-lacustrine setting from ***c.*** AD 1210 onwards. These original locations are unlikely to have been a long distance away from the core LPCA site as many were picked as complete carapaces. *Ilyocypris*, which can tolerate subterranean springs and groundwater, may also have been transported to the lake and again short distances are implied from the good preservation of the individuals (Meisch, 2000).

In other Mediterranean region lakes between 5 and 11 different genera have been identified in any one lake (e.g. Juliá *et al.*, 1998; Frogley *et al.*, 2001; Sohar and Meidla, 2010). At Lake Plav 8 different genera were identified, but genera such as: *Darwinula*, *Limnocythere*, *Cyrideis*, *Paralimnocythere, Leucocythere*, *Physocypria, Herpetocypris, Psychrodromus* and *Notodromas* were absent (***cf.*** Mezquita *et al.*, 1996; Belis *et al.*, 1999, 2008; Juliá *et al.*, 1998; Külköylüoğlu 2003; Külköylüoğlu and Dügel, 2004; Riera *et al.*, 2004; Kiss, 2007; Reed *et al.*, 2008; Frogley *et al.*, 2001; Sohar and Meidla, 2010). The majority of these groups belong to the Darwinuloidea and Cytheroidea super-families, neither of which was identified at Lake Plav. The only super-family identified at Lake Plav was Cypridoidea (Meisch, 2000), which includes *Notodromas*, *Physocypria*, *Herpetocypris* and *Psychrodromus*. Within the reference material, Lake La Cuz and the Estranya lakes situated in the Pre-Pyrenees were most comparable to Lake Plav setting, as they were situated within karst topography and at altitudes between 1000 – 670 m a.s.l., but both maintained Darwinuloidea populations (Juliá *et al.*, 1998). Other lakes in karstic situations with populations of Darwinuloidea and Cytheroidea included Lake Pamvotis in North-Western Greece, Lake Ohrid in Albania/Macedonia and Lake Fehér in Hungary (Frogley *et al.*, 2001; Kiss 2007; Belmecherie *et al.*, 2010). Only Laguna Zoñar (Zoñar Lake), Southern Spain (Valero-Garcés *et al.*, 2006; see Chapter 2), had an assemblage similar to Lake Plav. It lacked Darwinuloidea and Cytheroidea types and was dominated by Candonidae types, with *Potamocypris* indicating shallow and flowing water. This may suggest that Darwinuloidea and Cytheroidea types tend not to populate areas dominated by fluvial inputs. Furthermore, Valero-Garcés *et al.* (2006) recovered low numbers of ostracoda from Laguna Zoñar and suggested that low oxygen content and restricted water circulation inhibited the survival of benthic types. Such conditions were indicated by lamination of sediments and authigenic calcite. However, the LPCA sediments lacked laminations or authigenic calcite and endemics such as *Leptocythere* (Frogley *et al.*, 2001). Therefore, the argument that fluvial activity and influence of the distal delta edge at the site impeded or removed benthic types is preferred. The taphonomic data also suggest that there has been some limited reworking of ostracoda, as some appear orangey red in colour; an appearance associated with either iron staining and thus soil formation, or a change in chemical composition of the sediments (Dr David Horne *pers. comm.*). As there is no evidence for soil development at these points, the chemical composition of the surrounding sediments may have caused the valve staining and this could be assessed using analysis of the redox chemistry (Pruysers *et al.*, 1991).

Overall, the ostracoda record from LPCA were indicative of a permanent water body which was affected by fluvial inputs throughout its history reducing ostracoda carapace concentrations before these inputs intensified around ***c.*** AD 1640. This led to an absence of these crustaceans as fluvial inputs and more recently overbank flooding and seasonal terrestrialisation of the site occurred.

6.1.6 Assessing the Lake Plav palynological record

The plant taxa recorded in the LPCA palynological profile have been divided into 6 different groups (Table 6.4) with the aim of allowing climatic and anthropogenic variations to be identified. Thermophilous types used in previous studies have included *Quercus* (deciduous and evergreen), *Fraxinus*, *Tilia*, *Ulmus*, *Corylus*, *Betula*, *Alnus*, *Hedera* and *Ilex* (e.g. Wright, 1967; Huntley, 1990; Bergmeier and Dimopoulos, 2001; Finsinger *et al.*, 2006; Belis *et al.*, 2008). Huntley (1993) discussed the comparatively tolerant taxon *Corylus* which, more so than other thermophilous taxa, can withstand cooler winter temperatures and responds to changing climate conditions quickly as it will flower as soon as possible. Although *Corylus* was found sporadically up to 1800 m a.s.l. within the catchment, it was felt that following Huntley (1993) it was a useful thermophilous indicator. *Quercus*, *Betula* and *Alnus* were all found to be scattered throughout the catchment and although these types have been used in studies from Swedish to Southern Alpine sites (Kullman, 1998; Finsinger *et al.*, 2006), it was felt that due to their high pollen productivity these types may disguise more subtle changes within the other arboreal taxa and were therefore, excluded from the thermophilous group. During the vegetation survey *Tilia* and *Ulmus* were not identified, but scattered specimens of *Ostrya* and *Carpinus* were found up to ~1200 m a.s.l. and these taxa were included due to similarities with *Corylus*. Finally, *Fraxinus*, *Acer, Hedera* and *Sorbus* were added to this group because they were scattered throughout the lower catchment below 1200 m a.s.l. and downstream at Murino (837± 6 m a.s.l.; Figure 6.8). This distribution suggested an intolerance of the current conditions above 1200 m a.s.l.

Using the vegetation survey data, a further group was created; ‘Alpine meadow/scrub’. Types such as *Saxifraga* are common in the Balkans and prefer upland limestone bedrock, crevices and screes (Polunin, 1969), with *Saxifraga luteoviridis* growing above 1500 m a.s.l. (Polunin, 1980). Cruciferae (exc. *Brassica*) were recorded in the upper XIV zone and some species are known to prefer rocky mountainous areas, whilst *Artemisia* has been described as a 'high-elevation herb' by Jimenez-Moreno *et al.* (2010). *Juniperus, Rhamnacea* and Ericaceae were present above 1800 m a.s.l., with the former dominating shrub cover, and were thus included within the 'Alpine meadow/scrub' group (Figure 6.9); which aims to record any expansion of the vegetation currently dominating zones XIII and XIV (> 1800 m a.s.l.). All of the types included within this group are more tolerant of cooler conditions, which may have led to their expansion during climatic events such as the LIA. In conjunction with this group *Picea* and *Abies* are used to determine any expansion of high altitude vegetation, as these types are typical of high altitudes (Körner and Paulsen, 2004; Jimenez-Moreno *et al.*, 2010; see Figure 6.9), and would have a competitive advantage at the tree line over deciduous taxa.

Figure 6.8: Arboreal types identified in and around the Lake Plav catchment, (A) *Acer* and (B) *Prunus.*

The group 'Ruderals' was introduced, which Mazier *et al.* (2009) described as species that may be representative of pastoral or arable farming. The group includes: Chenopodiaceae, *Urtica*, *Rumex,* Compositae and *Filipendula* which have been cited by various authors as indicators of disturbed ground (e.g. Punning *et al.*, 2004; Joly and

Visset, 2009; Mazier *et al.*, 2009). Species of Chenopodiaceae have been associated with the Mediterranean Pleistocene steppe environment (similar to *Artemisia*; Tzedakis and Bennett, 1995; Frogley *et al.*, 1999; Allen, 2003; Tzedakis *et al.*, 2004b), but species such as *Chenopodium alba* have been associated with contemporary annual crops preferring arid summer environments (Allen, 2001). During the vegetation survey, *Urtica,* Compositae and *Rumex* were found particularly around roads and the edges of fields. This group represents ground disturbance as a result of farming without differentiating between 'Arable' or 'Pastoral' types.

Group	Genera included
Thermophilous	*Fraxinus*; *Ostrya/Carpinus*; *Tilia*; *Ulmus*; *Corylus*; *Acer*; *Sorbus*; *Hedera*
Alpine meadow/scrub	*Juniperus*; Ericaceae (*Erica* type); *Saxifraga*; Cruciferae (exc. Brassica); *Artemisia*; *Rhamnacea*
Ruderals	Chenopodiaceae; *Urtica*; *Rumex*; *Compositae*; *Filipendula*
Arable	*Juglans*; *Prunus*; *Secale cereale*; *Hordeum*-type; *Plantago major*; *Brassica*
Pastoral	*Plantago lanceolata*; Ranunculaceae; Lactuceae; *Campanula*; Valerianceae; *Melampyrum*
Lake periphery/wetland	*Typha latifolia*; Cyperaceae; *Equisetum*

Table 6.4: Vegetation groups used for discussion and the pollen types included.

Prunus and *Hordeum* – type (Barley) were added to the arable group and Valerianceae and *Melampyrum* to the pastoral indicators (see Polunin, 1980; Figures 6.8, 6.9). Finally, the lake periphery/wetland group aimed to define pollen types associated with a lake edge environment (*Typha latifolia*) and waterlogged areas i.e. *Equisetum* and Cyperaceae, which currently colonise peripheral areas of the lake (Figure 6.8).

Phase	Average pollen concentration excluding non-polleniferous (g cm^{-3}) samples
LPCA VI	8,233
LPCA V	6,724
LPCA IVc	37,508
LCPA IVb	32,279
LPCA IVa	11,409
LPCA III	13,331
LPCA II	23,038
LPCA I	26,601

Table 6.5: Average pollen concentrations in core LPCA in grains per cm^3.

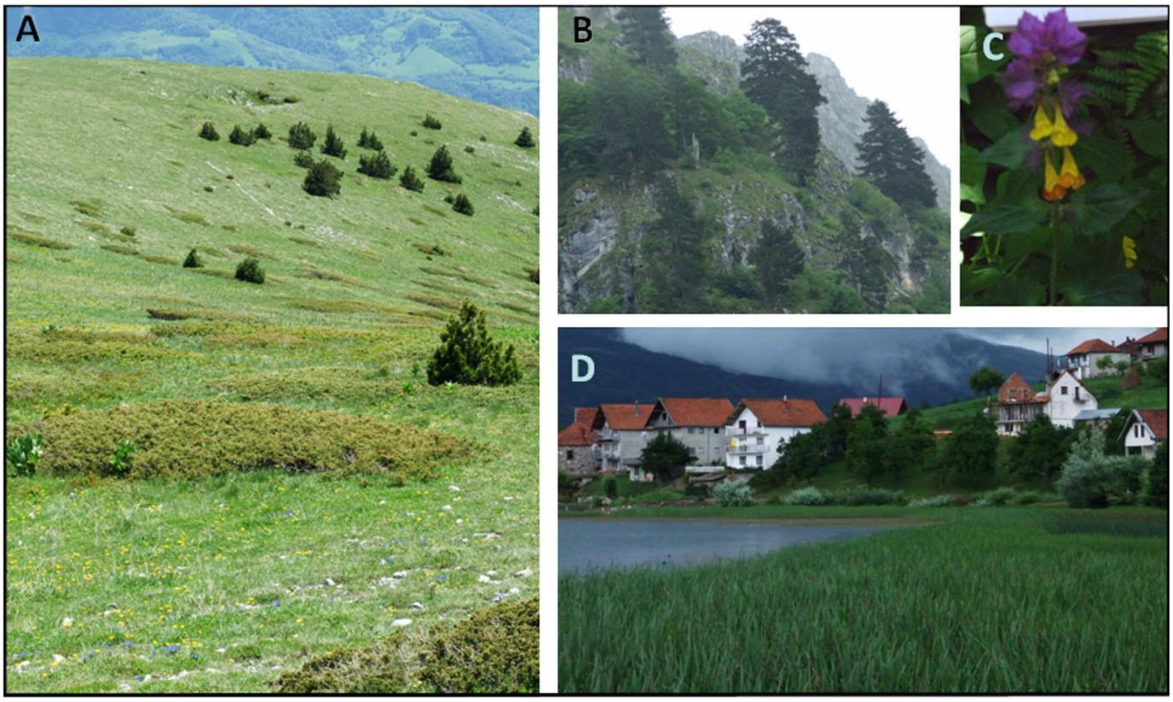


Figure 6.9: Selected vegetation seen around the Lake Plav catchment with reference to pollen types described in Table 6.1. (A) land above 1900 m a.s.l., Visitor Mountain, *Juniperus communis* subsp. *nana* (following description by Polunin, 1969) and small *Pinus* visible. (B) *Abies* on the steep Grbaja valley slopes. (C) *Melampyrum* in a pastoral field and (D) Lake edge vegetation dominated by Cyperaceae (not in flower).

6.1.7 Interpreting the Lake Plav pollen record

Changes in the pollen record represent variations in vegetation across the catchment and are never exclusively representative of the changing flora at the coring site. Variations in the record may be skewed towards vegetation local to the site, but a continuous catchment signal remains due to the effects of long distance 'pollen rain' and sediment transport. The record here follows the division of pollen phases to estimate past temperatures, landscape change, land use and lake extent variations.

During the LPCA pollen profile (condensed version Figure 6.10) the two most significant changes are the steep decline in *Alnus* ***c.*** AD 1330, and later the increase in lake periphery/wetland types ***c.*** AD 1580 (LPCA IVa/b), which will be discussed chronologically within the associated phase descriptions. Poor pollen concentrations characterised the Lake Plav record, with shifts from 58,439 g cm^{-3} (680 cm) to only 848 g cm^{-3} (240 cm) and even a number of non-polleniferous samples particularly during phase LPCA V (see Figures 6.11; 5.24, Chapter 5). Concentrations below 1,000 g cm^{-3} are very low; with comparable studies registering minimum concentrations of 3,000 - 10,000 g cm^{-3} (e.g. Watson, 1996; Wick *et al.*, 2003; Schwab *et al.*, 2004). Low pollen concentrations may result from high sedimentation rate reducing the concentration of pollen grains (Mudie and Byrne, 1980), or by sustained surface exposure leading to degradation and oxidation of pollen grains (Cushing, 1967). In contrast, higher pollen concentrations may be the result of reduced sedimentation rate allowing longer periods of pollen accumulation before being overlain by sediments (Watson, 1996; Wick *et al.*, 2003), or erosion from pollen-rich sediments inputting greater quantities of pollen to a site (Davis, 1999; Zhao and Sun, 2010).

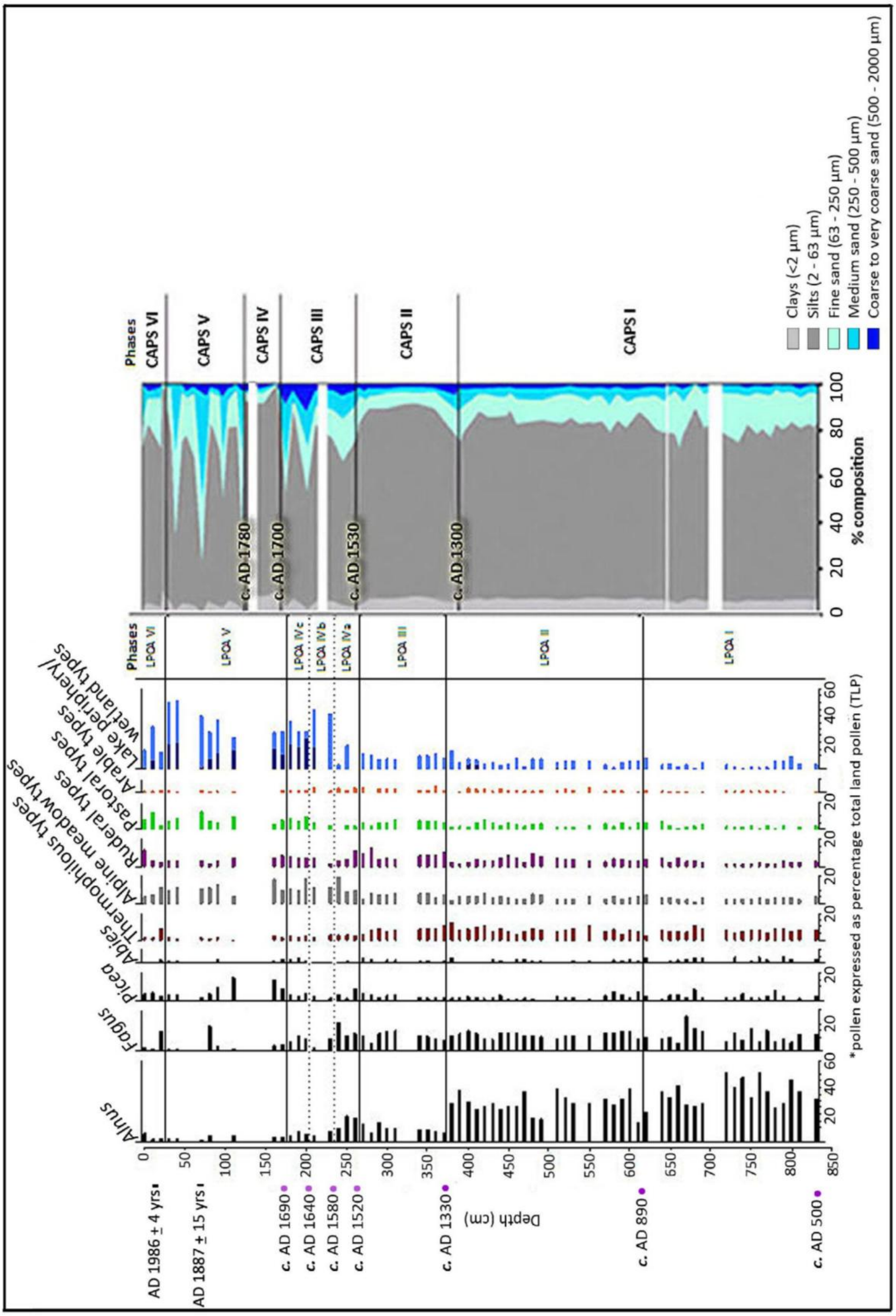


Figure 6.10: Pollen groups and particle size record from core LPCA.

The taxonomically diverse pollen assemblage of phases LPCA I and II are indicative of an ‘open water’ site between ***c.*** AD 500 – 1330, as researchers have reported greater pollen diversity at permanently submerged lake body sites (e.g. Lowe, 1992; Mohammed *et al.,* 1995; Wick *et al.,* 2003). Open water indicators *Nuphar* (water-lily) and *Myriophyllum* (Polunin, 1969; 1980) along with a minerogenic stratigraphy confirm that Lake Plav extended beyond the LPCA site between ***c.*** AD 500 – 1330. Evidence for local land disturbance occurs throughout ***c.*** AD 500 – 1330 with the presence of ruderal types (i.e. Chenopodiaceae, *Filipendula*, *Urtica*, *Rumex*), but there is no evidence of further land clearance as arboreal types remain relatively stable. At this time the catchment appears to be dominated by *Alnus*, *Fagus, Pinus* and *Salix.* As *Alnus* and *Salix* prefer more waterlogged river edge conditions, their dominance may be a product of their proximity to waters feeding Lake Plav, and the edge of the lake itself. However, *Alnus* and *Pinus* are high pollen producers and can be over-represented within the pollen assemblage, compared to their actual vegetation abundance and thus, may not be as dominant as the record suggests. Warmer local conditions compared to later phases (LPCA IV to VI) are indicated by the comparatively higher contribution of thermophilous taxa to the %TLP, at around 10%. Evidence of anthropogenic activity is minor during LPCA I, with indicators remaining, on average, below 5% until ***c.*** AD 840 when they rise to over 10%. Ruderal types dominate the anthropogenic indicators and by ***c.*** AD 840 pastoral types rise to ~5% indicative of local land disturbance. Soon after, ***c.*** AD 890 - 1330, arable indicators rise consistently above 2%, suggesting the local population may be intensifying their farming practices particularly with the cultivation of *Hordeum* – type (Barley) and *Juglans* (Walnut). The increase in anthropogenic indicators may well be the result of areas being eroded and sediments from anthropogenically modified areas being inputted to the lake.

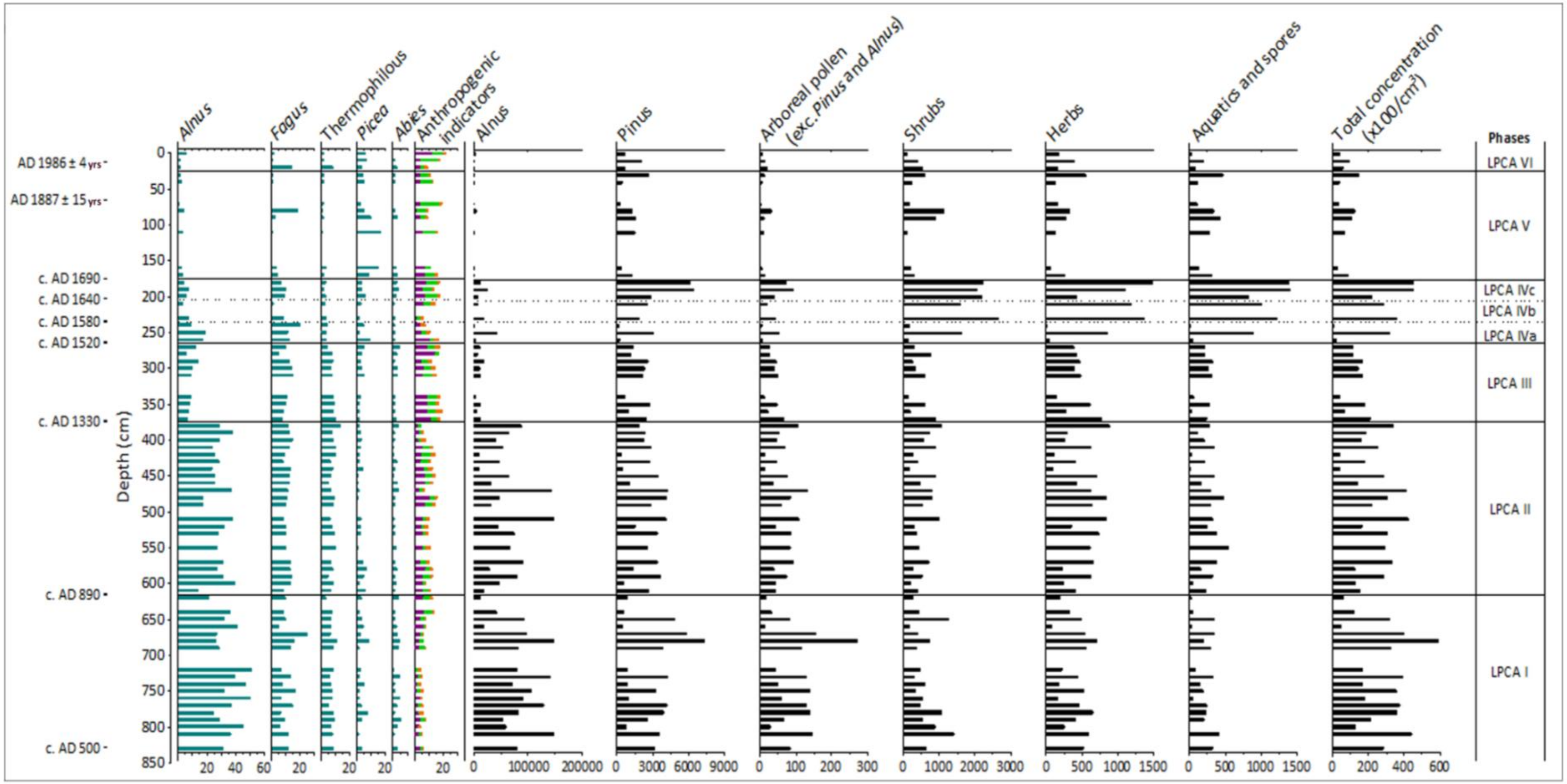


Figure 6.11: Total pollen concentration data for core LPCA (black histograms) compared to selected species presented as percentage TLP (Teal histograms). Anthropogenic indicators include, ruderal types (purple), pastoral types (green) and arable types (orange).

LPCA II (***c.*** AD 890 - 1330) continues in a similar fashion, but more stable *Fagus* and *Pinus* signals replace the reduced *Alnus* frequencies. *Salix* also declined slightly during this period, suggesting that local waterlogged habitats for *Alnus* and *Salix* were reduced, either as a result of climatic variation or anthropogenic impacts within the catchment. The latter cause is indicated within the pollen assemblage, as *Poaceae*, ruderal, pastoral and arable types all increase ***c.*** AD 890 – 1330. This suggests continued land clearance and anthropogenic activity, perhaps focused on the floodplain. Within the pastoral indicators, *Plantago lanceolata* dominates the assemblage indicating land reclamation for pastoral farming. Thermophilous types indicate that the climate remained relatively warm, which may account for the rising levels of pastoral and arable farming within this phase and perhaps also lower levels of *Picea* and *Abies*. However, it is difficult to decipher the dominant forces at this time i.e. did climate change force anthropogenic activity to expand - this period is thought to represent the beginning of LIA deterioration at the site (see section 6.5.3), or did anthropogenic activity increase erosion and thus input of pollen from anthropogenically modified areas? Considering the expansion of lake periphery/wetland indicators during LPCA II a changing climate seems the likely cause for the expansion in anthropogenic activity using the data available.

An abrupt *Alnus* decline characterises the beginning of LPCA III (***c.*** AD 1330 – 1520), alongside a rapid increase in anthropogenic activity indicators (Figure 6.10). A similar *Alnus* trend has been recorded in other regions when an expansion of rye (*Secale cereale*) cultivation occurred (Sarmaja-Korjonen, 2003; Saarse *et al.*, 2010) and in the LPCA profile *Secale cereale* also increases at this level. Anthropogenic wetland clearance appears to have been the primary cause of the rapid *Alnus* decline. Increased pressure upon the land may have been the result of either a growing population at the time or an effect of the LIA climatic deterioration. Many parts of Europe encountered diminished food supplies in the early 1300s e.g. 'The great Hunger' of AD 1315 - 1319 (Fagan, 2000) and this may have forced wetland clearance to expand the land available for cultivation, at this point historical documentation would enable these suggestions to be confirmed/rejected. Further evidence for the clearance of *Alnus* from floodplain waterlogged areas is the synchronous expansion of *Equisetum*, which would benefit from a more open but still wet habitat, comparable to its current growing conditions around Lake Plav. Phase LPCA III is thought to represent the beginning of LIA climatic deterioration in all but the pollen record as a slow waning of thermophilous types does not occur until ***c.*** AD 1460. Average total pollen concentrations fall by 43% during the third phase which coincides with elevated silt content but stable magnetic susceptibility and LOI levels. This is indicative of slow moving water with stable levels of allochthonous materials, which does not seem conducive to suppressed pollen concentrations. Three explanations may be considered; (1) surface exposure caused grain degradation (Cushing, 1967), (2) reduced pollen production as a result of the LIA climatic deterioration (Birks and Birks, 2001) or (3) higher rates of sediment accumulation in the later stages of the phase reduced the pollen concentration. The first suggestion is discounted, because there is no evidence for surface exposure in either the core stratigraphic or LOI. The remaining suggestions may both provide a valid explanation, as climatic deterioration is evident during the latter half of LPCA III and the LPCB particle size record indicates increased rates of sediment transport into the lake (see Figure 6.12).

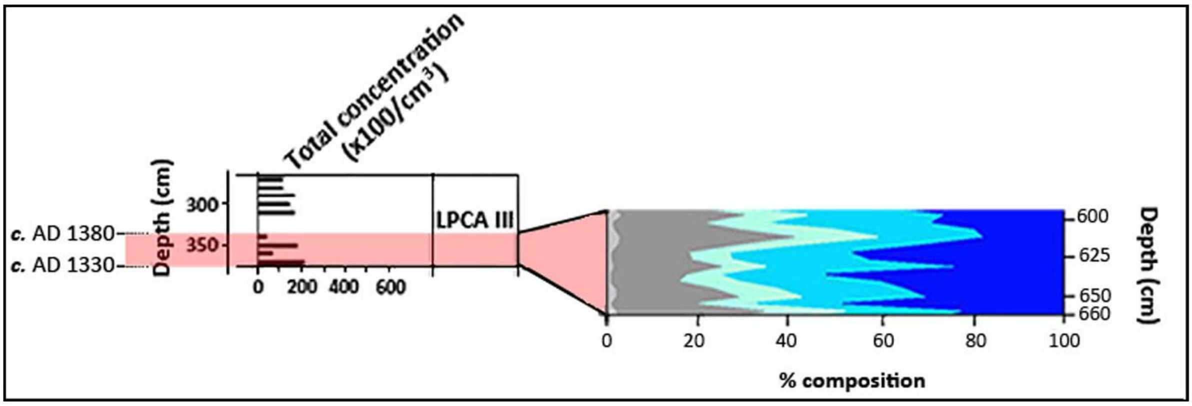


Figure 6.12: Period of reduced pollen concentrations within core LPCA alongside the same time period in particle record for core LPCB. Clays and silts (<63 µm) are shown in silver and grey and sands (>250 µm) in pale green to blue.

During phase LPCA IVa (***c.*** AD 1520 - 1580) alpine meadow/scrub types increase rapidly, indicating either expansion of vegetation zone XIV or greater erosion of such sites. Trees associated with these higher altitude sites, i.e. *Abies* and *Picea* (Figure 6.10) lag behind somewhat as they do not multiply until ***c.*** AD 1640 (LPCA IVc) and ***c.*** AD 1690 (LPCA V) respectively. This may be due to the longer time required for trees to establish. The second obvious change at ***c.*** AD 1550 is the sudden rise in lake periphery/wetland types suggesting an expansion in the mid LIA. The fluctuating total pollen concentrations at this point (between 31,466.9 g cm^{-3} – 847.9 g cm^{-3}) may offer some reason for the abrupt fluctuations seen in some taxa during phase LPCA IV e.g. *Picea*, *Equisetum* and *Typha latifolia*, which may result in data being skewed towards more local sources. *Equisetum* levels rise abruptly at 250 cm which may be more representative of lake peripheral areas than catchment-wide variations. *Equisetum* is tolerant of waterlogged land and rising sedimentation rates, due to its rhizome structure, thus fluctuations of this type may indicate increased sedimentation rates and possibly seasonal terrestrialisation of the site (Gill, 1973). However, with a lack of soil development, organic bands or plant macrofossils, terrestrialisation is not thought to have occurred at the LPCA site, but in the surrounding lake periphery. Pollen concentrations at this time fluctuate widely, with higher concentrations associated with coarser material implying the input of sediments from pollen rich sources. The diversity of arable types begins to wane within this zone, alongside that of pastoral types, which poses the question were taxa such as, *Juglans*, *Prunus*, *Secale cereale* and *Plantago major* unable to cope with the climatic deterioration associated with the LIA or did people chose to stop growing these types? Following the other Lake Plav datasets climatic deterioration was occurring between ***c.*** AD 1520 – 1580, but cultural decisions cannot be assessed from the datasets presented and it difficult to determine the cause for this change. Furthermore, during LPCA IVa rapid fluctuations occur in the Ericaceae and *Picea* records, indicative of upper catchment erosion as both were identified in contemporary vegetation zones, XIII and XIV, and *Picea* is a known European 'high altitude' species (Kienast and Kuhn, 1989; Körner and Paulsen, 2004). Coupled with the Ericaceae and *Picea* changes abrupt increases in *Alnus* and *Fagus* pollen between 260 cm and 240 cm, suggest that variations may be due to changing source areas associated also with the variable particle size record during LPCA IVa (Figures 6.10, 6.11). Changing source areas may be a product of anthropogenic activity possibly as a result of climatic deterioration, but this is difficult to define.

Phase LPCA IVb (***c.*** AD 1580 - 1640), continues to indicate an expansion of vegetation along the lake shoreline as *Equisetum* remains prevalent, and an explosion of *Typha latifolia* occurs at the end of the phase. *Typha latifolia* prefers silty marsh/wetland areas and flashy flood regimes (Kercher and Zedler, 2004). Its presence corresponds to low magnetic susceptibility readings, a rise in LOI and a coarsening of sediments thought to represent inundated shoreline lake vegetation where flood waters regularly reach the site. *Alnus* and *Fagus* pollen frequencies both decline once more between ***c.*** AD 1580 - 1690, suggesting that the dominant pollen source area has reverted to that dominating phase LPCA III. An expansion of wetland areas at Lake Plav is indicated by the now dominant lake periphery/wetland types, but a rise in *Poaceae* illustrates an open landscape where anthropogenic activity continues. The continued reduction of thermophilous taxa, are indicative of diminished temperatures. Throughout LPCA IVb pollen concentrations (~36,000 g cm^{-3}) and LOI (~15%) are comparatively high suggesting a possible reduction in sedimentation rates. The pollen assemblage is dominated by shrubs (primarily Ericaceae and *Juniperus*), herbs and aquatics. This indicates pollen derived from 'local' sources, but areas of the upper catchment where Ericaceae and *Juniperus* grew could have been undergoing erosion for comparatively high quantities of pollen to reach Lake Plav from such a remote source.

Indication of wetland expansion continues during LPCA IVc (***c.*** AD 1640 - 1690), but at this time *Equisetum* is not registered, and *Typha latifolia* dominates alongside Cyperaceae. The abrupt disappearance of *Equisetum* during LPCA IVc may suggest that the Lake Plav water level began to rise, as both Cyperaceae and *Typha latifolia* can withstand such submersions, but *Equisetum* is less tolerant (Gill, 1973; Lieffers 1983; ISSG, 2006). Comparably high and stable pollen concentrations are noted during this zone which would suggest ideal anaerobic conditions (waterlogged or submerged sediments) for the preservation of pollen grains, consistent with the notion of a large lake and widespread wetland at this time (see Figure 6.10; Table 6.4). Evidence for infilling of the site can be inferred from the coarse particle size. However, there is no evidence for terrestrialisation within the stratigraphic or pollen record, where presence of *Typha latifolia* indicated a saturated or flooded area for most of the growing season (Lieffers 1983; ISSG, 2006).

Therefore, the lake may have been becoming shallower as infilling occurred, whilst maintaining a large lake extent. Within the catchment ruderal and pastoral types dominated anthropogenic indicators ***c.*** AD 1640 - 1690 (LPCA IVc). This is the last time *Juglans* is identified in the palynological record and the declining diversity of arable types suggests that pastoral farming became more viable during periods of climate instability.

Phase LPCA V (***c.*** AD 1690 – 1960) coincides with the mid- to late LIA as seen across the Mediterranean. The lowest average pollen concentrations of the whole record, at 6,723 g cm^{-3} (Figure 6.11; Table 6.5) were recorded at this time. This is thought to have resulted from a combination of taxa failing to flower during the cooler climate, but primarily an increase in turbulence and sediment deposition at the site, as indicated in the particle size records of both LPCA and LPCB. Cooler climatic conditions are further indicated by the near-absence of thermophilous types and sustained rise of *Picea* and *Juniperus* (Figure 6.10), both of which can tolerate colder conditions, as seen in the contemporary vegetation zones XIII and XIV. Climatic cooling is then inferred throughout the 1700s and 1800s. Within the anthropogenic indicators ruderal types reach a low similar to that of LPCA I, but as the number of low count or non-polleniferous samples within this zone is high, the taxonomic diversity of the samples is likely to have been reduced, skewing results towards dominant taxa. Throughout LPCA V, lake periphery/wetland, herbs and shrub types dominate the assemblage suggesting a large lake extent with shallow areas ideal for wetland colonisation. Between ***c.*** AD 1690 – 1960 significant sediment inputs to the lake are indicated by PS records of LPCA and LPCB and this is characterised by spontaneous peaks in *Fagus* and *Salix* pollen resulting from variable quantities and sources of sediment input. This is corroborated by peaks being preceded and followed by troughs in frequency unlikely to represent vegetation loss and then recovery over two decades (the approximate sampling interval for the palynological record).

A reduction in tree cover throughout the catchment during phase LPCA V is indicated by a reduction in AP and also increasing levels of Lactuceae, which Collins *et al.* (2005) suggested to be indicative of a largely tree less environment. *Typha latifolia* maintains a strong presence during LPCA V until the end of the phase. At a depth of 70 cm a reduction in *Typha latifolia* coincides with the largest input of coarse grained material at LPCA, therefore, it may have been the result of increased sedimentation rate diluting the pollen assemblage, or input of sediments from a different area dominated by other pollen types. This fluctuation is also seen in the *Picea*, *Abies*, *Fagus*, *Salix*, Ericaceae and *Juniperus* records, suggesting that variable rates of sedimentation caused some taxa to be under-represented.

The brief zone LPCA VI represents the past 45 years, with the final sample being indicative of the present day vegetation at the site. A steep decline in the lake periphery/wetland group is primarily the result of *Typha latifolia* disappearing, and <5% *Equisetum*. Lake Infilling is a probable cause with a 42% reduction in the size of Lake Plav as a result of climate amelioration during the 20th Century, i.e. a temperature rise and reduction in precipitation. Thermophilous types recover, particularly *Corylus* and *Ostrya*/*Carpinus*, indicative of 20th Century warming, which is confirmed by the reduced percentages of *Picea* and *Juniperus* within this phase. Pollen concentrations become more stable at this time and the stratigraphic record indicates the presence of intertwined rootlets, indicative of a more terrestrial setting. The continued dominance of pastoral farming within the anthropogenic indicators suggests that the areas around Lake Plav have continued to be cleared for pastoral agriculture, but arable farming did not recover from late Holocene climatic deterioration. Arable types remain at low levels, as seen ***c.*** AD 1580 which suggests that arable cultivation has been maintained on a more 'domestic' scale. The dominance of farming, particularly pastoral in the immediate area surrounding the LPCA site may go some way to explain the low levels of *Fagus* seen in the upper record. Locally sourced pollen is likely to dominate the pollen profile compared to that from the wider catchment where the *Fagus* arboreal type dominates.

Overall, the palynological profile suggests that since ***c.*** AD 500 anthropogenic influences at Lake Plav have risen, but by ***c.*** AD 1520 arable farming was in decline with pastoral farming preferred – most likely the result of LIA climatic deterioration. The LIA deterioration is thought to have occurred between ***c.*** AD 1330 – 1520, using the LPCA data, with evidence for a temperature depression ***c.*** AD 1460 inferred from the decline of thermophilous types. The LIA is then characterised by rapid sedimentation at the site coupled with higher precipitation, creating a large but shallow lake. These conditions are thought to have provided the ideal environment for the expansion of lake periphery/wetland types from ***c.*** AD 1520 – late 1880s. Finally, throughout the 1900s pastoral farming expanded as lake level dropped creating the current LPCA site dominated by pastoral and ruderal types and seasonal flooding as discussed in Chapter 3.

6.1.8 Lake Plav synthesis

This section aims to bring together a selection of records from cores LPCA and LPCB to provide a synthesis of the main changes at the Lake Plav site since ***c.*** AD 1100.

The beginning of the records in Figure 6.13 (***c.*** AD 1100), highlight the different processes occurring at sites LPCA and LPCB. At this time Lake Plav extended beyond its present day extent with < 20% sand grained material (>250 μm) registered between ***c.*** AD 1100 – 1300 in core LPCA. Although high quantities of sands are registered in core LPCB these are thought to represent deltaic deposits at Lake Plav. The first major change in the records was the *Alnus* collapse, which occurs synchronous to declining levels of sands in both cores LPCA and LPCB (Figures 6.13). This change is thought to be the result of anthropogenic activity in the catchment, possibly as a result of climatic deterioration as temperatures decline in the Northern Hemisphere around this time (e.g. Briffa *et al.*, 1988; Luterbacher *et al.*, 2002, 2004; Moberg *et al.*, 2005) and shortly after LPCA thermophilous types decline. The deterioration may have caused poor/failed harvests, a problem seen across Europe (Fagan, 2000), forcing people to utilise greater areas of land to try and maintain the necessary volume of food production.

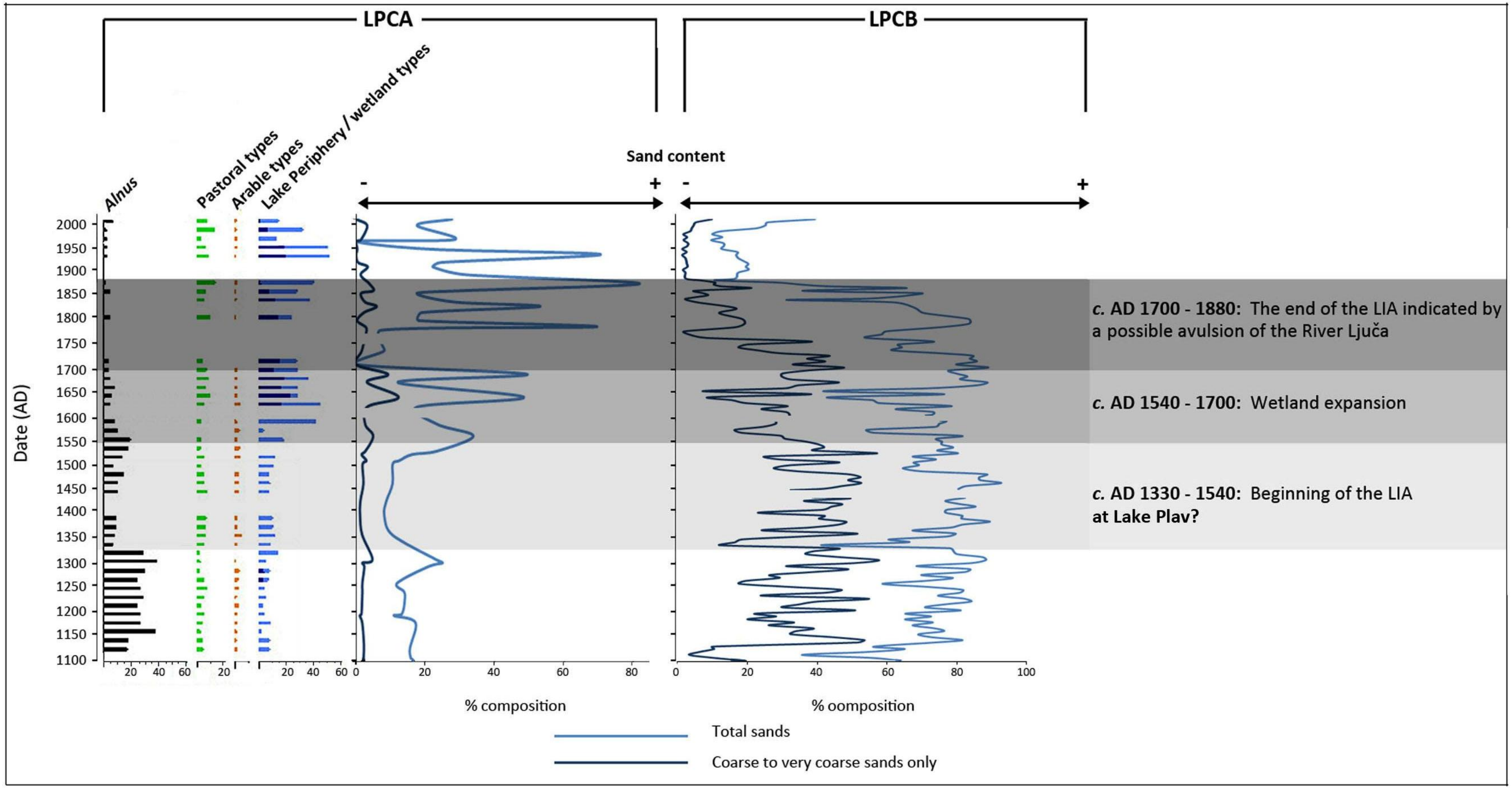


Figure 6.13: Selected record from cores LPCA and LPCB. Pollen are expressed as percentage TLP, with the dark blue lake periphery/wetland type being *Typha latifolia*.

Between ***c.*** AD 1330 – 1540 the total sands registered in core LPCA fall to a record minimum, indicative of rising lake levels. Increased precipitation is likely to cause lake level rise and this is supported by the coarse nature of the core LPCB record. A minor increase in core LPCB total sands indicates flood waters, which would enable larger volumes of water and sand sediments to be transported to the site. The records from core LPCA suggest that lake periphery/wetland vegetation (excluding *Typha latifolia*) began to expand ***c.*** AD 1540. Around this time total sands rise, thought to indicate deltaic expansion rather than falling lake levels. This would account for both the increase in core LPCA total sands (Figure 6.13), but also the low organic content of the sediments (Digerfeldt, 1986; Figure 5.22; Table 5.8, Chapter 5). By ***c.*** AD 1640 the lake periphery/wetland types are dominated by *Typha latifolia* and Cyperaceae which can withstand flooded habitats. Coupled with the continued evidence of inorganic coarse to very coarse sands in cores LPCA and LPCB, evidence that Lake Plav was infilling rather than the lake level falling is suggested. *Typha latifolia* becomes more dominant during the late 1600s as the volume of sands in core LPCA and LPCB increases. This suggests precipitation rose during the late 1600s allowing greater quantities coarse grained material to reach core LPCA causing the lake to infill and become shallower.

Total sands decline in core LPCA between ***c.*** AD 1700 – 1775, with a broad decline in the LPCB record between ***c.*** AD 1700 – 1880. The total sand records for LPCA and LPCB show an apparent conflict between ***c.*** AD 1775 – 1880, with rising values of sand in core LPCA and falling values in core LPCB. Coarse to very coarse sands fall below 20% by ***c.*** AD 1800 in core LPCB suggesting that the site was less affected by flood waters. In core LPCA fine to medium sands begin to peak. The lake periphery/wetland indicators remain dominant, but levels of *Typha latifolia* fall. Therefore, during ***c.*** AD 1775 – 1880 it is possible that falling precipitation lead to an overall reduction in the volume of sands reaching core LPCB. Reduced precipitation would inevitably cause lake levels to fall, increasing the proximity of core LPCA and LPCB to the lake shoreline. By ***c.*** AD 1880 an avulsion of the River Ljuča, north-east towards Visitor Mountain (Figure 6.3) caused the sudden reduction of sands in the core LPCB record (Figures 6.13, 6.14).

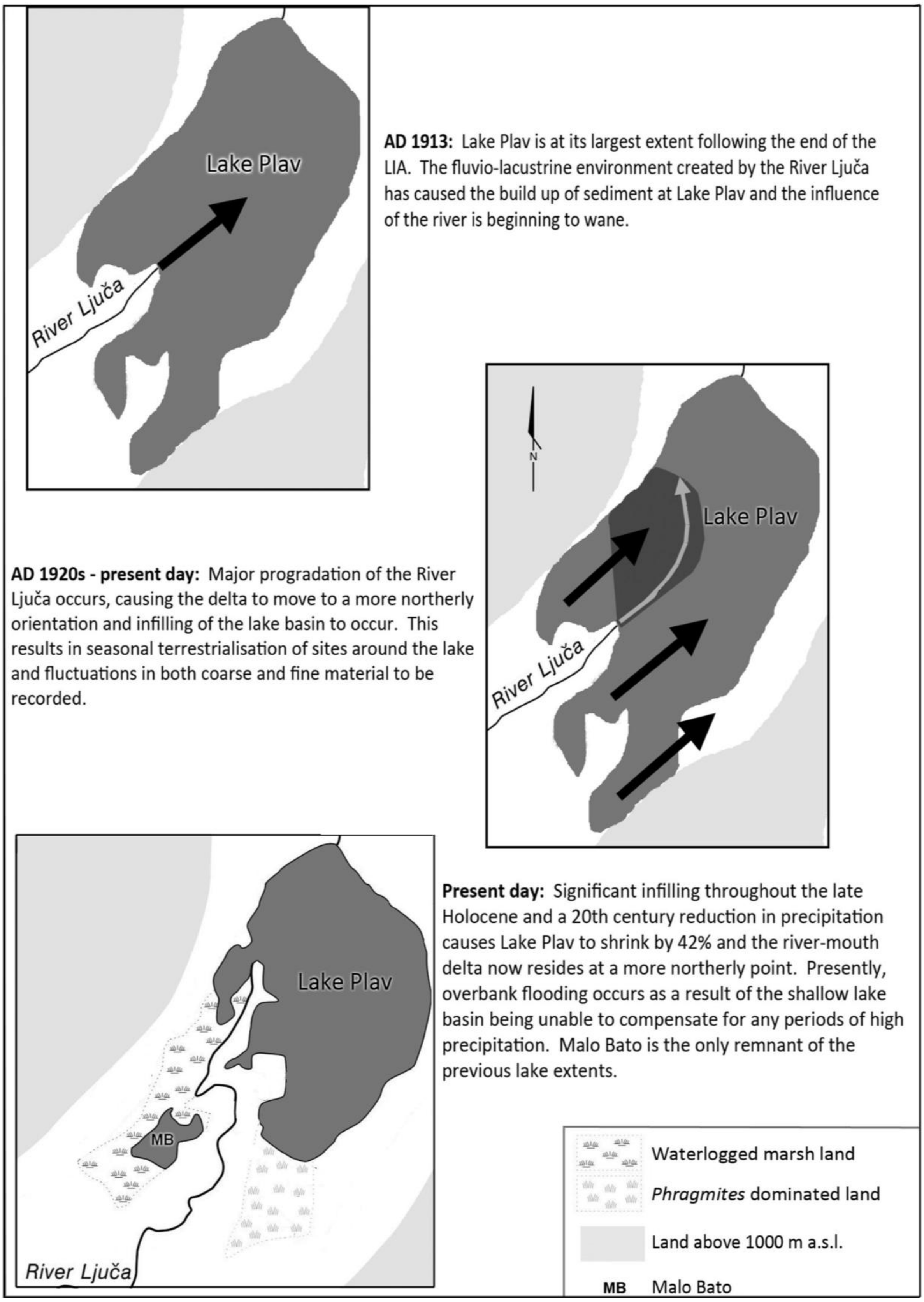


Figure 6.14: Schematic illustration of the changing extent of Lake Plav since Cvijić mapped the site in AD 1913.

Sediments from the early 1900s are characterised by > 80% fine grained material and un-polleniferous sediments. At this time Cvijić (1913) indicates that Lake Plav was 42% larger than its current extent. Although the map lacks any indication of lake depth around the basin the LPCA and LPCB records suggest the lake may have been at its shallowest non-terrestrial phase (Figure 6.13). Maps of the site dated to AD 1939 (Figure 6.5) indicate that Lake Plav was close to its current extent. As climatic amelioration generally characterised AD 1913 – 1939 across Europe and the Mediterranean (e. g. Schweingruber *et al.*, 1988, 1991; Serre-Bachet, 1994; D'Orefice *et al.*, 2000; Luterbacher *et al.*, 2004) it is suggested that the reduction in lake extent was a result of falling lake level exposing shallow lake areas. Sedimentation rates fell dramatically at this time by up to 1.133 cm yr^{-1} and current relatively shallow depth of Lake Plav (9 m) support a theory of falling lake level as opposed to rapid inundation. Deeper areas of the lake basin remain inundated, such as Malo Bato, thought to be a remnant of the past lake extent (Figure 6.14). During the 1940s to present, both LPCA and LPCB have become seasonally terrestrial. Records from core LPCA indicate overbank flooding by peaks in sandy material and terrestrialisation of the site by rising LOI levels (Figure 6.6). Figure 6.14 provides a summary of the changes indicated by the core LPCA and LPCB data since Cvijić (1913) mapped the site.

6.2 LAKE C, BUNI I JEZERCES SITE

The record from Lake C above the tree-line aimed to provide a more climate driven record, further from anthropogenic influences. This provides a palaeoclimate record for the Lake Plav catchment and direct comparison to the glacial record near-by (see section 6.3).

6.2.1 The stratigraphic record of BJC1

Organic fragments were found in the lower 5 cm of the core and at 113 cm, with the latter being radiocarbon dated. Lake C is a seasonal lake with much of the water evaporating during the summer months. The mottled appearance of the BJC1 sediments may be the result of either bioturbation, whereby different colours occur around burrows, or reducing conditions, which cause mineral changes and new minerals to form (Tucker, 2003; Schaetzl and Anderson, 2005). Bioturbation was mentioned earlier in section 5.2.3, chapter 5, as the Cs^{137} record in the upper units of core BJC1 seemed broad, but there is no mottling in the upper section which suggests an alternative source of discolouration. Mottling dominated the colour of the core below 60 cm, but other records (pollen and magnetic) show no evidence for bioturbation at these depths. The short-lived spike in magnetic susceptibility at 110 cm (***c.*** 1660 BC) indicates a brief event, which, under biological mixing would have been characterised by a lower value and importantly a broader shape. The palynological records are similarly diverse, with variance notable in for example, *Plantago lanceolata, Cruciferae*, *Scutellaria* type, *Saxifraga* and *Pediastrum* records. Therefore, the mottled sediments are thought to reflect the reducting conditions in the sediments of core BJC1.

6.2.2 Magnetic susceptibility record for Lake C (core BJC1)

Magnetic susceptibility at Lake C remained relatively low throughout the mid- to late Holocene (Figure 6.15), but two periods of increased magnetism can be dated to ***c.*** 1660 BC (BJMS I) and ***c.*** AD 150 – 640 (BJMS II). Around 110 cm (***c.*** 1660 BC) a large spike in magnetic susceptibility suggests that erosion in the catchment increased, inputting more magnetic material to the lake. It is difficult to indicate the probable source of this, as the surrounding geology is limestone dominated and as the geological map (Figure 3.11, Chapter 3) does not cover the Albanian Buni i Jezerces site any e.g. sandstone outcrops may not be identified. The dominant types around the catchment are dark grey organic limestone, thick bedded dolomitic limestone and dolomite. Inputs from the latter are likely to have increased the magnetic susceptibility, as Shogenova (1999) indicated that some dolomites can have enough iron content to change magnetic properties. Further work upon the mineral composition of the core BJC1 sediments would be required to determine this. The Thera eruption occurred ***c.*** 1645 – 1600 BC, the date of which is much contested (e.g. La Marche and Hirschboeck, 1984; Zielinski and Germani, 1998; Friedrichs *et al.,* 2009) coinciding with the core BJC1 spike of magnetic susceptibility. However, as tephra from the eruption has been identified in areas to the east of Santorini, in the Black Sea and South-Eastern Mediterranean (e.g. Guichard *et al.*, 1993; Eastwood *et al.*, 1999; Wulf *et al.*, 2002) it seems unlikely that the spike in magnetism was the result of a tephra layer. The eruption is thought to have caused global climatic perturbation (Friedrich *et al.*, 2006; Tsonis *et al.*, 2010), but this seems a very large rise in magnetic susceptibility to be the result of erosion around Lake C alone, considering no comparable spikes occur in the record. It is difficult to decipher the source for the magnetic susceptibility spike at 110 cm as there are no stratigraphic changes and no tephra particles were identified in the pollen samples due to the HF preparation of the samples. The current data suggest that a combination of processes may have occurred to create the spike. Therefore, local erosion coupled with surface weathering of the upper units may explain the size of the spike compared to the rest of the record, as both of these processes would increase sediment magnetic susceptibility.

Erosion in the catchment is suggested to have increased again ***c.*** AD 150 – 640, indicated by a rising magnetic susceptibility. This may have been a product of increased precipitation at the site washing greater quantities of allochthonous material into Lake C. However, the *Pediastrum* and thermophilous types records suggest this period was characterised by warm and dry conditions, which are less conducive to erosion. Anthropogenic indicators at this time are relatively stable, but *Poaceae* does increase suggesting possible land clearance in the area, which may have caused erosion in otherwise unsuitable conditions. A later rise in phase BJMS IV suggests intensification of eroded inputs at the site from ***c.*** AD 1620 to the present. This also coincides with wetland expansion seen in the lower catchment, which may together with rising *Pediastrum* be indicative of rising precipitation (Figure 6.15).

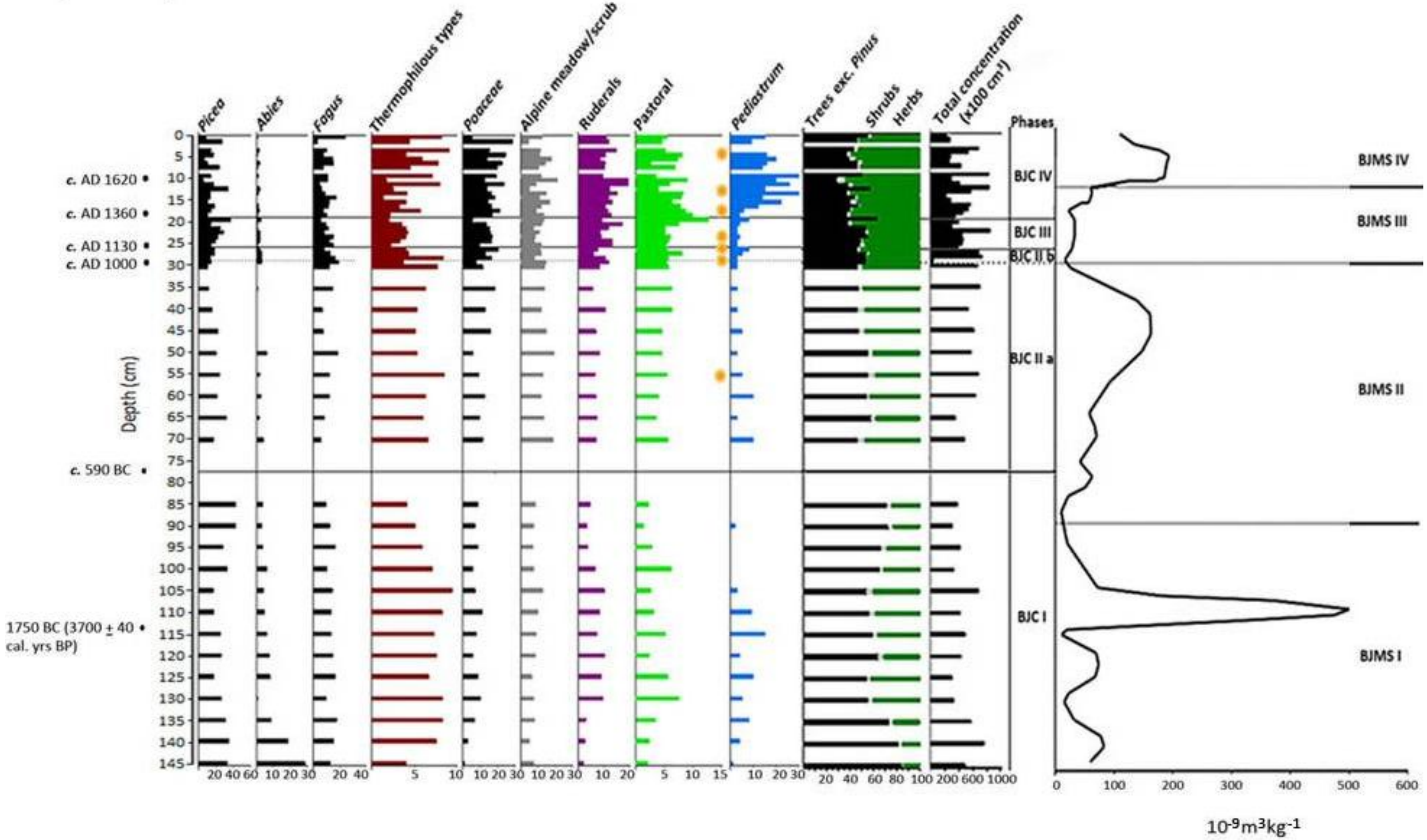


Figure 6.15: Stacked indicators with pollen concentration and magnetic susceptibility records from Lake C, Buni i Jezerces. Pollen is expressed as percentage TLP excluding *Pinus*, extrapolated dates are indicated by ***c.*** and the presence of arable types by ●.

6.2.3 Lake C palynological profile

Figure 6.15 presents the four indicator groups using TLP (exc. *Pinus*) on the BJC1 data: thermophilous types, alpine meadow/scrub, ruderals, arable and pastoral following Table 6.4. An addition to these groups was the aquatic *Pediastrum* which has been used as an indicator for environmental and climatic changes. Thermal changes at the site were ascertained from the record using the aggregated thermophilous types (see Table 6.4). In addition, following Harrison and Digerfeldt (1993, cited in Edwards *et al.*, 2000) levels of *Pediastrum* have been used to indicate water depth and thus climate. Sarmaja-Korjonen *et al.* (2006) and Edwards *et al.* (2000) used the taxon to indicate both climate changes and lake productivity, with lower levels of *Pediastrum* associated with a cool climate and poor lake productivity. Reynolds (1980) and Weckström *et al.* (2010) used *Pediastrum* as a proxy for nutrient levels in lake environments. As Lake C is seasonal the limiting factors upon *Pediastrum* growth is most likely water depth and to some degree nutrient levels.

During the last 4,730 year (***c.*** 2720 BC extrapolated base date) vegetation Lake C surrounding has been dominated by coniferous montane forest and deciduous *Fagus* woodland with near absent levels of *Poaceae* pollen at the base of the core. *Abies* declined in-line with a reduction in pollen concentration to its lowest levels of <26,000 g cm^3, ***c.*** 2270 – 2110 BC and these trends were accompanied by rising levels of thermophilous types, alpine meadow/scrub, ruderal and pastoral groups with *Poaceae* also increasing. Synchronous with this, levels of *Pediastrum* began to rise, indicating that lake level was rising and that the climate may have been more humid (e.g. Veski, 1994; Edwards *et al.*, 2000; Weckström *et al.*, 2010). Thermophilous and alpine meadow/scrub taxa remained below 10% as *Abies* recovered to ***c.*** 10% (125 cm; ***c.*** 2110 BC), and *Picea* levels began to wane suggesting that the dynamics within the local montane forest may have been changing, rather than as a result of climatic variability. *Pediastrum* peaks at 115 cm (***c.*** 1810 BC); just prior to the largest magnetic susceptibility spike at 110 cm which coincides with slightly reduced pollen concentrations and montane types, but relatively stable pollen abundance in the remaining taxa. The depletion of coniferous types ***c.*** 1660 - 1490 BC (see Figures 6.15, 5.32b, Chapter 5) coincides with the a suggested erosion and surface weathering event and the reduction may be a result of dominant source areas changing, rather than an actual reduction in coniferous types. Thermophilous types did not appear responsive to this event as a group, but *Corylus* actually rises during this period as others remain stable or decline, again suggesting the taxa benefited from the changes occurring at this time. Alpine meadow/scrub and ruderals increase ***c.*** 1500 BC as levels of *Pediastrum* fall suggestive of drier conditions allowing expansion of certain taxa. Following this event pollen concentrations fall to < 40,000 g cm^3 with *Pediastrum* becoming sparse, but *Picea* rises to its highest levels at 90 – 85 cm (***c.*** 1000 - 840 BC) and thermophilous types plateau implying slightly drier and warmer conditions.

Throughout BJC IIa *Pediastrum* levels recover, with *Fagus* and *Picea* at levels lower than at the end of BJC 1. This is accompanied by indication from the ruderal and pastoral records that grazing herds may be using the area at this time, although *Picea* does recover indicating the tenacity of the local montane forest, with *Abies* becoming scarce. Throughout BJC IIa (***c.*** 580 BC – AD 1000) *Poaceae* and alpine meadow/scrub seem to vie for dominance, whilst there is also evidence for remote arable farming during this phase. Magnetic susceptibility begins to rise mid-BJC IIa (***c.*** AD 0 - 100), indicative of rising detrital inputs suggesting climatic deterioration as thermophilous types wane from ***c.*** 8% to ***c.*** 5%. Montane forest also seemed to suffer at this time, with *Picea* frequencies falling and the conditions favouring herb types such as *Artemisia* and *Lactuceae*, which prefer open conditions (Collins *et al.*, 2005). Detrital inputs fall by the end of BJC IIa, but pollen concentrations remain relatively stable and as magnetic susceptibility plateaus during BJC IIb and III, thermophilous types rise. Falling detrital inputs may well have been the result of a drier or warmer climate between ***c.*** AD 800 – 1300 which would coincide with the MWP (***c.*** AD 800 - 1200; Broecker, 2001) and goes someway to explain the rise in thermophilous types between 40 – 28 cm (***c.*** AD 640 – 1000; Figure 6.15). ***c.*** 1070 (27 cm) thermophilous taxa collapse from 8.1 – 4%, synchronous with further evidence for arable farming, remaining ***c.*** 0.5% between 27 – 12 cm (***c.*** AD 1070 - 1560). This collapse may therefore, be attributed to anthropogenic or climatic factors, but as arable types form a minor component, it is likely that the change was driven by climatic deteriorations. The end of BJC III is characterised by a peak in *Picea* against a much reduced abundance of thermophilous types (< 4%) and this is thought to indicate cooling temperatures, but as *Pediastrum* remains stable no real change in humidity.

By BJC IV (***c.*** AD 1360) ruderal types and *Poaceae* expanded and *Picea* declined, with the *Pediastrum* record providing evidence of a general rise in humidity between ***c.*** AD 1430 – 1700 consistent with many records of LIA climatic change (Lamb 1965, 1977; Jalut *et al.*, 2000; Cronin *et al.*, 2003; Trouet *et al.*, 2009). Total pollen concentrations were variable throughout this period, possibly as a result of climatic perturbations, but most likely a product of changing erosion rates within the catchment. However, at 10 cm (***c.*** AD 1620) thermophilous taxa rise above 5%, where they remain for much of the following record, suggesting climatic amelioration. Coincident to a warming climate magnetic susceptibility rises and levels of *Pediastrum* decline to <20 (raw count). The rising detrital inputs from ***c.*** AD 1620 may have been the result of glacial recessions at the site. However, the decline in *Pediastrum* suggests seasonal evaporation of Lake C, as erosion would most likely create an unstable and enriched water column, cited by Reynolds (1980) as ideal conditions for *Pediastrum* expansion, in contrast to the record presented.

During the uppermost sediments of the BJC1 pollen profile (3 – 0 cm), the rapid increase in thermophilous types ***c.*** AD 1850 is accompanied by the disappearance of *Pediastrum*. Following the interpretation that *Pediastrum* levels denote lake level changes, this would suggest that either Lake C evaporated or was frozen for a period of time. Using the changes in thermophilous taxa it is thought that Lake C most likely seasonally evaporated as climatic amelioration at the end of the LIA occurred. By the end of the profile changes between ***c.*** AD 1950 and present day are pronounced. Even though pollen concentrations were stable, *Fagus* rose by 500% as thermophilous types increased by around a third, and both *Poaceae* and *Picea* declined, suggesting a switch towards *Fagus* in the mountain forest throughout the 20^{th} and 21^{st} centuries, a change not replicated in the LPCA record.

Overall, levels of thermophilous types continually shifted throughout the 1600s – present day from as little as 1.8% at 10 cm (***c.*** AD 1620) to 8 – 9% at: ***c.*** AD 1590, ***c.*** AD 1750, from ***c.*** AD 1850 to the present day. These periods correspond to periods of glacial expanse across Europe (e.g. Stotter, 1846 and Michelier, 1887 cited in Grove, 1988; González Trueba *et al.*, 2008; Ivy-Ochs *et al.*, 2009), but as these dates were extrapolated from the radiocarbon date at 113 cm it is difficult to constrain these periods more effectively. The thermophilous pollen record characterises three periods: phases BJC I – BJC II (***c.*** 2700 BC – AD 1100) average 6.5% thermophilous taxa, BJC III – BJC IV (10 cm; ***c.*** AD 1100 - 1620) were cooler with 3.36% thermophilous taxa and the latter half of BJC IV by a typical 6.25% (***c.*** AD 1620 - present). This suggests that in the upper catchment, the LIA climatic deterioration took effect around AD 1300, before the latter half of the LIA i.e. AD 1600 – 1800 was characterised by heterogeneous climatic conditions and a certain degree of warming. Levels of *Pediastrum* begin to wane by the 1800s and are absent ***c.*** AD 1820, this absence could have resulted from lake desiccation or snow cover throughout the year. This would coincide with glacial expansion in the European Alps and Pyrenees (Grove, 1988; Julián and Chueca, 1998; Cía *et al.*, 2005) and below average summer temperatures in Southern European temperatures and the Calimani Mountains in Romania (Briffa *et al.*, 2002; Popa and Kern, 2009). Since then *Pedisatrum* levels have remained relatively stable suggesting that Lake C became the lake seen today during the 20th century.

6.3 RECONSTRUCTION OF THE MAJA E KOLJAET GLACIER

Glacial geomorphological mapping within the Lake Plav catchment has enabled the reconstruction of the only existing LIA glacial history in Northern Albania. The work carried out at the Maja e Koljaet glacier enabled palaeotemperature reconstructions since the LIA glacial maximum (LIA_{GM}), providing an independent climate record for the Lake Plav catchment.

6.3.1 The Maja e Koljaet glacier

Glaciers are known to be sensitive to climate, with temperature and precipitation being the most influential factors upon a glacier (Ohmura *et al.*, 1992). In particular, small glaciers in more marginal areas have been found to be particularly sensitive to climatic variations throughout the LIA (e.g. Grove, 1988; Grudd, 1989; Kuhn, 1993; Kuhn *et al.*, 1997). During the LIA two periods of glacial expansion have been identified at ***c.*** AD 1650 and ***c.*** AD 1850 (Matthews and Briffa, 2005). The climatic perturbations associated with the LIA are recorded in the form of end moraines at glaciers across the Mediterranean and Europe (Thompson Davis, 1988; Winkler and Matthews 2010). However, glaciers are inherently erosive and glacier advances can remove evidence of less extensive earlier glaciations (Gibbons *et al.*, 1984). Therefore, sometimes only historical evidence can provide an idea of previous glacial extents at a site where more recent advances have eradicated those that occurred before. As historical documentary information was unavailable for the Maja e Koljaet glacier, the glacial history of the site could only be reconstructed using geomorphological evidence from the maximum glacier extent of the LIA, which occurred during the early to mid 1800s, to the present day. Nevertheless, this 150 - 200 year period spans a significant interval that is also recorded in the lake cores retrieved from both Lake Plav and Lake C at Buni I Jezerces.

Throughout the LIA, Mediterranean regional equilibrium line altitudes (ELA) ranged from 2600 m a.s.l. in the west to 2400 – 2500 m a.s.l. in the east. LIA conditions allowed glaciers in areas such as the Cantabrian Mountains and Sierra Nevada to extend to 2190 m a.s.l. (González Trueba, 2003/2004, 2005; González Trueba *et al.*, 2008) and down to 1740 m a.s.l. for the Miage glacier in the Mont Blanc Massif (Revelli, 1911 cited in Imhof, 2010). Reconstruction of Maja e Koljaet ELA suggests that whilst the ELA has remained relatively stable at between 2025 – 2040 m a.s.l., yet the glacier has existed at altitudes nearly 400 m below that of other Southern European glaciers located at similar latitudes. The low ELA and frontal elevation of Maja e Koljaet highlights the importance of local topographic controls in the development and maintenance of the glacier. Low insolation and avalanching snow would have provided ideal conditions for the accumulation and preservation of snow at the site, as the northwest to northeast orientation of the glacier and enclosure by steep cliffs and peaks provides deep shading and abundant source of avalanching snow (Hughes, 2009b).

Maja e Koljaet was a small glacier during the LIA, with a maximum area of 0.538 km^2. Although this is six times bigger than the present area of the glacier (0.084 km^2) the glacier median altitude still had a similar ELA because of expansion at upper, lateral and lower margins. As such, the glacier would have still been dominated by local topoclimatic controls in much the same way as it is today, especially with respect to accumulation from avalanching snow. Accumulation at the site would have still been affected by avalanching, (and also to a lesser degree windblown snow) in addition to direct meteorological precipitation.

As there is no other information regarding Albanian high mountain glaciers in marginal glacial environments during the LIA, this thesis offers the first reconstruction of LIA glacial development in the country. Figures 5.36 and 5.39 presented in Chapter 5, suggest that the Maja e Koljaet glacier reached its furthermost extent around AD 1859/1864, but prior to this in AD 1834 the glacier had expanded vertically creating a lateral moraine between 1900 - 1950 m a.s.l.. The LIA_{GM} is taken as AD 1859 because this is associated with the lowest altitude the glacier reached during the LIA. Lichenometric dating of the lateral moraine at the site suggest that the glacier retreated from this moraine around AD 1834 before retreating from the hummocky moraines further down valley around AD 1859. Therefore, the glacier first preferentially lost ice depth before losing glacier length. It is likely that thickening of the glacier around AD 1834 was the result of a rapid positive increase in net mass balance creating a 'bulge' over the glacier surface. This was then later expressed down-valley by frontal advance of the glacier recorded at AD 1859. This would suggest a 20 - 25 year lag in response between positive net balance expressed by glacier thickening and vertical expansion in the upper cirque and the transfer of mass down-valley to a lower glacier front. Relics of the LIA LIA_{GM} at the site highlight that any previous glacial extensions at the site were smaller in size and were overridden by the later LIA maximum.

The Maja e Koljaet LIA glacier maximum occurred just after the end of the LIA_{GM} seen across the Western Mediterranean (AD 1845-1856), but coincided with the beginning of a warm period in the Calimani Mountains, Romania, suggesting that the site may have been influenced by the climatic variations seen in this area (Popa and Kern, 2009). Furthermore, the LIA glacier maximum in Albania seems relatively detached from that of Northern Montenegro (Debeli Namet glacier, Durmitor Massif) where the glacier maximum appears to have occurred just under 20 years later in AD 1878. Although, Hughes (2007) calculated a mean percentage error of ***c.*** 11% in the regression between observed lichen size and the predicted lichen size for *Aspicilia calcarea* – the same lichen growth regression used in this thesis. This could mean an error of 15 – 20 years. Nevertheless, possible differences between the timings of the LIA_{GM} in the Durmitor and Prokletije Massifs can be explained by the local topographic effects and conditions which can influence the glacial development between seemingly 'close' massifs (Hughes, 2004). The effects of the NAO could also account for the variance, if the Maja e Koljaet was more influenced by variations in the index. Periods of reduced NAO index are associated with anomalous northerly flow across the Mediterranean (Hurrell, 1995) which could cause greater quantities of wind-blown snow onto the glacier from the Maja e Koljaet and Maja e Jezerces peaks located southeast of the Maja e Koljaet glacier. These conditions were seen throughout the late 1700s and early 1800s which may have contributed to the extension of the glacier during this time. Furthermore, the NAO index rose between the 1850s and 1860s to a positive state, leading to warmer conditions in Southern Europe and the Mediterranean (Hurrell, 1995; Hurrell and van Loon, 1997) coinciding with the exposure of the terminal moraine created by the maximum glacial extent of the Maja e Koljaet glacier. However, it must be noted that the synchronicity within the Maja e Koljaet record and NAO index may not account for the 11% error associated with the lichen regression used within this thesis, but nevertheless remains a point of concern.

The no. 3 moraine (Figure 5.36, Chapter 5) suggests that glacier advance or stabilisation also occurred at AD 1908/1922 coinciding with the two early 20$^{\text{th}}$ century advances or stabilisations at ***c.*** AD 1904 seen in Montenegro (Hughes, 2010) and the Western Mediterranean (Grove, 1988). However, as noted earlier the error associated with the lichen growth regression may mean the age differences are not statistically significant. Nevertheless, differences in dates may be indicative of uneven ice retreat or variable conditions for lichen colonization. If it is argued that the ages are statistically indistinguishable then the later AD 1922 obtained for the Maja e Koljaet glacier may be the most accurate age representing glacier advance or stabilisation in this area of the Balkans and actually coincides with a negative NAO (Cook *et al.*, 2002; Luterbacher *et al.*, 2002). Thus local topographic conditions may have controlled the glacial retreat until this time (González Trueba, 2003/2004; González Trueba *et al.*, 2008; Milivojević *et al.*, 2008; Hughes, 2009b), with regards windblown snow and the NAO (see preivous paragraph).

Figure 6.16: Doline in front of the current Maja e Koljaet glacier and shaded by the steep northwest facing slopes of the surrounding relief. The AD 1859 glacier terminus is marked by hummocky moraines below this doline (out of picture) whilst the no.3 (AD 1908/1922) and no.4 (AD 1928/1944) moraines are situated around the doline.

The no.4 lateral moraine (Figure 5.36, Chapter 5) has yielded lichen ages of between AD 1928 to AD 1944, suggesting that it may have been similarly and irregularly uncovered by ice. Lichen would only colonise the rock beneath when the conditions for colonisation were suitable, possibly leading to this variance in age. Again this coincides with glacial expansion in the Western Mediterranean and a negative NAO (Cook *et al.*, 2002). During the exposure of the terminal (no.3) and lateral (no.4) moraines, the doline, identified in Figure 5.36 (Chapter 5) and Figure 6.16, may well have continued to contain ice throughout this period, a suggestion made by Hughes (2010) in his study of these features in the Durmitor Massif. However, as few suitable lichen were found for dating within this feature it is difficult to assess when the doline became ice-free. The bleached appearance of the doline floor and sparse lichen could suggest that the site allows snow preservation for significant periods of the year, thus not providing conditions conducive to lichen colonisation.

The final moraine to be dated at the site corresponds to a period of glacial expansion in the Alps throughout the 1970s (D'Orefice *et al.*, 2000). At this time the climate seems to vary across Europe and the Mediterranean, with dendrochronological records in Europe registering the beginning of recent warming in the mid to late 1970s (Luterbacher *et al.*, 2004; Cía *et al.*, 2005). Similar records from Romania suggest cool conditions persisting throughout the 1970s (Popa and Kern, 2009) and climate records from Podgorica in Montenegro, indicate that the late 1970s was a period with cooler summer temperatures (Hughes, 2008). This suggests that although the Maja e Koljaet glacial variations correspond to those of the remaining glaciers in the Western Mediterranean, the variations are also synchronous with climatic variations occurring across Eastern Europe. Furthermore, with the presence of ice along the western trim-line at Maja e Koljaet glacier the expanse of ice or possibly snow cover may have remained beyond the period of glacial retreat seen in the Western Mediterranean. The two moraines in front of the present day glacier which were free from lichen may have been formed between 10 and 20 years, the lag time for *Aspicilia* type (Miller, 1969; Hughes, 2009b), and could possibly correspond to the glacial advances of AD 2004 and 2006 described at the Durmitor Massif by Hughes (Hughes, 2008).

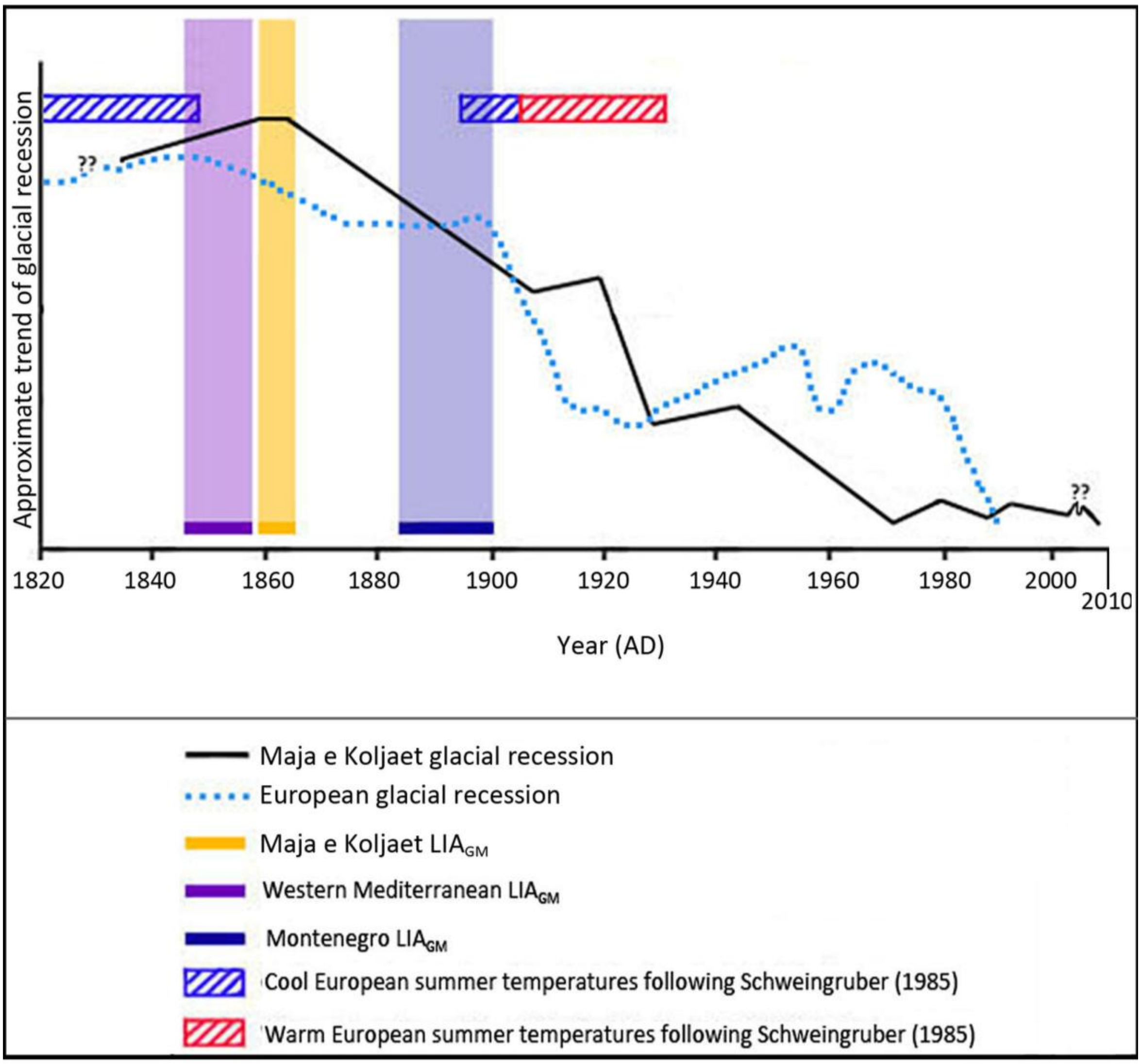


Figure 6.17: Comparison of Maja e Koljaet and European glacial recession at the end of the LIA.

The LIA timing of the Albanian glacial maximum, following the Maja e Koljaet glacier, occurred just after that of Europe and the Western Mediterranean and before the Montenegro LIA_{GM} (see Figure 6.17). Glacial recession of Maja e Koljaet generally lags that seen in Europe by around a decade, with the general shape of the curves presented in Figure 6.17 being similar until the 1920s. Interestingly the Maja e Koljaet glacier does not record any extension during the late 1800s, even though this period represents expansion or glacial maxima across the Mediterranean coupled with cooler temperatures in Europe (e.g. Schweingruber, 1985; D'Orefice *et al.*, 2000; Hughes, 2007, 2009b). This indicates the influence of local topographic factors not only on the expansion of the Maja e Koljaet glacier, but also its retreat. By 1920, the halt to glacial recession and expansion of glaciers across the Mediterranean is mirrored by the Maja e Koljaet glacier suggesting that the glacier is possibly responding to climatic variations causing similar changes across Europe and the Mediterranean. However, the divergence from the European glacial recession record is evident during the 1930/40s. During this time, local factors may have been particularly important or the climate of the Eastern Mediterranean may have become more influential. The latter suggestion is supported by recent variations at the Debeli Namet glacier by Hughes (2008), a glacier which seems to be more synchronous to climatic variations of the Eastern Mediterranean than Europe. During the 1970s the Maja e Koljaet glacier is relatively synchronous with Alpine and Mediterranean glacial records where glacier expansion was registered (Wood, 1988; D'Orefice *et al.*, 2000; Hughes, 2008). Two expansions at Maja e Koljaet glacier occurred in the past 10 to 20 years as the moraines associated with these expansions are lichen free, and thus may correspond to Eastern Mediterranean climatic variations.

Figure 6.18: Surface debris and the 'sunken' central core of the Maja e Koljaet glacier in October 2009. Active glaciers tend to exhibit a more convex cross profile, especially in the upper glacier area where positive mass balance creates a mass 'bulge' which is then propagated down-glacier to the snout.

The continued presence of the Maja e Koljaet glacier indicates that the site is favourable for glacier maintenance with shading and avalanching sustaining the glacier. However, the visit to the site in 2009 suggested that the glacier was dormant as no evidence of deformational structures (curved debris layers or crevassing) were visible on the ice surface and detrital material was visible across the upper ice surface (see Figure 6.18). The glacier also appeared 'sunken' (Figure 6.18) and lacked the convex surface profile of most active glaciers. This 'sunken' or low surface gradient characteristic is indicative of glacier decay *in-situ* (Quincey *et al.*, 2009). The presence of frontal moraines do suggest that the glacier has been recently active, and the presence of a measurement wire at least 3 m above the current ice level highlights the 'sunken' depression of the glacier central core (as discussed in section 3.1.3, Chapter 3). Although, in contrast to the Debeli Namet glacier in Montenegro which has been described as less sensitive to regional climate change by Hughes (2010) the Maja e Koljaet seems sensitive to some variations affecting European Alpine and Mediterranean sites until recent years.

6.3.2 Palaeoclimate reconstruction from Maja e Koljaet glacier

The mapped past glacial extents of the Maja e Koljaet glacier indicated that the glacier began to retreat from its minimum LIA altitude of 1800 m a.s.l. ***c.*** AD 1859, which was mapped as the '1800s' extent in Figure 5.39, Chapter 5. Results from the Maja e Koljaet glacier provide the first mapped LIA $_{GM}$ extents of any glacier in Northern Albania. The site provides the only record of LIA climatic variations and a solitary proxy climate record for the Lake Plav catchment, which continues to lack regular climatic monitoring.

The reconstructed temperatures for the glacier used the frontal moraine altitudes and were compared to contemporary climate data from Kolašin 40 km northwest of the site. However, the glacier did not occur within the range of the Ohmura *et al.* (1992) model for temperature reconstruction. Therefore, the DDM described in Hughes (2008) was used to assess the temperature depressions required for glacial extension to the altitudes mapped at Maja e Koljaet glacier. The altitudinal differences between the frontal moraines for the different glacier phases produces identical calculations of temperature depressions based on a lapse rate of 0.6°C per 100 m altitude (but mean annual temperature in the case of the degree-day model (DDM) and mean summer temperatures in the case of the Ohmura *et al.* (1992) regression. Both the DDM and Ohmura *et al.* (1992) regression data suggest that temperature depressions within the Lake Plav catchment would have been around 0.9°C at the LIA_{GM} i.e. ~AD 1859, which is consistent with other palaeotemperature reconstructions from elsewhere in Europe (Matthes, 1939; Grove, 1979, 1988; Lamb, 1995). Palaeoclimate reconstructions for the different glacier phases are outlined below and Figure 6.19 shows the reconstructed temperatures for the Maja e Koljaet glacier front (1800 – 1950 m a.s.l.) and Lake Plav (904 m a.s.l.) using the DDM data. The volume of accumulation predicted by each approach differs, with the DDM suggesting **6504 mm w.e.** and the Ohmura model calculating a figure of **5101 mm w.e.**. Contemporary values of precipitation around Maja e Koljaet are unknown because no climate stations exist. Estimations in the high mountains of this area have indicated between 2500 to 3000 mm yr^{-1} (Bošković and Bajković, 2004). Degree day modelling performed using extrapolated climate data and the methods described in section 4.12, Chapter 4 suggested the accumulation required to balance melting at the Maja Koljaet glacier is around **6504 mm w.e.** This is more than double the value of estimated annual direct meteorological precipitation and is likely to reflect the fact that accumulation via avalanching and windblown snow makes a major (in fact dominant) contribution to glacier mass balance in this area (Hughes, 2009).

Results from the DDM were seen as more reliable than the Ohmura model. The summer temperature data for Maja e Koljaet glacier is not represented on the polynomial curve created by Ohmura *et al.* (1992) and the temperature-precipitation combination required to sustain equilibrium glacier mass balance is beyond the scale of the curve (since none of the glaciers in the Ohmura model dataset of 70 glaciers worldwide exist under such warm and wet conditions as the Maja e Koljaet glacier). Consequently, the Ohmura model cannot be verified for accuracy for glaciers that exist under high summer temperatures and require high precipitation to survive (see Figure 4.14, Chapter 4). The reconstruction completed by degree day modelling is likely to be more representative, and modelling predicts that accumulation in excess of **6000 mm w.e.** is required to balance melting at the Maja e Koljaet glacier. It is possible that greater precipitation caused the Maja e Koljaet glacier to expand during the 19th century. There are few examples of glaciers around the world with accumulation values of more than 6000 mm, hence the problem with the Ohmura *et al.* (1992) curve, which is further compounded by the small size of the Maja e Koljaet glacier (<0.5 km^2 since the 1800s). If local controls dominate the mass balance regime of this small glacier, then small variations in direct meteorological precipitation are unlikely to have much impact. For this reason, temperature is likely to be most important control on glacier front fluctuations.

Date (AD)	1859	1908	1979	2009
Temperature depression required to allow glacial expansion (°C)	-0.9	-0.3	0	0
Reconstructed average annual temperature at Kolašin (954m a.s.l.)	7.63	8.23	8.53	8.53
Extrapolated average annual temperature for (904m a.s.l.)	**7.93**	**8.53**	**8.83**	**8.83**

Table 6.6: Reconstructed temperatures for Lake Plav (904 m a.s.l.), since AD 1859.

The DDM data suggest that the AD 1859 moraine formed under a temperature depression of -0.9 ^oC compared with present-day and this represents the temperature depression for the Albanian LIA_{GM}. This is in-line with other palaeotemperature reconstructions from across Europe which indicate depressions of 1 - 2^oC (Matthes, 1939; Grove, 1979, 1988; Lamb, 1995). Whilst the temperature depression reconstructed based on the glacier frontal variability is slightly lower this is insignificant because of possible variability in lapse rates used in temperature extrapolations. The temperature depression of -0.9 ^oC for the AD 1859 glacier assumes that precipitation (accumulation) was the same as today and that the advance in the glacier front was the result of temperature depression. Even if precipitation was greater or lesser, given the very strong local controls on accumulation (see below) it is unlikely that modest variations in direct precipitation would have made much difference to actual accumulation. Average annual temperatures increased after the LIA_{GM} by 0.6^{o}C to AD 1908 and then another 0.3^{o}C to the present day (see Table 6.6). Therefore, to identify the temperature change at the base of the Lake Plav catchment mean annual temperature (MAT) were modified to suggest that in AD 1859 the MAT at Lake Plav would have been 7.93^{o}C (Table 6.6).

Figure 6.19 shows the reconstructed temperatures for the Maja e Koljaet glacier front (1800 – 1950 m a.s.l.) and Lake Plav (904 m a.s.l.) using the DDM data. Temperatures from AD 1980 onwards have been extrapolated from the available Kolašin climate records (Tutiempo, 2009) to provide a more recent record of climate change within the Lake Plav catchment. The data highlight the 6.27^{o}C temperature range that exists across the catchment, with the area warming since AD 1959. The general trend since the LIA_{GM} in the Lake Plav catchment has been a warming pattern. Temperatures increased by 2.17^{o}C between AD 1859 and AD 2008, which was an exceptionally warm year compared to the AD 1978 - 2008 extrapolated mean for Lake Plav. Prior to the AD 1979 recession climate records from Kolašin suggest that the previous three winters were above the 0.5^{o}C winter (AD 1978 - 2008 Nov-Feb mean) average and also slightly drier than the mean. Up until AD 1980 the glacial record responded to the increasing temperatures by following a general trend of recession. This suggests that the Maja e Koljaet glacier responded to local climatic changes during the late 1970s causing it to retreat in AD 1979. Following this period, winters became cooler once more with winter temperatures averaging between -0.51^{o}C and -2.18^{o}C, similar to those seen at the LIA_{GM}. However, no expansion was identified in the 1980s and this may be due to the climatic deterioration of the early AD 1980s being too short lived for effective accumulation to cause expansion at the Maja e Koljaet glacier. Finally, the previous suggestion, that two of most recently formed moraines at Maja e Koljaet may correspond to the AD 2004 and AD 2006 moraines identified by Hughes (2008) at the Debeli Namet glacier, is supported by evidence of climatic deterioration throughout Montenegro and Albania at these times. The Kolašin climate data suggests that annual precipitation doubled between AD 2003 - 2006, with the majority falling during the winter periods. This would provide ideal conditions for glacier expansion as the higher precipitation would provide a source of accumulation and the lower temperatures would reduce the volume of ice lost via ablation. The continued presence of the Maja e Koljaet glacier in the marginal Northern Albania environment confirms the glaciers ability to respond to climatic variations, thus providing a reliable climate reconstruction for the Lake Plav catchment. In this sense, it appears that the glacier may well not be 'dead', and is better described as inactive or dormant. The critical point when such a glacier does become 'dead' is when the glacier loses sufficient ice mass that it does not present a dynamic response to short-term colder or wetter climate changes. At the time of observation (in 2009), the Maja e Koljaet glacier did not appear to have reached this point.

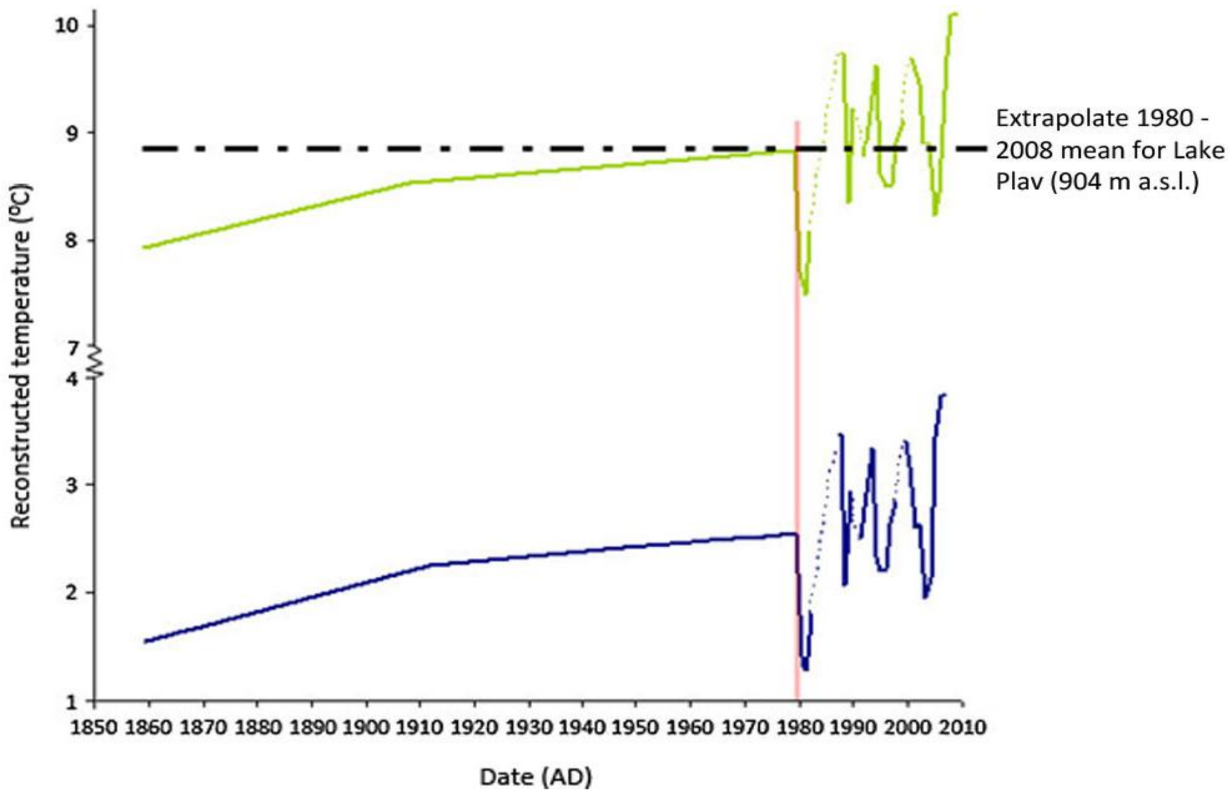


Figure 6.19: Reconstructed temperatures for Lake Plav (green line) and the Maja e Koljaet glacier (blue line) since AD 1859. The final glacial expansion at Maja e Koljaet is represented by a red line.

6.4 THE LAKE PLAV CATCHMENT DURING THE LATE HOLOCENE

This section aims to bring together the different records previously presented and discussed to synthesise the upper and lower catchment records, beginning ***c.*** AD 500.

6.4.1 Medieval times and the MWP at Lake Plav catchment, *c.* AD 500 – 1300

At approximately AD 500, both upper catchment and lower catchment records are available for comparison and during the period. AD 500 – 700 the area was characterised by relatively stable conditions. Particle size evidence, dominance of Candonidae and Ilyocyprididae ostracoda (Table 5.10, Chapter 5) and a varied pollen assemblage from core LPCA suggest that Lake Plav was a permanent water body extending southwest beyond core LPCA. Bell and Walker (1992) mention the MWP as a period of drier and warmer conditions beginning around AD 700. Evidence for this event is apparent around a century later, with thermophilous types in the upper catchment doubling between c. AD 640 - 800 whilst LF magnetic susceptibility fell in the upper catchment ***c.*** AD 700 - 800 (as indicated on Figure 6.20). The only change noted in the lower catchment was a slight increase in ruderal types, **c.** AD 750 (see Figure 6.10). Hughes and Diaz (1994) discussed the sensitivity of higher elevation dendrochronological records which responded to MWP warming, unlike low elevation records. All thermophilous types used were arboreal types, and they seem to have responded most noticeably to the MWP, in the upper catchment records which may have been more climatically marginal. Using these data the MWP appears to have begun later than other European sites at around AD 800. This may have been the result of the larger sampling intervals below 30 cm in the Lake C core (BJC1), or due to the remaining timescale uncertainty, but could be a genuine timing difference. In the lower catchment the presence of types such as *Juglans* and *Secale cereale* during the MWP suggest warm and possibly dry summer conditions that are necessary to ripen cereal crops (see Figure 5.25a, Chapter 5). This is corroborated by the low levels of lake periphery/wetland types which prefer damp to seasonally inundated habitats. Therefore, the wetland habitats that dominated from the 1600s to the present were less extensive during the MWP (see Figure 6.20). It is difficult to say whether lower lake levels were associated with the MWP at Lake Plav, as the particle size analysis, ostracoda record and varied pollen assemblage of core LPCA do not suggest reduced lake level. The particle size distribution of core LPCA between ***c.*** AD 500 – 1530 was dominated by silt particles (2 – 63 µm) indicative of a more profundal environment (Digerfeldt, 1986; see Figure 5.21, Chapter 5) and Candonids dominated the ostracoda assemblage, indicative of a permanent water body (Tables 5.9, 5.10, Chapter 5). Therefore, it seems unlikely that the lake level fell significantly.

Around AD 1000 thermophilous types in the upper catchment peaked to levels comparable to 20th century values, at a time when Greenland ice core records suggest temperatures were 1 – 2^{o}C warmer than present (Dansgaard *et al.*, 1975). Unlike the Greenland record, that from Lake C suggests that temperatures in the Prokletije Mountains were similar to those of the present day, rather than warmer. Precipitation was relatively stable at this time with indicators such as *Pediastrum* and total sands relatively stable (see Figure 6.20). By ***c.*** AD 1090 all but the thermophilous types of Lake C remain relatively unchanged (see Figure 6.20). The reduction in thermophilous types suggests that the Prokletije Mountains may have undergone a shorter MWP than in North-West Europe, between ***c.*** AD 800 – 1090, before climatic deterioration occurred. This compares well to the temperature curve presented in Figure 2.5, Chapter 2 (Moberg *et al.*, 2005) and the study by Serre-Bachet (1994) in North-Eastern Italy. Moberg *et al.* (2005) infer a minor temperature downturn at the end of the 1000s, followed by a continuous temperature decline in the mid 1100s. In North-Eastern Italy, two 'very cold' periods between AD 1100 – 1135 and AD 1170 – 1190 were inferred, prior to identification of the LIA between AD 1430 and AD 1860 (Serre-Bachet, 1994). The end of the MWP is difficult to define from the data presented here, as changes are limited before ***c.*** AD 1300 (Figure 6.20). Rather than showing an abrupt change to LIA conditions, the period between ***c.*** AD 1090 – 1300 is identified as the MWP/LIA transitional period at the site with climatic deterioration evident in the upper catchment thermophilous pollen record, but not evident in other proxies (see Figure 6.20).

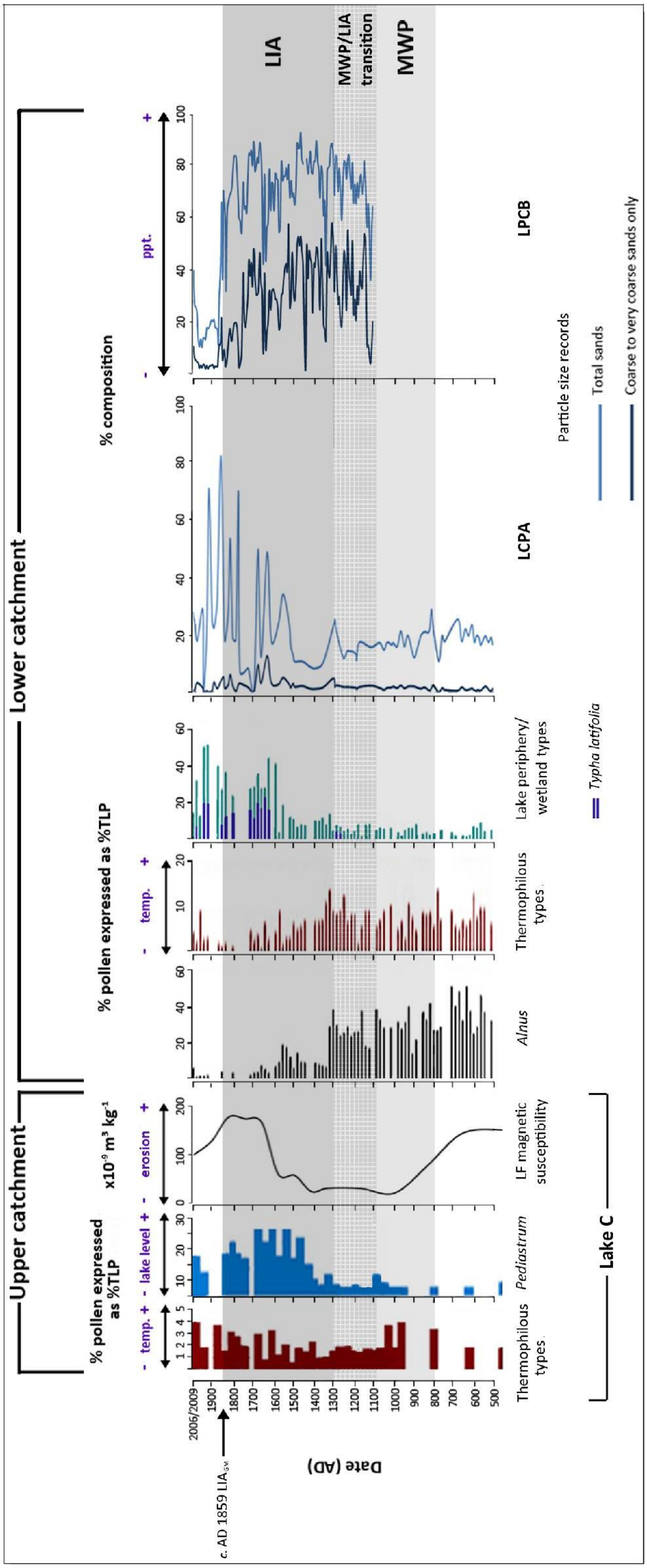


Figure 6.20: Synthesis of the Lake Plav catchment records since *c.* AD 500, with the MWP and LIA periods indicated by grey shading.

6.4.2 The LIA, *c.* AD 1300 – 1860

Around AD 1300 a number of changes are apparent in the catchment. Thermophilous pollen types fall below 2% of TLP and *Pediastrum* levels begin to rise at Lake C. The clearest changes occur in the lower catchment, where an *Alnus* decline between ***c.*** AD 1310 – 1330 coincides with falling thermophilous types and peaks in ruderals, and a sedimentary change to increased sand (see Figures 6.10, 6.20). The decline in thermophilous types that accompanies the *Alnus* decline lags similar changes in the upper catchment by around 200 years. This may be due to a number of reasons; either the greater sensitivity to minor climatic changes thought to exist in high mountain areas (as discussed in Chapter 1, Beniston *et al.*, 1997; Birks and Ammann, 2000); thermophilous types continuing to flower at lower more sheltered altitudes until temperatures became particularly cooler, or differences in the chronological accuracy of each core. It is possible that a combination of these effects caused the lag, but these reasons cannot be disentangled with the data available. The levels of sand, particularly coarse in the LPCB record ***c.*** AD 1300 indicates that precipitation in the region may have increased, causing coarser material to be deposited at site LPCB. This suggestion is supported by a doubling in Cyprididae types (Tables 5.9, 5.10, Chapter 5) ***c.*** AD 1350 which prefer flowing waters. Overall, the beginning of the 'full' LIA after the intermediate phase suggested above is defined in the lower catchment where evidence for a cooler and wetter climate is apparent from ***c.*** AD 1300.

Between ***c.*** AD 1400 – 1500, the Lake C records suggest wetter conditions and greater erosion at the site, following the trends in the *Pediastrum* and LF magnetic susceptibility records (see Figure 6.20). During this time thermophilous types at Lake C increase to 2.1%, but this abundance is half that seen during the MWP. Therefore, temperatures as the site may have risen, but were still comparatively low compared to the MWP. Around the AD 1550s an expansion in lake periphery/wetland types (excluding *Typha latifolia*) occurs, coinciding with elevated levels of *Pediastrum*. These changes suggest precipitation increased, enabling expansion of lake periphery/wetland types, possibly as previously terrestrial sites became inundated. The 1500s and 1600s in the Lake C records were characterised by continual changes in temperatures and lake levels coupled with greater erosion. Thermophilous types and *Pediastrum* records rose and fell simultaneously from ***c.*** AD 1500, suggesting that cooler periods were characterised by drier conditions. By ***c.*** AD 1620 thermophilous types and *Pediastrum* peak simultaneously with the largest expansion in *Typha latifolia* so far. The warmer, humid conditions indicated by the upper catchment records may have enabled the expansion of *Typha latifolia*. However, unlike previous minor expansions in *Typha latifolia* (see Figure 5.26, Chapter 5) the growth at ***c.*** AD 1620 is maintained throughout the cool and wet mid- to late-1600s, as it becomes the dominant lake periphery/wetland type through to the 1700s. This suggests that once *Typha latifolia* had colonised the area it could maintain its population through the unstable and cooler periods of the LIA (see Figure 6.20 upper catchment thermophilous types). By the mid-1600s the Lake C thermophilous pollen types declined to <1%, the second lowest abundance of the whole record, coupled with continued >20 spores of *Pediastrum* suggesting particularly cool and wet conditions, coeval to 'very cold' conditions in North-Eastern Italy. This supports documentary records from the Southeastern Mediterranean that indicate the Late Maunder Minimum (LMM) was characterised by cooler and wetter winter/springtime (Xoplaki *et al.* 2001). Overall the 1500 and 1600s support the previous suggestion by many authors (e.g. Jalut *et al.*, 2000; Cronin *et al.*, 2003; Trouet *et al.*, 2009) that the LIA was climatically heterogeneous, with shifts between warmer and cooler periods.

During the 1700s to ***c.*** AD 1850 a climatic amelioration similar to that of ***c.*** AD 1620 is evident within the thermophilous types at Lake C and water levels at the site fell slightly, but remained high compared to earlier records i.e. ***c.*** AD 500 – 1300. Overall, turbulent conditions are implied for the later phases of the LIA around Lake C. Erosion at the site was at its highest point since ***c.*** AD 500 – 600, which is echoed by the dominance of sands in the Lake Plav sediments. By ***c.*** AD 1750 coarse to very coarse sands begin to fall in abundance (see Figure 6.20, core LPCB) indicative of falling levels of peak precipitation. Around 50 years later the conditions that previously encouraged expansion of *Typha latifolia* seem to be waning, as Cyperaceae and *Equisetum* frequencies increase. The Albania LIA_{GM} at Maja e Koljaet (AD 1859) occurs just prior to a doubling of thermophilous types in the upper catchment alongside an absence of *Pediastrum,* inferring warm but dry conditions, but in the lower catchment thermophilous types decline. This disparity could have resulted from chronological differences between the glacial and sediment records from Buni i Jezerces (also see section 6.5.3). The abrupt disappearance of *Pediastrum* may represent a time when Lake C evaporated completely, or froze over, most likely the former as no similar disappearance occurs during apparently cooler times (indicated by minima in thermophilous types, Figure 6.20).

These records indicate that the LIA began ***c.*** AD 1300 at the Lake Plav catchment. The Lake C records indicate that the period was climatically heterogeneous, with temperatures fluctuating throughout the ***c.*** AD 1300 - 1860 period. However, evidence for continually increasing precipitation is supported by both upper and lower catchment records (particle size, *Pediastrum*, lake periphery/wetland types, ostracoda types). Levels of precipitation are not thought to have decreased until the late 1800s and this is supported by the non-polleniferous samples and low pollen concentrations noted in phase LPCA V (see Table 5.9, Chapter 5).

6.4.3 The catchment environment after the LIA, *c.* AD 1860 – present

The end of the LIA at the Lake Plav catchment was characterised by four changes: (1) a rapid increase in thermophilous types in the upper catchment, (2) a lack of *Pediastrum* at Lake C (3) expansion of Cyperaceae and *Equisetum* at Lake Plav and (4) a reduction in total sands in core LPCB of 60% at approximately AD 1880. These suggest that conditions became drier as temperatures rose at the end of the LIA, unlike earlier warm and wet phases. By ***c.*** AD 1880 infilling of the Lake Plav basin caused a possible avulsion of the River Ljuča, moving the river channel north and removing the source of coarse grained material to site LPCB. Lake Plav is thought to have remained larger than its present day extent, but must have become shallower as a result of infilling. During the early 1900s cooler and wetter conditions returned, as thermophilous types fell and *Pediastrum* levels increased in the upper catchment. This coincides with glacial expansion at Maja e Koljaet around AD 1908 – 1922 (see Figures 6.17, 5.36, 5.37 Chapter 5). Around this time Cvijić (1913) mapped Lake Plav, indicating that its extent was much larger than the present day (see Figure 6.1). Falling lake levels throughout the 20th century are implied, which may have made conditions for lake periphery/wetland types less hospitable. Anthropogenic activity in the area may have exacerbate this problem by draining or reclaiming marshland for cultivation which is supported by the continued clearance of the lower catchment for pastoral farming (see Figure 6.10). Rising temperatures accompanied these changes, evident in the thermophilous indicators of both the upper and lower catchment. During the late 20th to 21st century, a continued rise in temperatures is registered by thermophilous types across the catchment. The quantities of sands in the lower catchment records become indicative of floodplain sites affected by over-bank flooding and seasonal inundation. *Pediastrum* levels in the upper catchment recover in the uppermost 1 cm suggesting that Lake C has become deeper or more consistent, and that the current climate may be more humid than the late 1800s.

6.5 THE WIDER CONTEXT

The palaeoenvironmental records from the Lake Plav catchment have captured not only the climatic changes during the Late Holocene, but also changes in anthropogenic activity in the catchment, and the development and infilling of Lake Plav. Data from the Maja e Koljaet glacier provided a palaeotemperature record and the first available glacial reconstruction able to date the LIA_{GM} for Northern Albania in the Prokletije Mountains. During this section these records are compared to other Mediterranean and wider European datasets, to identify periods of correlation and non-correlation during the late Holocene. Where possible, connections to potential drivers of late Holocene climate are also made.

The variables used in the summary diagrams presented in this section have been chosen as they represent predominantly climate driven records. Furthermore, records from a variety of Mediterranean and European locations have been included to ensure a wide spatial coverage.

6.5.1 The period *c.* 2700 BC – AD 500

The longer time period has been reconstructed in less detail than the later phases, as it is not the main focus of the thesis. This discussion is included, however, on the basis that it provides a longer term context of change within which to place the LIA data. Records are available from Lake C, with an interpolated age at the base of ***c.*** 2700 BC, although the timescale must be considered provisional. Between ***c.*** 2700 - 1500 BC, thermophilous types suggest that temperatures in the catchment warmed to conditions comparable to the MWP, rising to more than 4% of TLP (see Figure 6.21). The climate at this time was also humid enough to support relatively high levels (for the Lake C record) of *Pediastrum*. Similar conditions are identified in central to South-Eastern Mediterranean records, with $\delta^{18}O$ foraminifera records from the Adriatic Sea (Piva *et al.*, 2008) and South-Eastern Mediterranean planktonic records (Schilman *et al.,* 2001) indicating warm and more humid conditions than those seen after ***c.*** 1500 - 0 BC. In the Western Mediterranean and central Europe contrasting conditions existed. Lower lake levels were identified at Zoñar Lake in Spain (Martin-Puertas *et al.*, 2008, see Figure 6.21) and in the Jura Mountains (Magny, 2004) and cooling temperatures are inferred by the foraminifera record from Tyrrhenian Sea (Antonioli *et al.*, 2001, Figure 6.21). These may be linked to the increased aridity that began 2200 BC mentioned by Weiss *et al.* (1993) and identified in cores extracted from the Alboran Sea, South-Western Mediterranean between 2050 – 850 BC by Martin-Puertas *et al.* (2010).

Drier and cooler conditions are evident in the Lake Plav catchment between ***c.*** 1500 - 840 BC. Similar temperature trends are echoed in both Tyrrhenian Sea foraminifera and the Adriatic Sea records (Antonioli *et al.*, 2001; Piva *et al.*, 2008). However, humid conditions were maintained until ***c.*** 1450 BC at Lake Van in Turkey (Gat and Magaritz, 1980) and in the South-Eastern Mediterranean ***c.*** 1250 BC (Schilman *et al.*, 2001). By ***c.*** 500 BC

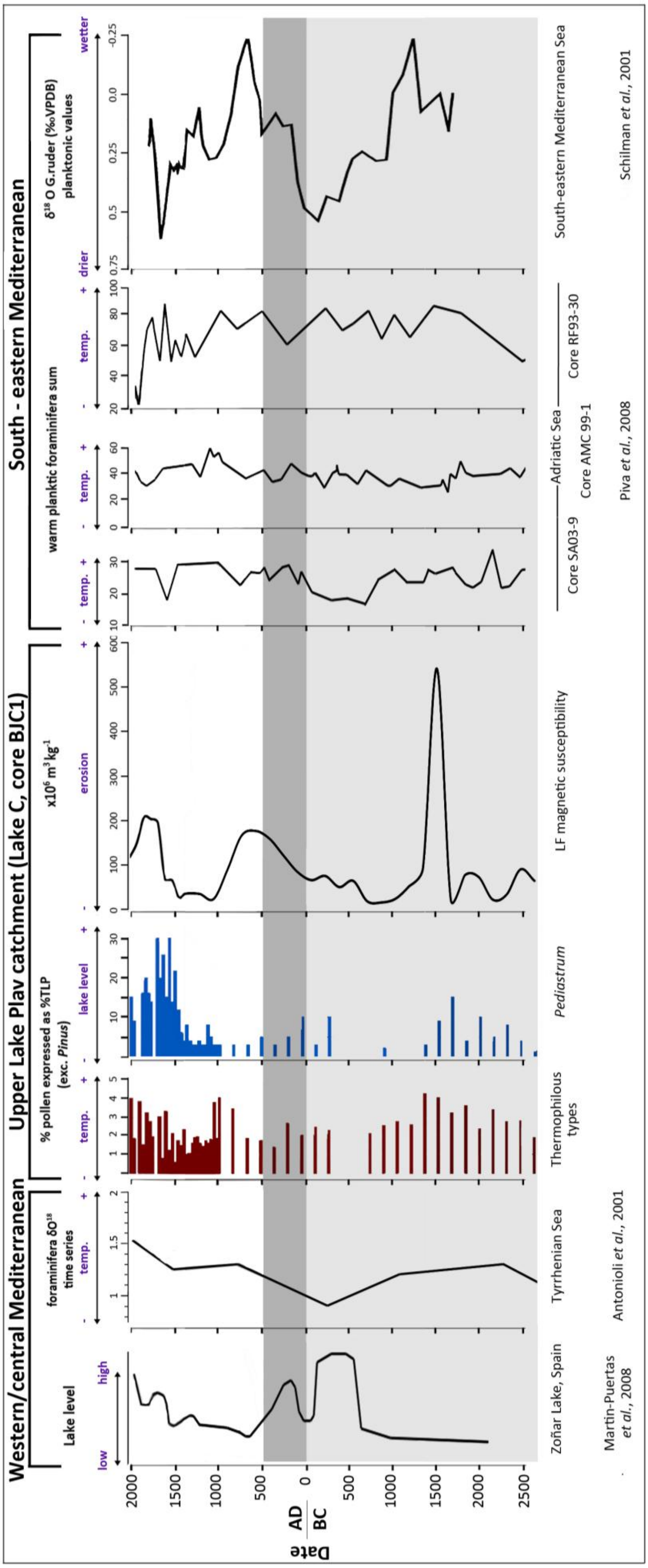


Figure 6.21: Palaeoclimatic and environmental records from Lake C, Buni i Jezerces, compared to climatic records from around the Mediterranean.

particularly dry conditions were seen across the South-Eastern and Eastern Mediterranean area (Schilman *et al.*, 2001; Gat and Magartiz, 1980), coeval to the disappearance of *Pediastrum* and a period of non-polleniferous samples at Lake C, ***c.*** 840 – 350 BC. Conversely Western Mediterranean records indicate increasing precipitation from ***c.*** 600 BC (Figures 6.21, 6.22 Holzhauser *et al.,* 2005 record). Whilst conditions across the central and Eastern Mediterranean at this time are thought to have been cooler, as indicated by the Adriatic Sea records (Piva *et al.*, 2008). Therefore, the likely cause of non-polleniferous samples at Lake C was seasonal evaporation of lake waters under drier and cooler conditions, with pollen oxidising on the dry lake bed.

Between 350 and 0 BC the Lake C thermophilous types indicate a rising temperature trend, as seen in the Adriatic Sea data (cores SA03-9, AMC 99-1). In the Western Mediterranean, elevated lake levels and precipitation persisted at Zoñar Lake, Spain until ***c.*** 150 BC (Martin-Puertas *et al.*, 2008, 2010; Figures 6.21, 6.22). The *Pediastrum* record from Lake C indicates higher lake levels at ***c.*** 350 BC, but any synchrony with the Spanish records are then lost. A rapid drop in *Pediastrum* soon after suggests drier conditions and a similar trend can be inferred by the rapid rise in $\delta^{18}O$ values from South-Eastern Mediterranean Sea, which indicates increasingly dry conditions (Schilman *et al.*, 2001; see Figure 6.21). Records from around the Mediterranean suggest that drier conditions prevailed across the region until AD/BC 0. At approximately this time, the Lake C records then become disengaged with those of the Western/Central Mediterranean, where rising aridity and falling temperatures characterise the climate. In the Eastern Mediterranean region warmer, more humid conditions dominated and these are evident in the rising LF magnetic susceptibility and *Pediastrum* levels at Lake C (see Figure 6.21).

Temperature and lake level proxies suggest cooler and drier conditions ***c.*** AD 0 – 500 at Lake C. This is in general agreement with all but the South-Eastern Mediterranean records, where generally more humid conditions dominated (see Figure 6.21). Falling lake levels characterised Iberian records until ***c.*** AD 500 (Martin-Puertas *et al.*, 2008), in contrast *Pediastrum* levels remain relatively stable suggesting little change in lake level, but inferred erosion into Lake C rises. The Lake C and Piva *et al.* (2008) records indicate rising temperatures for the majority of this period and therefore the stability in *Pediastrum* may be a result of greater evaporation causing little variation in lake level. The rising humidity seen in the Schilman *et al.* (2001) curves are thought to be captured by the Lake C (BJC1) magnetic susceptibility record (see Figure 6.22). Humid conditions persisted in the Eastern Mediterranean records until ***c.*** AD 750 (see Figures 6.21, 6.22), around 400 years longer than the Western Mediterranean. Lake levels in West-Central Europe began to rise around 0 BC (Holzhauser *et al.*, 2005; Figure 6.22) with above average lake levels from AD 100 – 300 (AD 0 – 250 in the pre-Alps, Magny, 2004). This was short lived, unlike the continual rise in humidity in the South Eastern Mediterranean (see Figure 6.21, Schilman *et al.*, 2001).

Overall, the Lake C record is broadly synchronous to Eastern and South-Eastern Mediterranean records, and there is evidence to suggest that between ***c.*** 500 and 0 BC the warming and more humid climate that dominated the Mediterranean also affected the Lake Plav catchment.

6.5.2 The MWP, *c.* AD 500 – 1090

This period is captured in Figures 6.21 and 6.22, with the latter bringing together the lower and upper catchment records. Thermophilous types across the catchment suggest warm conditions and this is generally in agreement with Eastern Mediterranean records (Figure 6.21; Schilman *et al.,* 2002). Reconstructions from the Western Mediterranean and central Europe suggest that more humid conditions did not occur until ***c.*** AD 650 – 700 (see Figure 6.22), but in the South-Eastern Mediterranean rising humidity was inferred until this time, with the MWP suggested to be the final humid period in this region (Figure 6.21; Schilman *et al.* 2001, 2002). The Lake Plav record does show some evidence for decreasing humidity as the magnetic susceptibility record begins to decline by ***c.*** AD 750, even though *Pediastrum* levels remain relatively stable (Figures 6.21, 6.22). Therefore, the Mediterranean region seems to be experiencing different degrees of

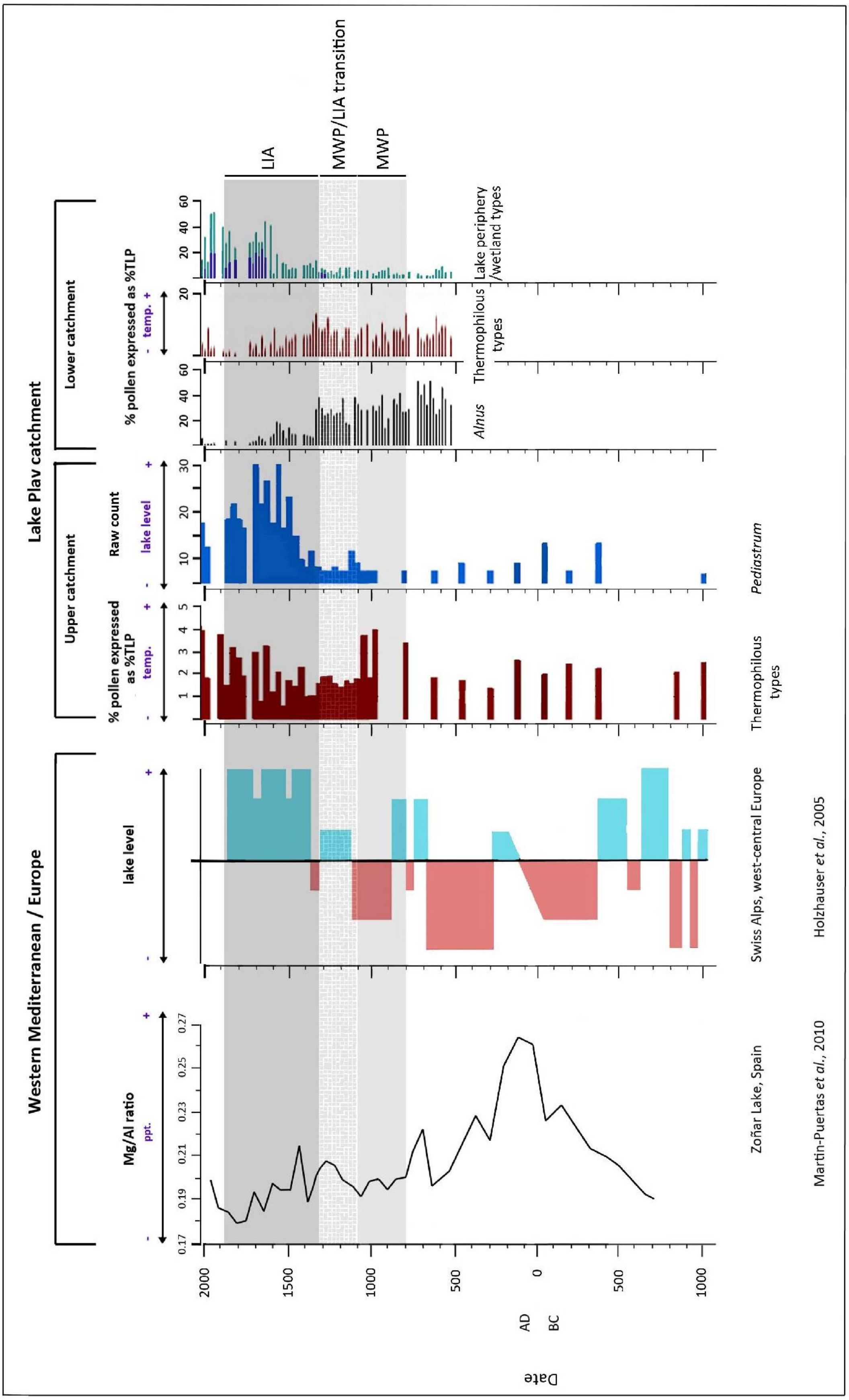


Figure 6.22: Climatic and environmental reconstructions from the Lake Plav catchment compared to selected Mediterranean and European palaeoclimatic records.

dryness from ***c.*** AD 700 – 1100 which may be associated with the MWP (beginning ***c.*** AD 800 at the Lake Plav catchment). Overall, the period ***c.*** AD 500 – 1090 seems to be characterised by a warm climate which is in agreement with the foraminifera records of the Adriatic Sea (see Piva *et al.*, 2008; Figure 6.21). Piva *et al.* (2008) did not quantify how much warmer the climate was, but the continued presence of *Globigerinoides sacculifer* (a warm species) indicates a minimum sea temperature of 16^{o}C (Jenkins and Shackleton, 1979). Later, during the LIA *Globigerinoides sacculifer* is absent, as cooler conditions and North Adriatic Dense Water (NAdDW) dominates. Furthermore, the peaks in *G. sacculifer* during the MWP suggest drier conditions, low turbidity of the water column and reduced river runoff (Piva *et al.,* 2008).

6.5.3 The Little Ice Age across Southern Europe and the Mediterranean, *c.* AD 1090 – 1880

The MWP/LIA transition began to take hold at the Lake Plav catchment ***c.*** AD 1090. At this time *Pediastrum* levels rose, as lake levels in Europe also began to rise (Figure 6.22, Holzhauser *et al.*, 2005). Around AD 1100 a brief period of below average temperatures were recorded in the Calimani Mountains, Romania 250 km north-east of Lake Plav (Figure 6.23; Popa and Kern, 2009). At Lake C, evidence for climatic deterioration was primarily registered in the thermophilous pollen record. Coeval to this, the RF93-30 record from the Adriatic Sea (Figure 6.21) indicates falling temperatures, but in other Mediterranean and European records (Figures 6.21 – 6.23) temperature reconstructions register less obvious change. Therefore, contrasting climate drivers seem to have been affecting the Eastern Mediterranean compared to the rest of the Mediterranean and Europe. This summary is supported by dendrochronological data from the Calimani Mountains and North-Eastern Italy where cool temperatures, in excess of 0.15^{o}C below the 1961 – 1990 mean, were reconstructed between AD 1100 - 1375 (Serre-Bachet, 1994; Popa and Kern, 2009). Given the similarities between the Italian and Romanian records to the Lake Plav catchment data, similar atmospheric circulation patterns may have dominated the region at this time, for example the North-Sea Caspian Pattern (NCP).

Overall, the Lake Plav catchment seems to have entered the LIA sooner than other Mediterranean sites, seemingly a product of its geographical position. Eastern European records indicate climatic deterioration around a century before Mediterranean/Southern European records, with the Lake Plav catchment synchronous with these Eastern European records. Lake levels in central Europe had risen by the mid 1300s, whilst in Spain precipitation was rising, but to a lesser extent than in earlier periods (see Figure 6.22 and Magny, 2004). Around AD 1300 – 1350 the abrupt *Alnus* decline occurred and inputs to Lake Plav became coarser, with sands dominating the sediments in core LPCB and Cyprididae types composing 32% of the ostracoda assemblage (Table 5.10, Chapter 5). In the upper catchment a possible lake level rise ***c.*** AD 1350 – 1360 is inferred by a brief rise in *Pediastrum*. Across Southern Europe and the South-Eastern Mediterranean, conditions were becoming drier following a brief humid period in the mid 1350s (Figure 6.22; Magny, 2004), which corresponds to similar variations at the Lake Plav catchment. Although precipitation may be influenced by local variations, there does seem to be some correlation between the Lake Plav data and that of both Western and Eastern Mediterranean sites during the 1300s. The Wolf minimum, a solar phenomenon that occurred ***c.*** AD 1280 – 1342 was registered in climate records across Europe as temperature deteriorations and elevated precipitation (e.g. Stuiver and Braziunas, 1995; Mauquoy *et al.*, 2002). At the Lake Plav catchment thermophilous types and anthropogenic indicators declined briefly ***c.*** AD 1330 as *Pediastrum* levels peaked, possibly linked to the Wolf minimum (Figure 6.23).

As discussed in Chapter 2, evidence from around Europe has highlighted the heterogeneous nature of the LIA. This is exemplified by the continual variations in thermophilous types, *Pediastrum* and total sands records. Around AD 1400 minor temperature amelioration is implied by an increase in thermophilous types at Lake C. Any similar trends are not mentioned in Western Mediterranean or European records and Piva *et al.* (2008) actually date the beginning of the LIA to ***c.*** AD 1400 with the disappearance of the warm planktic type, *Globigerinoides sacculifer*. Dendrochronological records from North-Eastern Italy and Calimani Mountains indicate the dominance of cold conditions until AD 1600 (Serre-Bachet, 1994; Popa and Kern, 2009; Figures 6.21, 6.23), interspersed with warmer conditions AD 1375 – 1430 in North Eastern Italy (Serre-Bachet, 1994) and ***c.*** AD 1480 in the Calimani Mountains (Popa and Kern, 2009). The Lake Plav data indicate that warmer conditions prevailed after similar conditions in North-Eastern Italy, but before these conditions were recorded in the Calimani Mountains (Serre-Bachet, 1994; Popa and Kern, 2009). It is difficult to determine which record the upper catchment data are most appropriately correlated with, as dendrochronological studies have the advantage of being annual records. The current data suggest that warm conditions spread from North-Eastern Italy towards Lake Plav, before reaching the Calimani Mountains. Further chronological data would be required to be certain about whether the Lake Plav records track the extension of warm conditions across the Eastern Mediterranean and Eastern Europe, or are synchronous with the North Eastern Italy or Calimani Mountains datasets.

By ***c.*** AD 1400 peaks in reconstructed precipitation at Zoñar Lake in Spain (Martin-Puertas *et al.*, 2010), coincided with higher lake levels across the European Alps (Magny, 2004; Holzhauser *et al.*, 2005). Similar changes are noted in the Lake Plav area, as thermophilous types decline in the lower catchment and an inferred lake level rise occurs at Lake C. Following Cook *et al.* (2002) the winter NAO was in a negative state until ***c.*** AD 1450 which, following Xoplaki *et al.* (2001), creates cool and wet conditions over the South/Eastern Mediterranean, inferred within the Lake Plav catchment records (see Figure 6.23). A similar cooling is noted in the Adriatic Sea records, and may be linked to the Spörer solar minimum ***c.*** AD 1460 – 1550 (e.g. Rigozo *et al.*, 2001; Wagner and Zorita, 2005). Temperature downturns are apparent in both the Lake Plav data presented here and in the Calimani Mountain data with increased precipitation inferred during the early 1500s, potentially also a characteristic of solar minima.

Cool conditions are implied at the Lake Plav catchment during the 1500s, a trend which is apparent in the Adriatic Sea records (Piva *et al.*, 2008) and Calimani Mountains (Popa and Kern, 2009). During this period the Spörer Minimum continued. These data contrast with the positive winter NAO data ***c.*** AD 1560 – 1600 (Cook *et al.*, 2002), which would be expected to cause a warmer and drier climate in the Southern and Eastern Mediterranean

(Xoplaki *et al.*, 2001), an expectation apparent in the Schilman *et al.* (2001) record (Figure 6.21). Alpine dendrochronological reconstructions show that AD 1470 – 1570 was warm (Table 2.2, Chapter 2), unlike the dendrochronological reconstructions from the Calimani Mountains (Popa and Kern, 2009). Furthermore, evidence for greater fluvial inputs and humidity throughout the Eastern Mediterranean suggest that these areas are not responding to the positive NAO index, as converse cool humid conditions dominate (see Figure 6.23; Jex *et al.*, *in press*). Throughout the 1500s, climatic reconstructions from Lake Plav are disassociated with those of the Western Mediterranean and Southern Europe and are more in tune with the Eastern/South-Eastern Mediterranean records (Figure 6.23). This is exemplified by the contrast between the Calimani Mountains and Alpine records for this time, with the Plav datasets more associated with the former.

During the early 1600s the reconstructed estimates of the NAO shift from a negative phase ***c.*** AD 1600 – 1620, to positive ***c.*** AD 1620 – 1640 before reverting back to negative for the remainder of the 1600s (Cook *et al.*, 2002). Again the Lake Plav data contradicts the expected effect of the NAO in the area, with rising levels of coarse grained material, *Pediastrum* and the final occurrence of ostracoda. Candonidae types dominated the ostracod assemblage indicating that LPCA site remained part of a permanent water body (Tables 5.9, 5.10, Chapter 5). Data from the Calimani Mountains and North-East Turkey (Gümüşhane province) in the early 1600s indicated cooler temperatures and rising levels of precipitation, similar to some of the Lake Plav records (Popa and Kern, 2009; Jex *et al.*, *in press*; Figure 6.23). At this time *Pediastrum* abundance, LF magnetic susceptibility and thermophilous types rise in the Lake C record whilst, coarse grained material in core LPCB rise and thermophilous types of core LPCA are low but variable (see Figure 6.23). Therefore, local controls upon the Lake Plav catchment - such as geographical location or local topographic variations, may have been dominating climate proxies at this time, causing the conflict within the catchment records. Around AD 1620 *Typha latifolia* multiplied and erosion rates in the upper catchment rose, both a likely product of wetter conditions as discussed previously in section 6.4.2. The North Sea-Caspian Pattern (NCP) regional circulation pattern is thought to be one of the primary controls of rainfall across Turkey (Jones *et al.*, 2006; Jex *et al.*, *in press*). The synchrony between the Lake Plav and

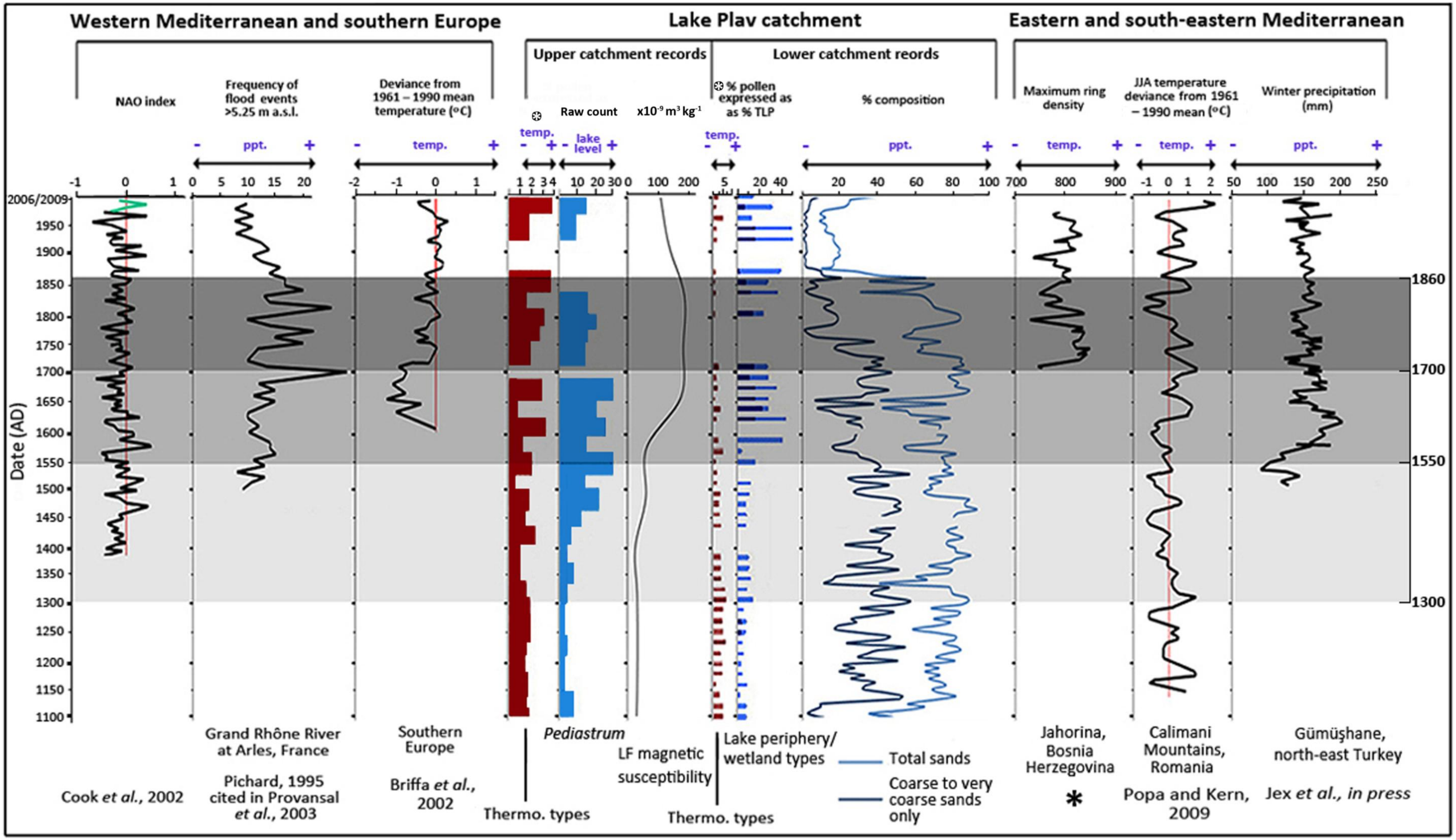


Figure 6.23: Selected Lake Plav catchment records compared to datasets from the Western Mediterranean/Southern Europe and the Eastern Mediterranean /Eastern Europe. * denotes Schweingruber, cited in NOAA (2008) and 'Thermo. types' indicates thermophilous type (see Table 6.4).

Gümüşhane data, coupled with the apparent contrast to Western Mediterranean and Southern European records suggests that circulation patterns such as the NCP or East Atlantic–West Russia pattern (EA–WR; Jex *et al.*, *in press*) were more influential upon the Prokletije climate through the 1600s. During the latter stages of the 1600s the records inferred from the pollen data have greater similarities to that of the Calimani Mountains than the Southern European record presented in Figure 6.23 (Briffa *et al.*, 2001). Popa and Kern (2009) highlighted a period of 'above-average [temperature] values' that lasted for just over 100 years, between AD 1630 – 1740. The Lake C thermophilous types indicate that the 1600s were on average relatively warm for the LIA, but for around 30 years during the mid 1600s cool conditions were noted at the site (Figure 6.23). This corresponds to changes in the NAO index, Central European and Portuguese sites where cooler summers and winters, respectively were recorded (Alcoforado *et al.*, 2000). These changes have been linked to reduced solar irradiance during the Maunder Minimum (Eddy, 1976; Lockwood, 2001; Bradley and Jones, 1993; Holzhauser *et al.*, 2005). The evidence of the Maunder Minimum within the Lake C dataset provides the first evidence in the eastern Mediterranean, as Xoplaki *et al*. (2001) mentioned only the Late Maunder Minimum (LMM; AD 1675 - 1715).

The 1700s were characterised by increasing similarities between the Western and Eastern Mediterranean records (Figure 6.23). Evidence for increased flood frequency at the Grand Rhône (Pichard, 1995 cited in Provansal *et al.*, 2003) and more humid conditions across Europe, Turkey and the South-Eastern Mediterranean (Schilman *et al.*, 2001; Jex *et al.*, *in press*) are presented in Figures 6.22 and 6.23 (see Magny, 2004; Holzhauser *et al.*, 2005). Increased freshwater inputs were evident at Lake Butrint in Albania from AD 1741 as $\delta 18$O (carbonate) values decreased (Ariztegui *et al.*, 2010). The wetter conditions seem to have affected the whole Mediterranean at this time and the Prokletije Mountains were no exception with peaks in *Pediastrum*, lake periphery/wetland types and coarse to very coarse sands all indicate rising precipitation. Temperature amelioration in the 1700s characterised Southern European and Eastern Mediterranean records; as seen in the Lake Plav data and dendrochronological reconstructions for Southern Europe (Briffa *et al.*, 2001) and Jahorina in Bosnia Herzegovina (Schweingruber, cited in NOAA, 2008; Figure 6.23). The particle size analysis of core LPCB showed a fall in coarse sand to almost zero (***c.*** AD 1775), indicative of drier and/or less turbulent hydrological conditions, supported by lower levels of *Pediastrum* and rising thermophilous types in the catchment. This trend replicated across Eastern Mediterranean datasets (Schweingruber cited in NOAA, 2008; Popa and Kern, 2009; Jex *et al.*, *in press*; Figure 6.23). Although precipitation seems to have fallen, erosion rates in the upper catchment remained stable throughout the 1700s. This suggests that erosion conditions were maintained from ***c.*** AD 1650, possibly as a result of anthropogenic activity around Lake C (see Figure 6.15). Evidence for pastoral farming expansion is noted in the core LPCA records at this time (see Figure 6.9), alongside possible clearance events as total arboreal types decline in both the Lake C and core LPCA records (see Figures 6.13; 5.25a,b, 5.32a,b, Chapter 5).

Throughout the early to mid 1800s, the Grand Rhône flood frequencies record indicate the second highest flood frequency between AD 1815 – 1820. This coincided with Southern European temperatures deviating from the 1961 – 1990 mean by 0^0C to -0.5^0C and a brief period of positive state NAO (see Figure 6.23). Similar conditions i.e. cooler and wetter are thought to have occurred in the Lake Plav catchment and Eastern Mediterranean, but to a lesser extent. At Lake C *Pediastrum* indicates a peak in lake level around ***c.*** AD 1800 followed by a decline to ***c.*** AD 1850, while thermophilous types infer a similar trend in temperature. Falling precipitation and temperatures are well represented in the particle size data, but also in the Jahorina, Bosnia Herzegovina maximum ring density and temperature record of the Calimani Mountains in Romania (see Figure 6.23). The Tambora eruption in AD 1815 has been widely acknowledged as the cause for climatic deterioration with falling temperatures and glacial expansion characterising European and Mediterranean records at this time (e.g. Grove, 1988; Büntgen *et al.*, 2005). Although, the effects of the Tambora eruption are thought to have had the greatest effect in AD 1816, the Lake Plav data fit well considering the chronologies applied.

Cooler conditions were then experienced across the Mediterranean, with precipitation at Gümüşhane Province in North-East Turkey increasing slightly. These changes are not only visible in the palaeoenvironmental records of the research catchment, but also at the Maja e Koljaet glacier where the glacier expanded vertically until AD 1834 (see section

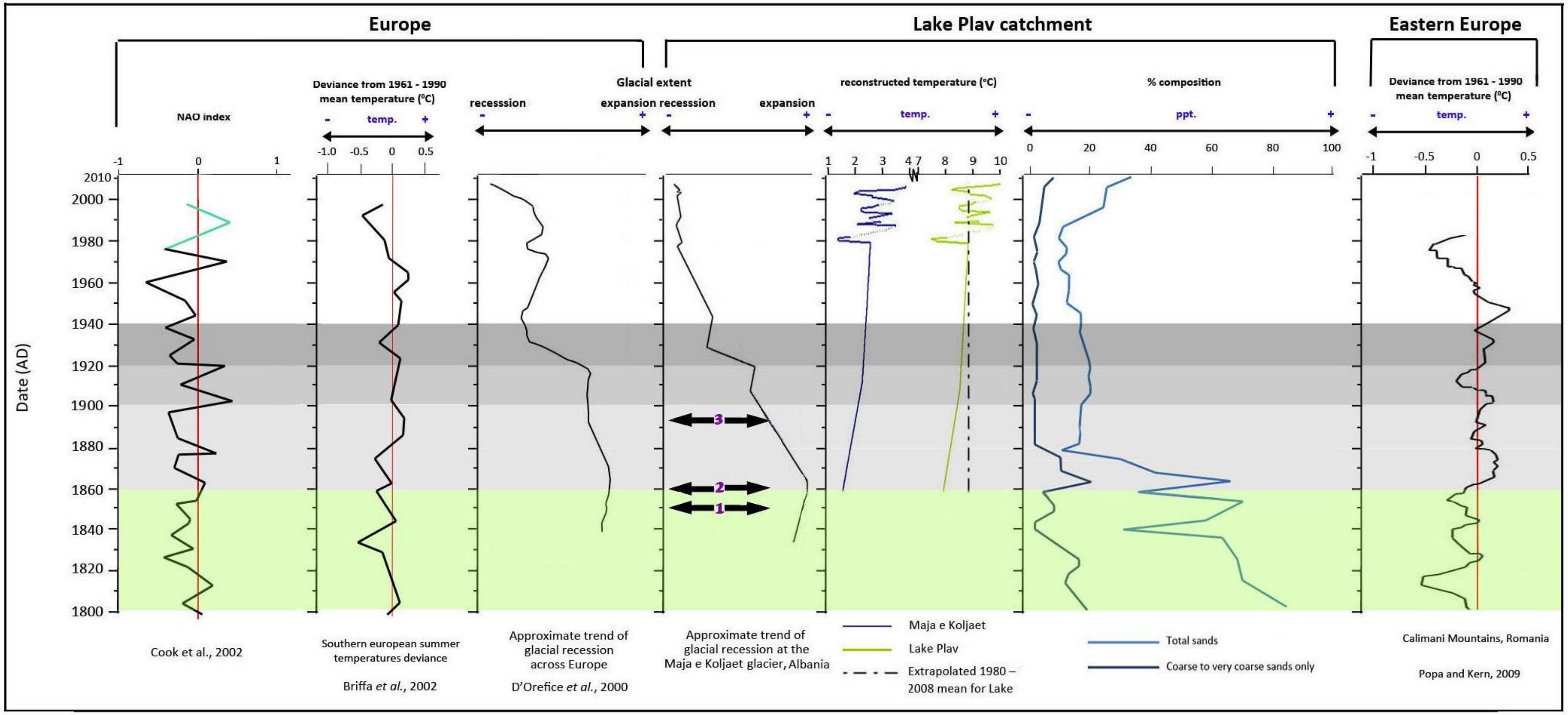


Figure 6.24: The end of the LIA throughout the Mediterranean, comparison of changes in the Lake Plav catchment to the Mediterranean region. Numbers 1 – 3 indicate various LIA glacial maxima; (1) European LIA_{GM} (D'Orefice *et al.*, 2000), (2) Maja e Koljaet LIA_{GM} (this project) and (3) Montenegrin LIA_{GM} (Hughes, 2009b).

6.3.1). After AD 1834 the Maja e Koljaet glacier is thought to have become thinner exposing the AD 1834 lateral moraine, before the glacier reached its greatest extent AD 1859 when it began to retreat from its LIA_{GM} position (see Figures 6.17, 5.36, 5.37 Chapter 5). The LIA_{GM} of the Western Mediterranean and Southern Europe occurred around AD 1850, nearly a decade before that of Maja e Koljaet (see Figure 6.24). Palaeotemperature reconstructions for the glacial record at AD 1859 (Albanian LIA_{GM}) indicate that temperatures at Maja e Koljaet would have been 0.9 ^{o}C below the present day mean (see Figure 6.23, Table 6.6). This is similar to previous reconstructed LIA thermal declines described in section 2.1, Chapter 2 (Matthes, 1939; Grove, 1979, 1988; Lamb, 1995; Gonzalez-Trueba *et al.*, 2008). Across the Eastern Mediterranean region, rising temperatures from ***c.*** AD 1850 have been reconstructed creating a possible catalyst for the glacial retreat that began around AD 1959 at Maja e Koljaet (see Figure 6.23; Serre-Bachet, 1994). Thermophilous pollen types in the upper catchment show continuing change around the time of the LIA_{GM} and the pollen data suggest rising temperatures. The apparent lag in glacial retreat to the changes in the palynological record are thought to be the result of reconstructed chronological differences, with the glacial chronology perhaps most reliable at this time (see Figure 6.24). Overall, the end of the LIA in glacial terms occurred at AD 1859 following temperature amelioration across the Eastern Mediterranean and Romania (e.g. Serre-Bachet, 1994; Popa and Kern, 2009). The climatic amelioration affecting the Eastern Mediterranean is also apparent in Western and Southern European records indicating similar climatic changes across the Mediterranean region by ***c.*** AD 1860. Temperatures around the Mediterranean were rising by ***c.*** AD 1860, whilst precipitation declined (see Pichard, 1995 cited in Provansal *et al.*, 2003; Jex *et al.*, *in press*). This could explain the lack of *Pediastrum* at Lake C at this time, and the decreasing levels of sands in core LPCB.

6.5.4 Palaeoenvironmental changes following the LIA and the impacts of 20th and 21st century changes (1880 – present day)

Following the Albanian LIA_{GM} a shift to finer sediments in core LPCB is one of the significant changes associated with the end of the LIA. A possible diversion of the River

Ljuča course shut down the supply of sediment to site LPCB ***c.*** AD 1880, coinciding with above mean temperatures in the Eastern Mediterranean and rising temperatures in the Western Mediterranean (Figure 6.24). Dry and warm conditions in the Eastern Mediterranean have been linked to either a positive (+) NAO which existed in the late 1800s (Xoplaki *et al.*, 2001; Cook *et al.*, 2002) or a negative (-) NCP (Kutiel and Benaroach, 2002). Precipitation levels were suggested to be more greatly impacted by local factors where the NCP dominates (Kutiel *et al.*, 2002), and thus the dry conditions may not necessarily be the result of - NCP (Figure 6.24). Records of NCP states to the late 1800s are not available, making it difficult to determine the dominant force affecting the Lake Plav catchment. Warming throughout the remainder of the 1800s is apparent throughout the Lake Plav and wider Mediterranean records, with temperatures above the mean for Southern Europe and around the mean in the Calimani Mountains (see Figure 6.24). Dendrochronological records from North-Eastern Italy and the European network indicate that AD 1860 – 1950 was generally warm (e.g. Schweingruber *et al.*, 1988, 1991; Serre-Bachet, 1994; Luterbacher *et al.*, 2004) and glacial recession was apparent in Europe (D'Orefice *et al.*, 2000; see sections 6.3 and 2.3, Chapter 2). At this time the Maja e Koljaet glacier had greater synchrony with European glacial recession trends, as glaciers were generally receding across Europe ***c.*** AD 1878 – 1890, but in Montenegro glaciers were reaching their LIA_{GM} maximum extent (Hughes, 2007; Figure 6.24).

By the 1900s greater synchrony than previously seen with Western Mediterranean and Southern European records is apparent in the Lake Plav catchment records. The negative NAO index ***c.*** AD 1905 – 1915 coincided with a brief period of glacial expansion in European, Apennine and Maja e Koljaet records, with cool temperatures registered in the Adriatic Sea (Figure 6.21) and across European and Eastern Mediterranean dendrochronological records (e.g. Briffa *et al.*, 1988; Luterbacher *et al.*, 2004; Popa and Kern, 2009). Such conditions would be conducive to glacial expansion, with a cooler, wetter climate prevailing, corroborated by the -0.3^{0}C palaeotemperature reconstruction for Maja e Koljaet at AD 1908 (see Table 6.6). Around this time Cvijić (1913) mapped the extent of Lake Plav, showing a much larger lake than its current extent (Figure 6.1). *Typha latifolia* and other lake periphery/wetland types continued to thrive at this time, supporting the suggestion of a wetter climate. Records from Zoñar Lake also indicate rising levels of precipitation at the turn of the century, supporting the view that the NAO was affecting most of the Mediterranean region including the Lake Plav catchment.

Two further glacial expansions were reconstructed for the Maja e Koljaet glacier with moraines dated to ***c.*** 1928/44 and ***c.*** AD 1979 – 1993. The first moraine was formed coeval to a sustained period of - NAO and glacial expansion across Europe (Figure 6.24). This offers the first indication that the catchment was becoming disengaged with climatic changes in Eastern Europe, as particularly warm temperatures were reconstructed at the Calimani Mountains between AD 1920 – 1950 (Popa and Kern, 2009). Precipitation fell between AD 1913 - 1940 as Lake Plav decreased in area by almost 42%. This is corroborated by a reduction in *Pediastrum*, but not in lake periphery/wetland types which only begin to decline ***c.*** AD 1950. The reduced extent of Lake Plav may have resulted from falling precipitation causing the shallowest areas of the site to become terrestrial. The lake periphery/wetland types that enjoyed inundated conditions seem to have been able to withstand terrestrialisation until ***c.*** AD 1950, perhaps maintained by over-bank flooding. The final Maja e Koljaet glacier expansions (AD 1979 – 1993) were coeval to European glacial expansion prior to glacial retreat during the mid 1980s. The NAO was once more in a negative state at the time of the 1970s and 1980s expansions, with the Mediterranean and Eastern Europe records implying cooler temperatures (see Figures 6.23, 6.24). By the 1990s the NAO returned to a positive state, but Southern European and Jahorina temperatures remained below average (Figure 6.24), which may explain how the Maja e Koljaet glacier continued to extend during the early 1990s. Warming since the early 1990s is captured throughout European glacial records and the Lake Plav catchment with rising thermophilous pollen percentages.

The Lake Plav catchment saw one of the earliest initiations of the LIA in the Mediterranean ***c.*** AD 1300. Climatic changes at the catchment are seemingly synchronised with those affecting the Calimani Mountains until the Albanian LIA_{GM} at AD 1859. After this event, greater synchrony to changes in the Western and central Mediterranean were evident. This provides some indication that the Lake Plav catchment has become disengaged with other records from the Eastern Mediterranean region. Climatic changes throughout the LIA suggest that the Lake Plav catchment is susceptible to change as a result of climatic fluctuations, possibly driven by major solar fluctuations and NAO and NCP variations.

7.1 INTRODUCTION

The aims of this research were to reconstruct late Holocene environmental changes at a new site in the Eastern Mediterranean region and investigate the effects of climatic change and land use upon the landscape, with a particular focus on the LIA. To achieve these aims, the mountainous Lake Plav catchment in the Prokletije Mountains was chosen. Upper and lower catchment sites were then used to provide the various records needed to reconstruct environmental change in the catchment; a catchment vegetation survey, two sediment cores from the lower catchment and an upper catchment sediment core, and glacial reconstruction of the Maja e Koljaet glacier. As the research progressed, it became apparent that the sediment cores offered longer records than expected, with the upper catchment record extending to around 2720 BC. This was used to the advantage of the project by establishing the environmental changes that had occurred prior to the LIA, allowing comparison to changes during the LIA.

This chapter aims to provide a summary of the main research findings and to discuss future research.

7.2 OBJECTIVES OF THE RESEARCH

The project objectives outlined in Chapter 1 were:

1) Reconstructing vegetation and environmental changes from sediment cores, using a variety of proxies to capture local and catchment-scale changes.

2) Determining LIA glacial extent and subsequent glacial fluctuations to provide a catchment specific temperature record through to the present.

3) Examining the palaeoenvironmental changes at a high altitude site where climatic forcing is likely to be amplified and anthropogenic factors less influential.

4) Synthesising the data collected to provide an integrated climate, land use and sedimentary history for the Lake Plav catchment.

5) Analysing this integrated record in the wider context of LIA environmental change.

The aim and objectives of this project have generally been successfully achieved. Records from around the catchment have provided information on the changing climate throughout the late Holocene. The non-polleniferous and ostracoda free levels in core LPCA provided indication of flooding events. At the site of core LPCB it was thought at the time of sampling that a central lake record, ideal for palynology, would be retrieved. This was not the case, as the sediments were largely non-polleniferous and instead a detailed, high resolution sedimentary record was obtained. This provided additional information on inferred precipitation levels and fluvial sediment inputs to the lake from ***c.*** AD 1100. In the upper catchment a longer record was retrieved providing greater context for the LIA changes that were identified in all three sediment records. Climatic variations dominate the records, with reconstruction of anthropogenic activity less apparent than expected. Pastoral and ruderal types dominated the anthropogenic indicators, with the most diverse arable indicators found during the MWP (***c.*** AD 800 – 1090).

The chronology created for each of the records produced (LPCA, LPCB, BJC1, Maja e Koljaet glacier) had variable accuracy and precision. Dating of the Lake Plav peripheral core LPCB and the geomorphic features around Maja e Koljaet glacier were most successful, and thus these proxies have the most robust chronologies. Where dating was not initially successful, further dating methods were implemented to create a more reliable chronology. AMS radiocarbon dating was unsuccessful in core LPCA and therefore, two complementary dating methods (^{210}Pb and ^{137}Cs) were used. For core BJC1 an AMS radiocarbon date provided a point of reference for the record as the mobility of ^{137}Cs in the sediments made this technique questionable.

7.3 SUMMARY OF FINDINGS

Palaeoenvironmental investigations at the Lake Plav catchment revealed a record of vegetation, hydrological, climatic and glacial response to late Holocene climate change.

Anthropogenic activity was evident throughout the catchment palynological records, with agricultural expansion first registered at ***c.*** AD 150. Cereal types were not registered until ***c.*** AD 570, around a century later than other Eastern Mediterranean and Balkan sites (Willis and Bennett, 1995). This may have been the result of the mountain location and tradition in such areas, with local populations concentrating on pastoral farming for a longer period of time. By ***c.*** AD 800 the MWP began prior to the MWP/LIA transitional period between ***c.*** AD 1090 – 1300. The LIA was characterised by a decline in thermophilous types from ***c.*** AD 1300 at Lake C, but arable and thermophilous types declined for a sustained period later in the lower catchment record, ***c.*** AD 1460 - 1530. However, as the upper catchment was used to provide a primarily climate sensitive record, with no cultivated ground in the vicinity and a natural forest edge, this earlier initiation date (***c.*** AD 1300) is preferred. Alternatively, the higher mountain area may have had a genuine climate shift earlier, although the ***c.*** AD 1300 date for LIA onset is corroborated by the coarser inputs of the LPCB sediment record between ***c.*** AD 1330 – 1380. The impact of the LIA across all records within the catchment is not consistent until ***c.*** AD 1520 – 1560, when evidence for increased flooding across the catchment is suggested. In the lower catchment, coarse grained events in the LPCA particle size record, coupled with the continued dominance of coarse grained material in the LPCB record, indicate fluctuating levels of coarse grained fluvial inputs into Lake Plav, thought to be a proxy for precipitation. The upper catchment record indicates rising levels of *Pediastrum* ***c.*** AD 1460, thought to indicate a rise in lake levels at Lake C, Buni i Jezerces, which is now seasonal. LIA deterioration is also thought to have caused the expansion in *Typha latifolia* and lake periphery/wetland types at the site, which was maintained until the mid 1900s.

Throughout the LIA greater synchrony with Eastern European areas i.e. Romanian and Turkish records was apparent than with proxies from the west. This implies that climate drivers in these areas e.g. NCP or EA–WR (Jex *et al.*, in press), were also driving the climate around Lake Plav, rather than the NAO which dominates Western Mediterranean areas (Cook *et al.*, 2002). This did not stop the records highlighting the heterogeneous nature of the LIA with continued variance in thermophilous pollen types, *Pediastrum*, pollen concentrations and particle size records; corroborating the conclusions by many researchers that the LIA was not a period of sustained cooler and wetter conditions (e.g. Jalut *et al.*, 2000; Cronin *et al.*, 2003; Trouet *et al.*, 2009). The glacial maximum for the LIA in the study area has been dated to AD 1859 using lichenometric dating of moraine ridge boulders, providing the first LIA_{GM} reconstruction for Albania. At around this time the links to Eastern European and Turkish records diminish, and a closer correspondence to the NAO and Western Mediterranean area is shown.

The most apparent change in the catchment at the end of the LIA was the rapid change of the Lake Plav extent as a result of infilling and falling lake levels. This infilling may have been the catalyst for the possible River Ljuča avulsion of ***c.*** AD 1880, when the coarse sediment supply was diverted away from LPCA and LPCB towards Visitor Mountain. A similar event is not thought to have occurred in any other Mediterranean record and highlights the dynamism of large flat Mediterranean floodplain areas (e.g. the Lake Plav catchment) which have been generally overlooked in Mediterranean sedimentary research.

7.3.1 Climate drivers

The Lake Plav catchment has been synchronous to different records throughout the late Holocene. Until the LIA_{GM} the greatest comparisons were made to Eastern Mediterranean and Eastern European sites, but after ***c.*** AD 1859 the records become more comparable to changes in the Western/Central Mediterranean and Southern Europe. The dominant climate drivers affecting the Lake Plav catchment are therefore thought to have changed. During periods of synchrony to Eastern Mediterranean and Eastern European sites, the NCP and/or EA–WR are thought to have dominated the area. Then during the latter half of the 19th century to the present day, changes in the NAO seem to be driving changes at the Lake Plav catchment.

Links to a number of solar minima were made throughout the Lake Plav catchment records, making these records briefly comparable to Western Mediterranean and Southern European sites. This was highlighted when the Maunder solar minimum (AD 1645 - 1715) was not identified in the Eastern European, Calimani Mountains dataset (Popa and Kern, 2009), but in the Lake Plav datasets. This may have provided indication of the future changes at the Lake Plav catchment, to a climate dominated by changes in the west rather than the east. Overall, climatic changes were most successfully identified in the Lake C record, but also in the particle size analyses of the lower catchment sites.

7.3.2 Climate change and anthropogenic activity

This project aimed to provide an integrated climate, land use and sedimentary history for the catchment. However, the sedimentary history overwhelmed the anthropogenic signal identified within the lower catchment, especially between ***c.*** AD 1730 – 1930 when low to absent pollen concentrations resulted in limited vegetation reconstruction. The data presented suggest that climate drove changes in anthropogenic activity at the site, especially when comparing the Lake C (climatically driven) record to the LPCA record, thought to provide a record of anthropogenic activity in the lower catchment. A MWP/LIA transitional period (***c.*** AD 1090 - 1300) was identified in the Lake C record by a reduction in thermophilous indicators. Climatic deterioration was not apparent in the lower catchment until ***c.*** AD 1330 when an abrupt *Alnus* decline occurred, attributed to anthropogenic clearance. The decline coincides with an expansion of pastoral types and reduction of arable types. It is suggested that the population tried to maintain arable farming practices until the LIA climatic deterioration, when they were forced to expand farming activities and thus cleared *Alnus* from the River Ljuča floodplain. More recently, increases in cultural indicators have mirrored the proxies for warming of the climate since the end of the 1800s.

7.4 METHODOLOGICAL CONSIDERATIONS

Throughout this study, the main points of contention are the chronologies established for the sites. As previously discussed, the accuracy of the age-depth records were continually increased by using different and complementary dating techniques. During comparison of the different records a concerted effort to adhere to the chronology created for each record was made. This was done to ensure that the records presented were not those expected or even wanted, but those provided during objective data collection. An alternative approach would have been to 'tune' the palaeoecological records to match known timescales, or correlate the upper and lower catchment records based on the environmental data. This approach would have been permitted if verification of some parts of the chronologies presented was completed, for example if an independent date had been achieved in the lower units of core LPCA or the upper units of core BJC1. Therefore the approach followed was objective, but may be improved in the future by increasing the number of dated points where possible, perhaps using longer sequences for Pb^{210} and Cs^{137} profiling, SCPs and even tephra. It is unlikely that better radiocarbon sequences can be obtained from these sites due to the paucity of datable organic remains at the key depths. Overall, the approach taken during this study has shown the benefits and drawbacks of a multi-proxy approach, but the primary limitation for this research has been the chronology developed.

Pollen and ostracoda preservation were absent from core LPCB and units of core LPCA. This meant that a direct and accurate comparison of records between these two cores was impossible, although absence of these fossils provided additional support for the rapid infilling and dynamism of the lake environment during the late Holocene. For future research, core extraction from the River Ljuča floodplain or the centre of the current lake may aid the recovery of these fossils, possibly at the expense of a 'catchment wide' record if taken from the floodplain.

7.5 FUTURE RESEARCH

A priority for future research into the climatic controls upon the Lake Plav catchment would be to improve on the records shown here by using other proxies. This could build upon the similarities seen in this project between Mediterranean records and Lake Plav catchment data after the LIA_{GM}. Two possible options could be pursued, either the construction of a $\delta^{18}O$ curve if a site with continuous ostracoda valves could be located, or construction of a catchment specific pollen transfer-function. Alternatively, tree-rings from the oldest *Pinus* or *Juniperus* in the upper catchment may be suitable for dendroclimatology, although problems disentangling the primary controls upon tree growth may limit the usefulness of this method. Vegetation in the area seems responsive to both temperature and precipitation change, especially in the upper catchment. The development of a transfer function using modern pollen analogues could be used in conjunction with the vegetation survey data to model vegetation changes in the catchment during climatic ameliorations/deteriorations.

The magnetic susceptibility records provided less information on erosion within lower catchment records than expected. Further assessment of the magnetic properties of these sediments and source areas in the catchment through fingerprinting would provide a record of sediment provenance to indicate changing source areas. Complementary to this would be a record of carbonate composition, as high $CaCO_3$ would be expected to dominate poorly magnetic and fine grained (<63 µm) material. Given the changing climate regime that seems to be affecting the site, further investigation into the sediments at the site may be of importance for future predictions of how this catchment may respond to future climate change. Have sediment sources changed in the past? Do intense rainfall events have a greater impact upon certain portions of the catchment? The answers to such questions could potentially be used to create a model for sediment load predictions.

7.6 FINAL CONCLUSION

'A multi-proxy study of late Holocene environmental change in the Prokletije Mountains, Montenegro and Albania' has provided new information on how a large Eastern Mediterranean mountain catchment responded to late Holocene and LIA climatic change. The MWP was identified at the catchment ***c.*** AD 800 – 1090 and, unlike in other areas, a period of transition to the LIA seems to have occurred at the site identified between AD 1090 – 1300. This provides one of the earliest dates for Mediterranean LIA onset at ***c.*** AD 1300. Not only did the climate change during the LIA, but land use practices became dominated by pastoral farming. The timing of the Albania LIA_{GM} has been identified to AD 1859 from this research, with a reconstructed temperature of 0.9^{o}C below the present day mean. Around this time, the climate became more closely in tune with the region to the west and south, with correlations to other Mediterranean mountain areas, rather than the climate data from areas to the east.

REFERENCES

Abraham, R. (2007) The mountains of Montenegro: A walker's and trekker's guide, Cicerone, Milnthorpe.

Aaby, B. and Berglund, B.E. (1986) *Characterization of peat and lake deposits* In: Berglund, B.E. (ed.) Handbook of Holocene palaeoecology and palaeohydrology, John Wiley and Sons, Chichester.

Aaby, B. and Digerfeldt, G. (1986) *Sampling techniques for lakes and bogs* In: Berglund, B.E. Handbook of Holocene palaeoecology and palaeohydrology, John Wiley and Sons, Chichester.

Aguzzi, M., Amorosi a, A., Colalongo, M.L., Lucchi, M.R., Rossi, V., Sarti, G. and Vaiani, S.C. (2007) 'Late Quaternary climatic evolution of the Arno coastal plain (Western Tuscany, Italy) from subsurface data' *Sedimentary Geology* **202** pp211–229.

Alçiçek, M.C., Ten Veen, J. H. and Özkul, M. (2006) *Neotectonic development of the Çameli Basin, South-western Anatolia, Turkey* In: Robertson, A.H.F. and Mountrakis, D. (eds.) Tectonic development of the Eastern Mediterranean region, Geological Society, London, Special publication **260** pp591-612.

Alcoforado, M.-J., Nunes, M. de F., Garcia, J.C. and Taborda, J.P. (2000) 'Temperature and precipitation reconstruction in Southern Portugal during the late Maunder Minimum (AD 1675 – 1715)' *The Holocene* **10** (3) pp333-340.

Allen, H.D. (2001) Mediterranean Ecogeography, Prentice Hall, Harlow, England.

Allen, H.D. (2003) 'Response of past and present Mediterranean ecosystems to environmental change' *Progress in Physical Geography* **27** pp359-377.

Allen, T. (1997) Particle Size Measurement 5th edition, Chapman and Hall, London.

Alley, R.B. and Ágústsdóttir, A.M. (2005) 'The 8.2k event: cause and consequences of a major Holocene abrupt climate change' *Quaternary Science Reviews* **24** pp1123–1149.

Alonso, F., Nicolás, P. and Martínez de Pisón, E. (1983) 'Los glaciares españoles actuales' INEGLA.

Alonso, V. and González- Suárez, J.J. (1998) 'Presencia de hielo glaciar en los Picos de Europa (Cordillera Cantábrica). El helero del Jou Negro' *Revista Cuaternario and Geomorfologia* **12** pp35-44.

Andersen, S.T. (1970) 'The relative pollen productivity and representation of North European trees, and correction for tree pollen spectra' *Danmarks Geologiske Undersøgelse Series II* **19** pp1-199.

Andersen, S.T. (1974) 'Wind Conditions and Pollen Deposition in a Mixed Deciduous Forest' *Grana* **14** (2) pp57-63.

Andersen, J. L. and Sollid, J. L. (1971) 'Glacial chronology and glacier geomorphology in the marginal zones of the glaciers Midtdalsbreen and Nigardsbreen, South Norway' *Norwegian Journal of Geography* **25** (1) pp1-38.

R e f e r e n c e s

Antonioli, F., Silenzi, S. and Frisia, S. (20010 'Tyrrhenian Holocene palaeoclimate trends from spelean serpulids' *Quaternary Science Reviews* **20** pp1661-1670.

Appleby, P.G. (2001) *Chronostratigraphic techniques in recent sediments* In: Last, W.M. and Smol, J.P. (eds.) Tracking environmental change using lake sediments, Volume 1: Basin analysis coring and chronological techniques, Kluwer Academic, Dordrecht.

Appleby, P.G (2008) 'Three decades of dating recent sediments by fallout radionuclides: a review' *The Holocene* **18** pp83-93.

Appleby, P.G., Oldfield, F., Thompson, R., Huttunen, P. and Tolonen, K. (1979) '^{210}Pb dating of annually laminated lake sediments from Finland' *Nature* **280** pp53-55.

Andreev, A.A., Forman, S.L., Ingólfsson, Ó. and Manley, W.F. (2006) 'Middle Weichselian environments on Western Yamal Peninsula, Kara Sea based on pollen records' *Quaternary Research* **65** pp275-281.

Ariztegui, D., Anselmetti, F.S., Robbiani, J.M., Bernasconi, S.M., Brati, E., Gilli, A. and Lehmann, M.F. (2010) 'Natural and human-induced environmental change in Southern Albania for the last 300 years – Constraints from the Lake Butrint sedimentary record' *Global and Planetary Change* **71** pp183-192.

Armenian, H.K., Noji, E.K. and Oganesian, A.P. (1992) 'A case – control study of injuries arising from the earthquake in Armenia, 1988' *WHO Bulletin OMS* **70** pp251-257.

Armstrong, R. (2004) 'Lichens, lichenometry and global warming' *Microbiologist* pp32-35.

Armstrong, R. and Bradwell, T. (2010) 'Growth of crustose lichens: A review' *Geografiska Annaler: Series A Physical Geography* **92** (1) pp3-17.

Arnaud, F., Magand, O., Chapron, E., Bertrand, S., Boës, X., Charlet, F. and Mélières, M.-A. (2006) 'Radionuclide dating (^{210}Pb, ^{137}Cs, ^{241}Am) of recent lake sediments in a highly active geodynamic setting (Lakes Puyehue and Icalma - Chilean Lake District)' *Science of the Total Environment* **366** pp837-850.

Arnaud-Fassetta, G. (2003) 'River channel changes in the Rhone Delta (France) since the end of the Little Ice Age: geomorphological adjustment to hydroclimatic change and natural resource management' *Catena* **51** pp141-172.

Arnaud-Fassetta, G. and Provansal, M. (1999) 'High frequency variations of water flux and sediment discharge during the Little Ice Age (1586–1725 AD) in the Rhône Delta (Mediterranean France). Relationship to the catchment basin' In: Garnier, J. and Mouchel, J.-M. (eds.) Man and River Systems, *Hydrobiologia* **410** pp241-250.

Atanassova, J. and Stefanova, I (2003) 'Late-glacial vegetational history of Lake Kremensko-5 in the Northern Pirin Mountains, South-western Bulgaria' *Vegetation History and Archaeobotany* **12** pp1-6.

Atherden, M., Hall, J. and Wright, J.C. (1993) 'A pollen diagram from the Northeast Peloponnese, Greece: implications for vegetation history and archaeology' *The Holocene* **3** (4) pp351-356.

R e f e r e n c e s

Auer, B. (2007) 'Pleistocene glaciations in the Prokletije Mountains of Montenegro' unpublished MSc thesis, The University of Manchester.

Ausseil, A.G.E. and Dymond, J.R. (2010) *Evaluating ecosystem services of afforestation on erosion-prone land: A case study in the Manawatu catchment, New Zealand* In: Swayne, D.A., Yang, W., Voinov, A. A., Rizzoli, A. and Filatova, T. (eds.) International Environmental Modelling and Software Society (iEMSs), International Congress on Environmental Modelling and Software Modelling for Environment's Sake, Fifth Biennial Meeting, Ottawa, Canada.

Axford, Y., Geirsdottir, A., Miller, G.H. and Langdon, P.G. (2009) 'Climate of the Little Ice Age and the past 2000 years in Northeast Iceland inferred from chironomids and other lake sediment proxies' *Journal of paleolimnology* **41** (1) pp7-24.

Bahr, D.B., Pfeffer, W.T., Sassolas, C. and Meier, M.F. (1998) 'Response time of glaciers as a function of size and mass balance' *Journal of Geophysical Research* **103** pp9777–9782.

Baillie, M.G.L. (1982) Tree ring dating and archaeology, Croom Helm Ltd., London.

Bakels, C.C. (1991) *Western continental Europe* In: van Zeist, W., Wasylikowa, K. and Behre, K.E. (eds.) Progress in old world palaeoethnobotany, Balkema, Rotterdam.

Baković, M., Mihailović, B., Mihailović, D., Morley, M., Vušović-Lučić, Whallon, R. and Woodward, J. (2009) 'Crvena Stijena excavations 2004 – 2006, preliminary report' *Eurasian Prehistory* **6** (1-2) pp3-31.

Ballantyne, C.K. (2007) *Glacial landforms, ice sheets: Trimlines and Palaeonunataks* In: Elias, S.A. (ed.) Encyclopaedia of Quaternary Science, Elsevier Science, Amsterdam.

Barber, K.E., Maddy, D., Rose, N., Stevenson, A.C., Stoneman, R. and Thompson, R. (2000) 'Replicated proxy-climate signals over the last 2000 yr from two distant UK peat bogs: new evidence for regional palaeoclimate teleconnections' *Quaternary Science Reviews* **19** pp481-487.

Bard, E., Raisbeck, G., Yiou, F. and Jouzel, J. (2000) 'Solar irradiance during the last 1200 years based on cosmogenic nuclides' *Tellus* **52B** pp985-992.

Barriendos-Vallve, M. and Martín Vide, J. (1998) 'Secular climatic oscillations as indicated by catastrophic floods in the Spanish Mediterranean coastal area (14th–19th centuries)' *Climatic change* **38** (4) pp473–91.

Barry, R. G. and Chorley, R.J. (1987) Atmosphere, Weather, and Climate (5th edition), Methuen, London.

Bartlein, P.J., Webb, T. and Fleri, E. (1984) 'Holocene climatic change in the Northern Midwest: Pollen-derived estimates' *Quaternary research* **22** pp361-374.

Basavaiah, N., Juyal, N., Pant, R.K., Yadava, M.G., Singhvi, A.K. and Appel, E. (2004) 'Late Quaternary climate changes reconstructed from mineral magnetic studies from proglacial lake deposits of Higher Central Himalaya' *The Journal of Indian Geophysical Union* **8** (1) pp27-37.

R e f e r e n c e s

Baster, I., Pugin, A. and Wildi, W. (2003) 'High-resolution seismic stratigraphy of an Holocene lacustrine delta in Western Lake Geneva (Switzerland)' *Eclogae Geologicae Helvetiae* **96** Supplement 1, ppS11-S20.

Battarbee, R.W., Jones, V., Flower, R.J., Cameron, N., and Bennion, H. (2001) *Diatoms* In: Smol, J., Birks, H.J.B. and Last, W. (eds.) Tracking environmental change using lake sediments: Terrestrial, algal, and siliceous indicators, Kluwer Academic Publishers, Dordrecht, pp155-202.

Bavec, M., Tulaczyk, S.W., Mahan, S.A. and Stock, G.M. (2004) 'Late Quaternary glaciation of the Upper Soč a River Region (Southern Julian Alps, NW Slovenia)' *Sedimentary Geology* **165** (3-4) pp265-283.

Becker, B. (1993) 'An 11,000-Year German Oak and Pine Dendrochronology for Radiocarbon Calibration' *Radiocarbon* **35** pp201-214.

Behre, K.E. (1981) 'The interpretation of anthropogenic indicators in pollen diagrams' *Pollen et Spores* **23** pp225-245.

Behre, K.E. (ed.; 1986) Anthropogenic indicators in pollen diagrams, A.A. Balkema, Rotterdam.

Belis, C.A., Finsinger, W. and Ammann, B. (2008) 'The late glacial-Holocene transition as inferred from ostracod and pollen records in the Lago Piccolo di Avigliana (Northern Italy)' *Palaeogeography, Palaeoclimatology, Palaeoecology* **264** pp306-317.

Belis, C.A., Lami, A., Guilizzoni, P., Ariztegui, D. and Geiger, W. (1999) 'The late Pleistocene ostracod record of the crater lake sediments from Lago di Albano (Central Italy): Changes in trophic status, water level and climate' *Journal of Paleolimnology* **22** pp151-169.

Bell, M. and Walker, M.J.C. (1992) Late Quaternary Environmental Change: physical and human perspectives, Prentice Hall, London.

Belmecherie, S., von Grafenstain, U., Andersen, N., Eymard-Brdon, A., Régnier, D., Grenier, C. and Lézine, A.-M. (2010) 'Ostracod-based isotope record from Lake Ohrid (Balkan Peninsula) over the last 140 ka' *Quaternary Science Reviews* **29** pp3894-3904.

Bengtsson, L. and Enell, M. (1986) *Chemical analysis* In: Berglund, B. E. (ed.) Handbook of Holocene Palaeoecology and Palaeohydrology. John Wiley and Sons Ltd., Chichester, 423-451.

Beniston, M., Diaz, H.F. and Bradley, R.S. (1997) 'Climatic Change at High Elevation Sites; An overview' *Climate Change* **36** pp233-251.

Benn, D.I. and Evans, D.J.A. (1998) Glaciers and Glaciation, Arnold, London.

Benn, D.I. and Lehmkuhlm, F. (2000) 'Mass balance and equilibrium-line altitudes of glaciers in high-mountain environments' *Quaternary International* **65/66** pp15-29.

Bennett, K.D., Tzedakis, P.C. and Willis, K.J. (1991) 'Quaternary refugia of North European trees' *Journal of Biogeography* **18** pp103-115 1991.

Benoit, G. and Rozan, T.F. (2001) '^{210}Pb and ^{137}Cs dating methods in lakes: a retrospective study' *Journal of Paleolimnology* **25** pp455-465.

R e f e r e n c e s

Benvenuti, M. (2003) 'Facies analysis and tectonic significance of lacustrine fan-deltaic successions in the Pliocene–Pleistocene Mugello Basin, Central Italy' *Sedimentary Geology* **157** pp197-234.

Beraldi, H. (1898) 'Cent ans aux les Pyrenees' 1 Pau Les Amis des Livre Pyrenean (Reprinted 1977).

Berglund, B.E. and Ralska-Jasiewiczowa, M. (1986) *Pollen analysis and pollen diagrams* In: Berglund, B.E. (ed.) Handbook of Holocene palaeoecology and palaeohydrology, John Wiley and Sons, Chichester.

Bergmeier, E. and Dimopoulos, P. (2001) '*Fagus sylvatica* forest vegetation in Greece: Syntaxonomy and gradient analysis' *Journal of Vegetation Science* **12** pp109-126.

Beschel, R.E. (1954) 'Eine Flechte als Niederschlagsmesser' *Wetter und Leben* **6-7** pp56-60.

Beschel, R.E. (1961) *Dating rock surfaces by lichen growth and its application to glaciology and physiography (lichenometry)* In: Raasch, G.O. (ed.) Geology of the arctic (Vol.2), University of Toronto Press, Toronto.

Beta Analytic Inc. 4985 SW 74 Court, Miami, Florida, 33155, USA.

Beug, H.J. (1961) Leitfaden der Pollenbestimmungen, I.Fischer, Stuttgart.
Beuselinck, L., Govers, G., Poesen, J., Degraer, G. and Froyen, L. (1998) 'Grain-size analysis by laser diffractometry: comparison with the sieve-pipette method' *Catena* **32** pp193–208.

Birks, H.H. (2001) *4. Plant Macrofossils* In: J. P. Smol, H. J. B. Birks and W. M. Last (eds.) Tracking Environmental Change Using Lake Sediments, Volume 3: Terrestrial, Algal, and Siliceous Indicators, Kluwer Academic Publishers, Dordrecht.

Birks, H.H. (2003) 'The importance of plant macrofossils in the reconstruction of Late glacial vegetation and climate: examples from Scotland, Western Norway, and Minnesota, USA' *Quaternary Science Reviews* **22** pp453-473.

Birks, H.H. and Ammann, B. (2000) 'Two terrestrial records of rapid climate change during the glacial–Holocene transition (14,000–9000 calendar years BP) from Europe' *Proceedings of the National Academy of Sciences of the United States of America* **97** pp1390-1394.

Birks, H.J.B. and Birks, H.H. (1980) *Quaternary Palaeoecology*, Edward Arnold, London.

Birks, H. B. and Birks, H.J.B. (2001) 'Future uses of pollen analysis must include plant macrofossils' *Journal of Biogeography* **27** pp31-35.

Birks, H.H. and Birks, H.J.B. (2006) 'Multi-proxy studies in paleolimnology' *Vegetation History and Archaeobotany* **15** pp235-251.

Birks, H.J.B. and Line, J.M. (1992) 'The use of rarefaction analysis for estimating palynological richness from Quaternary pollen analytical data' *The Holocene* **2** pp1-10.

Björck, S., Noe-Nygaard, N., Wolin, J., Houmark-Nielsen, M., Hansen, H.J. and Snowball, I. (2000) 'Eemian Lake development, hydrology and climate: a multi-stratigraphic study of the Hollerup site in Denmark' *Quaternary Science Reviews* **19** (6) pp509-536.

References

Björck, S. and Wohlfarth, B. (2001) *^{14}C chronostratigraphic techniques in paleolimnology* In: Last, W.M. and Smol, J.P. (eds.) Tracking environmental change using lake sediments, Volume 1: Basin analysis coring and chronological techniques, Kluwer Academic, Dordrecht.

Björkman, L., Feurdean, A., Cinthio, K., Wohlfarth, B. and Possnert, G. (2002) 'Lateglacial and early Holocene vegetation development in the Gutaiului Mountains, Northwestern Romania' *Quaternary Science Reviews* **21** pp1039-1059.

Björkman, L., Feurdean, A. and Wohlfarth, B. (2003) 'Late-Glacial and Holocene forest dynamics at Steregoiu in the Gutaiului Mountains, Northwest Romania' *Review of Palaeobotany and Palynology* **124** pp79-111.

Blamey, M. and Grey-Wilson, C. (2004) Wild flowers of the Mediterranean, A & C Black, London.

Blondel, J. and Médail, F. (2009) *Biodiversity and conservation* In: Woodward, J.C. (ed.) The physical geography of the Mediterranean, Oxford University Press, Oxford.

Blott, S. J. and Pye, K. (2006) 'Particle size distribution analysis of sand-sized particles by laser diffraction: an experimental investigation of instrument sensitivity and the effects of particle shape' *Sedimentology* **53** pp671–685.

de Boer, G.B.J., de Weerd, Thoenes, D. and Goossens, H.W.J. (1987) 'Laser diffraction spectrometry: Fraunhofer diffraction versus Mie scattering' *Particle Characterisation* **4** pp138-146.

Böhm, R., Auer, I., Brunetti, M., Maugeri, M., Nanni, T. and Schöner, W. (2001) 'Regional temperature variability in the European Alps: 1760–1998 from homogenized instrumental time series' *International Journal of Climatology* **21** pp1779-1801.

Bohncke, S. and Wijmstra, L. (1988) 'Reconstruction of Late-Glacial lake-level fluctuations in The Netherlands based on palaeobotanical analyses, geochemical results and pollen-density data' *Boreas* **17** pp403-425.

Bonaparte, P.R. (1891) 'Les variations périodiques des glaciers Français III' *Annuaire de Club Alpine Français* **18** pp423-427.

Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M.N., Showers, W., Hoffmann, S., Lotti-Bond, R., Hajdas, I. and Bonani, G. (2001) 'Persistent solar influence on North Atlantic climate during the Holocene' *Science* **294** pp2130-2136.

Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., de Menocal, P., Priore, P., Cullen, H., Hajadas, I. and Bonani, G. (1997) 'A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates' *Science* **278** pp1257-1266.

Bonnefille, R. and Mohammed, U. (1994) 'Pollen-inferred climatic fluctuations in Ethiopia during the last 3000 *years' Palaeogeography, Palaeoclimatology, Palaeoecology* **109** pp331-343.

Boomer, I., Horne, D.J. and Slipper, I.J. (2003) *The use of ostracods in palaeoenvironmental studies, or what you can do with an ostracod shell* In: Park, L. E. and Smith, A. J. (eds.) Bridging the gap: trends in Ostracode biological and geological sciences, The Paleontological Society, Papers, **9** pp153–179.

R e f e r e n c e s

Bordon, A., Peyron, O., Lézine, A .-M., Brewer, S. and Fouache, E. (2009) 'Pollen-inferred Late-Glacial and Holocene climate in Southern Balkans (Lake Maliq)' *Quaternary International* **200** pp19-30.

Borisenkov, Y.P. (1994) 'North Atlantic Oscillation: Concept and an application' *Bulletin of American Meteorological Society* **68** pp1218-1225.

Bošković, M. and Bajković, I. (2006) 'Waters of Montenegro, BALWOIS: Water Observation and Information System for Balkan Countries' Paper number A-403, Accessed at: http://balwois.mpl.ird.fr/balwois/administration/full_paper/ffp-870.pdf (4.4.09).

Bottema, S. (1992) 'Prehistoric cereal gathering and farming in the Near East: the pollen evidence' *Review of Palaeobotany and Palynology* **73** pp21-33.

Bottema, S. (2001) 'A note on the pollen representation of ivy (*Hedera helix L.*)' *Review of Palaeobotany and Palynology* **117** pp159-166.

Bozilova, E.D. and Tonkov, S.B. (2000) 'Pollen from Lake Sedmo Rilsko reveals Southeast European postglacial vegetation in the highest mountain area of the Balkans' *New Phytologist* **148** pp315–25.

Bradley, R.S., Hughes, M.K. and Diaz, H.F. (2003) 'Climate in Medieval time' *Science* **302** pp404-405.

Bradley, R.S. and Jones, P.D. (1992) Climate since A.D. 1500, Routeledge, London.

Bradley, R.S. and Jones, P.D. (1993) 'Little Ice Age summer temperature variations: their nature and relevance to recent global warming trends' *Holocene* **3** (4) pp367-376.

Bradshaw, R. and Hannon, G. (1992) 'Climatic change, human influence and disturbance regime in the control of vegetation dynamics within Fiby Forest, Sweden' *Journal of Ecology* **80** (4) pp625-632.

Braithwaite, R.J. (2008) 'Temperature and precipitation climate at the equilibrium-line altitude of glaciers expressed by the degree-day factor for melting snow' *Journal of Glaciology* **54** (186) pp437-444.

Braithwaite, R.J. and Raper, S.C.B. (2007) 'Glaciological conditions in seven contrasting regions estimated with the degree-day model' *Annals of Glaciology* **46** (1) pp297-302.

Braithwaite, R.J., Raper, S.C.B. and Chutko, K. (2006) 'Accumulation at the equilibrium-line altitude of glaciers inferred from a degree-day model and tested against field observations' *Annals of Glaciology* **43** pp329-334.

Bravard, J.-P. (1991) 'La dynamique fluviale á l'épreuve des changements environnementaux: quels enseignements applicables á l'aménagement des rivières?' *La Houille Blanche* **7-8** 515-521.
Bravard, J.-P., Roux, A., Amoros, C. and Reygrobellet, J.L. (1992) *The Rhône river: a large alluvial temperate rive* In: Callow, P. and Petts, G. (eds.) The Rivers Handbook, Volume I, Blackwell, London pp426-447.

R e f e r e n c e s

Bravard, J-P., Vérot-Bourrely, A., Franc, O. and Arlaud, C. (1997) Paléodynamique du site fluvial de Lyon depuis le Tardiglaciaire In: Bravard, J-P. and Prestreau, D. (eds.), Dynamique du paysage Entretiens de géo-archéologie, Documents d'Archéologie en Rhône-Alpes, Lyons, France pp177-201.

Brehme, K. (1951) 'Jahrringchronologische und -klimatologische Untersuchungen an Hochgebirgsliirchen des Berchtesgadener Landes.-Ztschr' *Weltforstwirtschaft* **14** pp65-80.

Breitenlechner, E., Hilber, M., Lutz, J., Kathrein, Y., Unterkircher, A. and Oeggl, K. (2010) 'The impact of mining activities on the environment reflected by pollen, charcoal and geochemical analyses' *Journal of Archaeological Science* **37** pp1458-1467.

Bridgewater, N.D., Heaton, T.H.E. and O'Hara, S.L. (1999) 'A late Holocene palaeolimnological record from central Mexico, based on faunal and stable-isotope analysis of ostracod shells' *Journal of Paleolimnology* **22** pp383-397.

Briffa, K.R., Jones, P.D., Schweingruber, F.H. (1988) 'Summer temperature patterns over Europe: A reconstruction from 1750 A.D. based on maximum latewood density indices of Conifers' *Quaternary Research* **30** pp36-52.

Briffa, K.R., Jones, P.D., Schweingruber, F.H. and Osborn, T.J. (1998) 'Influence of volcanic eruptions on Northern Hemisphere summer temperature over the past 600 years' *Nature* **393** (6684) pp450-455.

Briffa, K.R., Osborn, T.J., Schweingruber, F.H., Harris, I.C., Jones, P.D., Shiyatov, S.G. and Vaganov, F.A. (2001) 'Low frequency temperature variations from a Northern tree-ring density network' *Journal of Geophysical Research* **106** (D3) pp2929-2941.

Briffa, K.R., Osborn, T.J., Schweingruber, F.H., Harris, I.C., Jones, P.D., Shiyatov, S.G. and Vaganov, E.A. (2002) 'Northern Hemisphere and Regional Temperature Reconstructions' IGBP PAGES/World Data Centre-A for Paleoclimatology Data Contribution Series #2002-015 NOAA/NGDC Paleoclimatology Program, Boulder CO, USA, Accessed at: ftp://ftp.ncdc.noaa.gov/pub/data/paleo/treering/reconstructions/n_hem_temp/briffa2001jgr2.txt (05.01.10).

Briffa, K.R. and Schweingruber, F.H. (1995) *Recent dendroclimatic evidence of Northern and central European summer temperatures* In: Bradley, R.S. and Jones P.D. (eds.) Climate since A.D. 1500, Routledge, Chapman and Hall, London.

Broecker, W.S. (2001) 'Was the Medieval Warm Period global?' *Science* **291** (5508) pp1497-1499.

Brugger, K.A. (2006) 'Late Pleistocene climate inferred from the reconstruction of the Taylor River glacier complex, Southern Sawatch Range, Colorado' *Geomorphology* **75** pp318– 329.

Buckland, P.C. and Edwards, K.J. (1984) 'The longevity of pastoral episodes of clearance activity in pollen diagrams: the role of post occupation grazing' *Journal of Biogeography* **11** pp243-249.

Büntgen, U., Esper, J., Frank, D.C., Nicolussi, K. and Schmidhalter, M. (2005) 'A 1052-year tree-ring proxy for Alpine summer temperatures' *Climate Dynamics* **25** pp141-153.

R e f e r e n c e s

Butzer, K.W. (2005) 'Environmental history in the Mediterranean world: cross-disciplinary investigation of cause-and-effect for degradation and soil erosion' *Journal of Archaeological Science* **32** pp1773-1800.

Callaway, J.C., DeLaune, R.D. and Patrick, W.H.(1996) 'Chernobyl ^{137}Cs used to determine sediment accretion rates at selected Northern European coastal wetlands' *Limnology and Oceanography* **41** (3) pp444-450.

Cameron, N.G., Schnell, O.A., Rautio, M.L., Lami, A., Livingstone, D.M., Appleby, P.G., Dearing, J.A. and Rose, N.L. (2002) 'High-resolution analyses of recent sediments from a Norwegian mountain lake and comparison with instrumental records of climate' *Journal of Paleolimnology* **28** (1) pp79-93.

Campbell, C. (1998) 'Late Holocene lake sedimentology and climate change in Southern Alberta' *Canadian Quaternary Research* **49** pp96-101.

Campbell, I.D. (1999) 'Quaternary pollen taphonomy: examples of differential redeposition and differential preservation' *Palaeogeography, Palaeoclimatology, Palaeoecology* **149** pp245-256.

Campbell, I.D. and McAndrews, J.H. (1993) 'Forest disequilibrium caused by rapid Little Ice Age cooling' *Nature* **366** pp336-338.

Camuffo, D. and Enzi, S. (1995) *Reconstructing the climate of Northern Italy from archive sources* In: Bradley, R.S. and Jones P.D. (eds.) Climate since A.D. 1500, Routledge, Chapman and Hall, London.

Carbonnières, L.F.E. (1789) *'Observations fait dans les Pyrenees'* Paris, Belin.

Čarni, A., Košira, P., Karadžić, B., Matevskic, V., Redžić, S. and Škvorc Z. (2009) 'Thermophilous deciduous forests in Southeastern Europe' *Plant Biosystems* **143** (1) pp1-13.

Carrión, J.S. and Navarro (2002) 'Cryptogam spores and other non-pollen microfossils as sources of palaeoecological information: case studies from Spain' *Annales Botanici Fennici* **39** pp1-14.

Carroll, J., Williamson, M., Lerche, I., Karabanov, E. and Williams, D.F. (1999a) 'Geochronology of Lake Baikal from ^{210}Pb and ^{137}Cs radioisotopes' *Applied Radiation and Isotopes* **50** pp1105-1119.

Carroll, J., Lerche, I., Abraham, J.D. and Cisar, D.J. (1999b) 'Sediment ages and flux variations from depth profiles of ^{210}Pb: lake and marine examples' *Applied Radiation and Isotopes* **50** pp793-804.

Cerdà, A. (1998) 'Relationships between climate and soil hydrological and erosional characteristics along climatic gradients in Mediterranean limestone areas' *Geomorphology* **25** pp123-134.

Cerling, T.E. and Craig, H. (1994) 'Geomorphology and in-situ cosmogenic isotopes' *Annual Review of Earth and Planetary Science* **22** pp273-317.

Chambers, F.M., Mighall, T.M. and Keen, D.H. (1996) 'Early Holocene pollen and molluscan records from Enfield Lock, Middlesex, UK' *Proceedings of the Geologists' Association* **107** pp1-14.

R e f e r e n c e s

Chapron, E., Faïn, X., Magand, O., Charlet, L., Debret, M. and Mélières, M.A. (2007a) 'Reconstructing recent environmental changes from proglacial lake sediments in the Western Alps (Lake Blanc Huez, 2543 m a.s.l., Grandes Rousses Massif, France)' *Palaeogeography, Palaeoclimatology, Palaeoecology* **252** (3-4) pp586-600.

Chapron, E., Juvigné, E., Mulsow, S., Aritzegui, D., Magand, O., Betrand, S., Pino, M. and Chapron, O. (2007b) 'Recent clastic sedimentation processes in Lake Puyehue (Chilean Lake District, 40.5°S)' *Sedimentary Geology* **201** pp365-385.

Cheetham, M.D., Keene, A.F., Bush, R.T., Sullivan, L.A. and Erskine, W.D. (2008) 'A comparison of grain-size analysis methods for sand-dominated fluvial sediments' *Sedimentology* **55** pp1905–1913.

Cheddadi, R., Lamb, H.F., Giot, J. and Van Der Kaars, S. (1998) 'Holocene climatic change in Morocco: a quantitative reconstruction from pollen data' *Climate Dynamics* **14** pp883–890.

Chen, F.H., Bloemendal, J., Zhang, P.Z. and Liu, G.X. (1999) 'An 800 ky proxy record of climate from lake sediments of the Zoige Basin, Eastern Tibetan Plateau' *Palaeogeography, Palaeoclimatology, Palaeoecology* **151** pp307-320.

Chen, J., Wan, G., Zhang, D.D., Chen, Z., Xu, J., Xiao, T. and Huang, R. (2005) 'The 'Little Ice Age' recorded by sediment chemistry in Lake Erhai, Southwest China' *Holocene* **15** pp925-931.

Chenet, M., Roussel, E., Jomelli, V. and Grancher, D. (2010) 'Asynchronous Little Ice Age glacial maximum extent in Southeast Iceland' *Geomorphology* **114** pp253–260.

Chiverrell, R.C., Oldfield, F., Appleby, P.G., Barlow, D., Fisher, E., Thompson, R. and Wolf, G. (2008) 'Evidence for changes in Holocene sediment flux in Semer Water and Raydale, North Yorkshire, UK' *Geomorphology* **100** pp70-82.

Chueca, C.J. and Julián, A.A. (2004) 'Relationship between solar radiation and the development and morphology of small cirque glaciers (Maladeta Mountain Massif, Central Pyrenees, Spain)' *Geograpfiska Annaler* **86A** (1) pp81-89.

Cía, J.C., Andrés, A.J., Saz Sánchez, M.A., Creus Novau, J. and López Moreno, J.I. (2005) 'Responses to climatic changes since the Little Ice Age on Maladeta Glacier (Central Pyrenees)' *Geomorphology* **68** pp167-182.

Clague, J.J., Barendregt, R., Enkin, R.J., Froit, F.F. (2003) 'Paleomagnetic and tephra evidence for tens of Missoula floods in Southern Washington' *Geology* **31** pp247-250.

Clark, R. (1981) 'Cold climate features of the Cordillera Cantábrica, Northern Spain' *Biuletyn Peryglacjalny* **24** pp5-18.

Cohen, M.C.L., Behling, H. and Lara, R.J. (2005) 'Amazonian mangrove dynamics during the last millennium: The relative sea-level and the Little Ice Age' *Review of Palaeobotany and Palynology* **136** pp93-108.

Collins, P.E.F., Rust, D.J., Bayraktutan, M.S. and Turner, S.D. (2005) 'Fluvial stratigraphy and palaeoenvironments in the Pasinler Basin, Eastern Turkey' *Quaternary International* **140-141** pp121-134.

R e f e r e n c e s

Cook, E.R., D'Arrigo, R.D. and Briffa, K.R. (1998) 'A reconstruction of the North Atlantic Oscillation using tree-ring chronologies from North America and Europe' *The Holocene* **8** pp9-17 .

Cook, E.R., D'Arrigo, R.D. and Mann, M.E. (2002) 'A Well-Verified, multiproxy reconstruction of the winter North Atlantic Oscillation index since A.D. 1400' *Journal of Climate* **15** pp1754-1764.

Craigie, R. and Sugden, H. (2008) 'Tilia and Tilia*Graph manual' Available: http://phytolith.missouri.edu/guide.htm

Crasquin-Soleau, S. and Kershaw, S. (2005) 'Ostracod fauna from the Permian–Triassic boundary interval of South China (Huaying Mountains, Eastern Sichuan Province): palaeoenvironmental significance' *Palaeogeography, Palaeoclimatology, Palaeoecology* **217** pp131–141.

Cronin, T.M., Dwyer, G.S., Kamiya, T., Schwede, S. and Willard, D.A. (2003) 'Medieval Warm Period, Little Ice Age and 20th century temperature variability from Chesapeake Bay' *Global and Planetary Change* **36** pp17-29.

Crowley, T.J. and Lowery, T.S. (2000) 'How warm was the medieval warm period?' *Ambio* **29** (1) pp51-54.

Crusius, J. and Kenna, T.C. (2007) 'Ensuring confidence in radionuclide-based sediment chronologies and bioturbation rates' *Estuarine, Coastal and Shelf Science* **71** pp537-544.

Cushing, E.J. (1967) 'Evidence for differential pollen preservation in late Quaternary sediments in Minnesota' *Review of Palaeobotany and Palynology* **4** pp87-101.

Cwynar, L.C., Burden, E. and McAndrews, J.H. (1979) 'An inexpensive sieving method for concentrating pollen and spore from fine-grained sediments' *Canadian Journal of Earth Science* **16** pp1115-1120.

Cushing, E.J. and Wright, H.E. (1967) 'Hand-operated piston corers for lake sediments' *Ecology* **46** pp380-384.

Cvijić, J. (1907) 'Beobachtungen über die Eiszeit auf der Balkanhalbinsel, in den Südkarpathen und auf dem mysischen Olymp' *Zeitschrift für Gletscherkunde* **3** pp1-35.

Cvijić, J. (1913) 'The ice age in the Prokletije and surrounding mountains (in Serbian)' *Glas Srpske Kraljevske Akademije Nauka* **XCI** pp1-149.

Cvijić, J. (1917) 'L'époque glaciaire dans la péninsule balkanique' *Annales de Geographie* **26** pp189-218.

Dale, M.B. and Walker, D, (1970) 'Information analysis of pollen diagrams' *Pollen Spores* **12** pp21-37.

Dalton, C., Birks, H.J.B., Brooks, S.J., Cameron, N.G., Evershed, R.P., Peglar, S.M., Scott, J.A. and Thompson, R. (2005) 'A multi-proxy study of lake-development in response to catchment changes during the Holocene at Lochnagar, North-East Scotland' *Palaeogeography, Palaeoclimatology, Palaeoecology* **221** pp175- 201.

References

Dansgaard, W., Johnsen, S.J., Reeh, N., Gundestrup, N., Clausen, H.B. and Hammer, C.U. (1975) *Nature* **255** pp24-28

Darrigo, R.D., Malmstrom, C.M., Jacoby, G.C., Los, S.O. and Bunker, D.E. (2000) 'Correlation between maximum latewood density of annual tree rings and NDVI based estimates of forest productivity' *International Journal of Remote Sensing* **21** (11) pp2329-2336.

David, F. (2010) 'Expansion of green alder (*Alnus alnobetula* [Ehrh] K. Koch) in the Northern French Alps: A palaeoecological point of view' *Comptes Rendus Biologies* **333** pp424-428.

David, M.B. and Ford, M.S. (1982) 'Sediment Focusing in Mirror Lake, New Hampshire' *Limnology and Oceanography* **27** (1) pp137-150.

Davis, O.K. (1999) 'Pollen analysis of Tulare Lake, California: Great Basin-like vegetation in Central California during the full-glacial and early Holocene' *Review of Palaeobotany and Palynology* **107** pp249–257.

Day, S.P. (1991) 'Post-glacial vegetational history of the Oxford region' *New Phytologist* **119** pp445-470.

Dearing, J.A. (1983) 'Changing pattern of sediment accumulation in a small lake in Scani, Southern Sweden' *Hydrobiologia* **103** pp59-64.

Dearing, J.A. (1991) 'Lake sediment records of erosional processes' *Hydrobiologia* **214** pp99-106.

Dearing, J.A. (1994) Environmental magnetic susceptibility: Using the Bartington MS2 system, Bartington instruments handbook (7) Available: http://www.gmw.com/magnetic_properties/pdf/Om0409%20J_Dearing_Handbook_iss7.pdf

Dearing, J.A. (1997) 'Sedimentary indicators of lake-level changes in the humid temperate zone: A critical review' *Journal of Paleolimnology* **18** pp1-14.

Dearing, J.A. (1999) *Magnetic Susceptibility* In: Walden, J., Oldfield, F. and Smith, J. (eds.) Environmental magnetism: A practical guide, Quaternary Research Association, Oxford, pp35-62.

Dearing, J.A., Elner, J.K., and Happey-Wood, C.M. (1981) 'Recent sediment flux and erosional processes in a Welsh upland lake-catchment based on magnetic susceptibility measurements' *Quaternary Research* **16** (3) pp356-372.

De Deckker, P. (1988) 'An account of the techniques using ostracodes in paleolimnology in Australia' *Palaeogeography, Palaeoclimatology, Palaeoecology* **62** pp463-475.

De Deckker, P. (2002) *Ostracod Palaeoecology* In: Holmes, J. and Chivas, A. (eds.): The Ostracoda: Applications in Quaternary Research, AGU Geophysical Monograph **Series 131** pp5-36.

Delgado Úbeda, J., Boada, J.M. and Hernández Pacheco, F. (1932) Guía de los sitios naturales de interés nacional. El Parque Nacional de la Montaña de Covadonga. Comisaría de Parques Nacionales.

Delmas, M., Gunnell, Y., Braucher, R., Calvet, M. and Bourlès, D. (2008) 'Exposure age chronology of the last glaciation in the Eastern Pyrenees' *Quaternary Research* **69** (2) pp231-241.

R e f e r e n c e s

Delorme, L.D. (1989) ‘Methods in Quaternary ecology no.7: Freshwater ostracodes’ *Geoscience Canada* **16** pp85-90.

Denèfle, M., Lèzin, A-M., Fouache, E. and Dufaure, J-J (2000) ‘A 12,000-Year Pollen Record from Lake Maliq, Albania’ *Quaternary Research* **54** pp423–432.

Denton, G.H. and Karlén, W. (1973) 'Lichenometry: Its application to Holocene moraine studies in Southern Alaska and Swedish Lapland' *Arctic and Alpine Research* **5** (4) pp347-372.

Desloges, J.R. and Gilbert, R. (1994) ‘Sediment source and hydroclimatic inferences from glacial lake sediments: the postglacial sedimentary record of Lillooet Lake, British Columbia’ *Journal of Hydrology* **159** pp375-393.

Devillers, B. and Provansal, M. (2003) ‘La Morphogenèse d’un géosystème cultivé depuis le Néolithique recent: les petits basins versants de la moyenne vallée de l’Herault (France) – Morphogenesis in catchments cultivated since the late Neolithic: case studies from the middle Hérault valley (France)’ *Géomophologie : relief, processus, envionnement* **2** pp83-98.

Digerfeldt, G. (1986) *Studies on past lake-level fluctuations* In: Berglund, B.E. (ed.) Handbook of Holocene palaeoecology and palaeohydrology, John Wiley and Sons, Chichester.

Dobiński, 2005 ‘Short Communication: Permafrost of the Carpathian and Balkan Mountains, Eastern and Southeastern Europe’ *Permafrost and Periglacial Processes* **16** pp395-398.

Doner, L. (2003) ‘Late-Holocene paleoenvironments of Northwest Iceland from lake sediments’ *Palaeogeography, Palaeoclimatology, Palaeoecology* **193** (3) pp535-560.

D’Orefice, M., Pecci, M., Smiraglia, C. and Ventura, R. (2000) ‘Retreat of Mediterranean glaciers since the Little Ice Age: case study of Ghiacciaio del Calderone, Central Apennines, Italy’ *Arctic, Antarctic and Alpine Research* **32** pp197-201.

Driese, S.G., Ashley, G.M., Li, Z.H., Hover, V.C. and Owen, R.B. (2004) ‘Possible Late Holocene equatorial palaeoclimate record based upon soils spanning the Medieval Warm Period and Little Ice Age, Loboi Plain, Kenya’ *Palaeogeography, Palaeoclimatology, Palaeoecology* **213** pp231-250.

Dyurgerov, M.B. and Meier, M.F. (2007) ‘Twentieth century climate change: Evidence from small glaciers’ *Proceedings of the National Academy of Sciences of the United States of America* **97** (4) pp1406-1411.

Eastwood, W.J., Pearce, N.J.G., Westgate, J.A., Perkins, W.T., Lamb, H.F. and Roberts, N. (1999) ‘Geochemistry of Santorini tephra in lake sediments from SouthWest’ *Global and Planetary Change* **21** pp17–29.

Eckstein, D. (1972) ‘Tree-ring research I Europe’ *Tree-ring Bulletin* **32** pp1-18.

Eckstein, D. (1982) *Europe* In: Hughes, M.K., Kelly, P.M., Pilcher, J.R. and La Marche, V.C. (eds.) Climate from tree-rings, Cambridge University Press, Cambridge pp142-148.

Eddy, J.A. (1976) ‘The Maunder Minimum: The reign of Louis XIV appears to have been a time of real anomaly in the behaviour of the sun’ Science **192** pp1189-1202.

R e f e r e n c e s

Edwards, M.E., Bigelow, N.H., Finney, B.P. and Eisner, W.R. (2000) 'Records of aquatic pollen and sediment properties as indicators of late-Quaternary Alaskan lake levels' *Journal of Paleolimnology* **24** pp55-68.

Edwards, J.K. and Hirons, K.R. (1984) 'Cereal pollen grains in pre-elm decline deposits: Implications for the earliest agriculture in Britain and Ireland' *Journal of Archaeological Science* **11** pp71-80.

Edwards, J.K. and Rowntree, K.M. (1980) *Radiocarbon and palaeoenvironmental evidence for changing rates of erosion at a Flandrian stage site in Scotland* In: Cullingford, R.A., Davidson, D.A. and Lewin, J. (eds.) Timescales in Geomorphology, John Wiley and Sons, Chichester, pp207-224.

Edwards, K.J. and Whittington, G. (2001) 'Lake sediments, erosion and landscape change during the Holocene in Britain and Ireland' *Catena* **42** pp143–173.

Edwards, K.J., Whittington, G., Robinson, M. and Richter, D. (2005) 'Palaeoenvironments, the archaeological record and cereal pollen detection at Clickimin, Shetland, Scotland' *Journal of Archaeological Science* **32** pp1741-1756.

Ely, L.L., Webb, R.H. and Enzel, Y. (1992) 'Accuracy of Post-bomb ^{137}Cs and ^{14}C in Dating Fluvial Deposits' *Quaternary Research* **38** pp196-2004.

Enters, D., Kirchner, G. and Zolitschka, B. (2006) 'Establishing a chronology for lacustrine sediments using a multiple dating approach - A case study from the Frickenhauser See, central Germany' *Quaternary Geochronology* **1** pp249-260.

Erdtman, G. (1969) Handbook of palynology (morphology-taxonomy-ecology): An Introduction to the study of pollen grains and spores, Munksgaard, Copenhagen.

Erlinger, C., Lettner, H., Hubmer, A., Hofmann, W. and F. Steinhäusler (2008) 'Determining the Chernobyl impact on sediments of a pre-Alpine lake with a very comprehensive set of data' *Journal of Environmental Radioactivity* **99** pp1294–1301.

Esteve, J.F., Imeson, A., Jarman, A., Jarman, R., Barberis, R., Rydell, B., Sánchez, V.C. and Vandekerckhove, L. (2004*) VOLUME II – TASK GROUPS ON SOIL EROSION*
Task group 1 Pressures and Divers Causing Soil Erosion In van Camp, L., Bujarrabal, B., Gentile, A.R., Jones, R.J.A., Montanarella, L., Olazabal, C. and Selvaradjou, S.-K. (eds.) Reports of the technical working groups established under the thematic strategy for soil protection Volume II Erosion, European Commission, Directorate-General Joint Research Centre, European Environment Agency.

Evans, M.E. and Heller, F. (2003) Environmental magnetism: Principles and applications of enviromagnetics, Academic Press, Amsterdam.

Everest, J. and Bradwell, T. (2003) 'Buried glacier ice in Southern Iceland and its wider significance' *Geomorphology* **52** pp347-358.

Eydoux, D. and Maury, l. (1907) 'Les glacier du Pic long' *La Geographie* **XVI** pp1-18.

Faegri, K. and Inversen, J. (1975) Textbook of pollen analysis (3rd edition), Blackwell Scientific Publications, Oxford.

R e f e r e n c e s

Fagan, B. (2000) The Little Ice Age, Basic Books, New York.

Faust, D., Zielhofer, C., Escudero, R.B. and Diaz del Olmo, F. (2004) 'High-resolution fluvial record of late Holocene geomorphic change in Northern Tunisia: climate or human impact?' *Quaternary Science Reviews* **23** (16-17) pp1757-1775.

Feng, Z.-D. and Johnson, W.C. (1995) 'Factors affecting the magnetic susceptibility of a loess-soil sequence, Barton County, Kansas, USA' *Catena* **24** (1) pp25-37.

Fernández-Illescas, F., Javier, F., Nieva, J., Inmaculada, S., Tormo, R. and Muñoz, A.F. (2010) 'Pollen production of Chenopodiaceae species at habitat and landscape scale in Mediterranean salt marshes: An ecological and phenological study' *Review of Palaeobotany and Palynology* **161** (3-4) pp127-136.

Feurdean, A., Mosbrugger, V., Onac, B.P., Polyak, V. and Veres, D. (2007) 'Younger Dryas to mid-Holocene environmental history of the lowlands of NW Transylvania, Romania' *Quaternary Research* **68** pp364-378.

Fey, M., Korr, C., Maidana, N.I., Carrevedo, M.L., Corbella, H., Dietrich, S., Haberzettl, T., Kuhn, G., Lücke, A., Mayr, C., Ohlendorf, C., Paez, M.M., Quintana, F.A., Schäbitz, F. and Zolitschka, B. (2009) 'Palaeoenvironmental changes during the last 1600 years inferred from the sediment record of a cirque lake in Southern Patagonia (Laguna Las Vizcachas, Argentina)' *Palaeogeography, Palaeoclimatology and Palaeoecology* **281** pp363-375.

Finsinger, W., Tinner, W., van der Knaap, W.O. and Ammann, B. (2006) 'The expansion of hazel (*Corylus avellana* L.) in the Southern Alps: a key for understanding its early Holocene history in Europe?' *Quaternary Science Reviews* **25** pp612-631.

Fisher, J.B., Lick, W.J., McCall, P.L. and Robbins, J.A. (1990) 'Vertical Mixing of Lake Sediments by Tubificid Oligochaetes' *Journal of Geophysical Research* **85** (C7) pp3997-4006.

Folland, C.K. and Karl, T.R. (2001) *Observed Climate Variability and Change* In: Houghton, J.T., Ding, Y., Griggs, D.J., Nouger, M., van der Linden, P.J., Dai, X., Maskell, K. and Johnson, C.A. (eds.) Climate Change 2001: The scientific basis, Cambridge University Press, Cambridge pp101-165.

Follieri, M., Magri, D. and Sadori, L. (1988) '250,000-year pollen record from Valle di Castiglione (Roma)' *Pollen et Spores* **30** pp329–356.

Follieri, M., Magri, D. and Sadori, L. (1989) 'Pollen stratigraphical synthesis from Valle Di Castiglione (Roma)' *Quaternary International* **3/4** pp81-84.

Ford, D. and Williams, P. (1989) Karst geomorphology and hydrology, Unwin Hayman, London.

Forester, R.M. (1991) 'Pliocene-climate history of the Western United States derived from lacustrine ostracodes' *Quaternary Science Reviews* **10** pp133-146.

Foster, I.D.L., Mighall, T.M., Proffitt, H., Walling, D.E. and Owens, P.N. (2006) 'Post-depositional ^{137}Cs mobility in the sediments of three shallow coastal lagoons, SW England' *Journal of Paleolimnology* **35** pp881-895.

R e f e r e n c e s

Fouache, E., Desruelles, S., Magny, M., Bordon, A., Oberweiler, C., Coussot, C., Touchais, G., Lera, P., Lèzine, A-M, Fadin, L. and Roger, R. (2010) 'Palaeogeographical reconstructions of Lake Maliq (Korça Basin, Albania) between 14,000 BP and 2000 BP' *Journal of Archaeological Science* **37** (3) pp525-535.

Fouache, E., Dufaure, J-J., Denèfle, M., Lézine, A-M., Léra, P., Prendi, F. And Touchais, G. (2001) 'Man and environment around Lake Maliq (Southern Albania) during the Late Holocene' *Vegetation History and Archaeobotany* **10** pp79-86.

Frank, D. and Esper, J. (2005a) 'Characterization and climate response patterns of a high-elevation, multi-species tree-ring network in the European Alps' *Dendrochronologia* **22** pp107-121.

Frank, D. and Esper, J. (2005b) 'Temperature reconstructions and comparisons with instrumental data from a tree-ring network for the European Alps' *International Journal of Climatology* **25** pp1437-1454.

Frenzel, P. and Boomer, I. (2005) 'The use of ostracods from marginal marine, brackish waters as bioindicators of modern and Quaternary environmental change' *Palaeogeography, Palaeoclimatology, Palaeoecology* **225** pp68- 92.

Friedman, G.M. (1979) 'Address of the retiring President of the International Association of Sedimentologists: Differences in size distributions of populations of particles among sands of various origins' *Sedimentology* **26** pp3-32.

Friedman, G.M. and Sanders, J.E. (1978) Principles of sedimentology, Wiley, New York.

Friedrich, W.L., Kromer, B., Friedrich, M., Hainemeier, J., Pfeiffer, T. and Talamo, S. (2006) 'Santorini eruption radiocarbon dated to 1627–1600 B.C.' *Science* **312** pp548.

Friedrichs, D., Büntgen, U., Frank, D.C., Esper, J., Neuwirth, B. and Löffler (2009) 'Complex climate controls on 20th century oak growth in Central-West Germany' *Tree Physiology* **29** pp39-51.

Frisia , S., Borsato, A., Spötl, C., Villa, I.M. and Cucchi, F. (2005) 'Climatic variability in the SE Alps of Italy over the past 17,000 years reconstructed form a stalagmite record' *Boreas* **34** pp445-455.

Fritts, H.C. (1971) 'Dendroclimatology and dendroecology' *Quaternary Research* **1** (4) pp419-449.

Fritts, H.C. (1976) Tree Rings and Climate, Academic Press, London.

Fritz, P., Morgan, A.V., Eicher, U. and McAndrews, J.H. (1987) 'Stable isotope, fossil coleoptera and pollen stratigraphy in late Quaternary sediments from Ontario and New York state' *Palaeogeography, Palaeoclimatology, Palaeoecology* **58** pp183-202.

Frogley, M.R., Griffiths, H.I. and Heaton, T.H.E. (2001) 'Historical biogeography and Late Quaternary environmental change of Lake Pamvotis, Ioannina (North-Western Greece): Evidence from Ostracods' *Journal of Biogeography* **28** (6) pp745-756.

Frogley, M.R., Tzedakis, P.C. and Heaton, T.H.E. (1999) 'Climate Variability in Northwest Greece during the Last Interglacial' *Science* **285** pp1886-1889.

R e f e r e n c e s

Furlan, D. (1977) *The climate of Southeast Europe* In Wallen, C.C. (ed.) Climates of central and Southern Europe, Elsevier, Amsterdam pp185-223.

Gabrovec, M. (1998) 'The Triglav glacier between 1986 and 1998' *Geografski zbornik* **38** pp89-110.

Galanidou, N., Tzedakis, P.C., Lawson, I.T. and Frogley, M. R. (2000) 'A revised chronological and palaeoenvironmental framework for the Kastritsa rockshelter, Northwest Greece' *Antiquity* **74** (284) pp349-355.

Galatowitsch, S.M. and van der Valk, A.G. (1996) 'The Vegetation of Restored and Natural Prairie Wetlands' *Ecological Applications* **6** (1) pp102-112.

García-Sellés, C., Peña, J.C., Martí, G., Oller, P. and Martínez, P. (2010) 'WeMOI and NAOi influence on major avalanche activity in the Eastern Pyrenees' *Cold Regions Science and Technology* **64** (2) pp137-145.

García, M., Zapata, M.B.R., Santisteban, J.I., Mediavilla, R., López-Pamo, E. and Dabrio, C.J. (2007) 'Late Holocene environments in Las Tablas de Daimiel (South central Iberian peninsula, Spain)' *Vegetation History and Archaeobotany* **16** pp241–250.

Gaston, K.J. and David, R. (1994) 'Hotspots across Europe' *Biodiversity Letters* **2** pp108-116.

Gat, J.R. and Magaritz, M. (1980) 'Climatic variations in the Eastern Mediterranean Sea area' *Naturwissenschaften* **67** pp80-87.

Gaus-Krigerova (1970) Prokletije; Jezerce; Plav maps 1:25,000.

Gellatly, A., Smiraglia, C., Grove, J.M. and Lathan, R. (1994) 'Recent variations of Ghiacciaio del Calderone, Abruzzi, Italy' *Journal of Glaciology* **40** (136) pp486-490.

Genty, D. and Massault, M. (1999) 'Carbon transfer dynamics from bomb-^{14}C and δ^{13}C time series of a laminated stalagmite from SW France - Modelling and comparison with other stalagmite records' *Geochimica et Cosmochimica Acta* **63** (10) pp1537-1548.

Geophysical network of Montenegro (2010) 'Montenegro Seismological Observatory', Accessed at: http://www.seismo.co.me/1e.html (04.09.10).

Gerstengarbe, F.-W. and Werner, P.C. (2009) 'A short update on Koeppen climate shifts in Europe between 1901 and 2003' *Climatic Change* **92** pp99-107.

Ghasemi, A.R. and Khalili, D. (2008) 'The effect of the North Sea-Caspian pattern (NCP) on winter temperatures in Iran' *Theoretical and Applied Climatology* **92** (1-2) pp59-74.
Ghosh, S. (2006) 'Regression-based age estimation of a stratigraphic isotope sequence in Switzerland' *Vegetation History & Archaeobotany* **15** pp273–278.

Gibbons, A.B., Megeath, J.D. and Pierce, K.L. (1984) 'Probability of moraine survival in a succession of glacial advances' *Geology* **12** (6) pp327-330.

R e f e r e n c e s

Gilbert, E., Travi, Y., Massault, M., Chernet, T., Barbecot, F. and Laggoun-Defarge, F. (1999) 'Comparing carbonate and organic AMS-^{14}C ages in Lake Abiyata sediments (Ethiopia): Hydrochemistry and palaeoenvironmental implications' *Radiocarbon* **41** (3) pp271-286.

Gill, D. (1973) 'Floristics of a plant sucession sequence in the Mackenzie Delta, Northwest territories' *Polarforschung* **43** pp55-65.

Giraudi, C. (2005) 'Middle to Late Holocene glacial variations, periglacial processes and alluvial sedimentation on the higher Apennine massifs (Italy)' *Quaternary Research* **64** pp176–184.

Gob, F., Petit, F., Bravard, J.-P., Ozer, A. and Gob, A. (2003) 'Lichenometric application to historical and sub-recent dynamics and sediment transport of a Corsican stream (Figarella River - France)' *Quaternary Science Reviews* **22** pp2111–2124.

Gob, F., Bravard, J.-P., Jacob, N. and Petit, F. (2005) *Determining the competence of mountainous Mediterranean streams using lichenometric techniques* In: Batalla, R.J. and Garcia, C. (eds.) Geomorphological Processes and Human Impacts in River Basins. Proceedings of the International Conference Held at Solsona, Catalonia, Spain **vol. 299** pp161-170 IAHS Red Book.

Gob, F., Jacob, N., Bravard, J.-P. and Petit, F. (2008) 'The value of lichenometry and historical archives in assessing the incision of submediterranean rivers from the Little Ice Age in the Ardèche and upper Loire (France)' *Geomorphology* **94** pp170-183.

Gómez, A., Palacios, D., Ramos, M., Tanarro, L.M., Schulte, L. and Salvador, F. (2001) 'Location of Permafrost in Marginal Regions: Corral del Veleta, Sierra Nevada, Spain' *Permafrost and Periglacial Processes* **12** pp93-110.

Goñi, M.S., Cacho, I., Turon, J., Guiot, J., Sierro, F., Peypouquet, J., Grimalt, J., and Shackleton, N. (2002) **'**Synchroneity between marine and terrestrial responses to millennial scale climatic variability during the last glacial period in the Mediterranean region' *Climate Dynamics* **19** (1) pp95-105.

González Suárez, J.J. and Alonso, V. (1994) 'Glaciers in Picos de Europa, Cordillera Cantábrica, Northwest Spain' *Journal of Glaciology* **40** pp198–199.

González Suárez, J.J. and Alonso, V. (19960 'Reply to the comments of Frochoso and Castañón on Glaciers in Picos de Europa, Cordillera Cantábrica, Northwest Spain' *Journal of Glaciology* **42** pp386–389.

González Trueba, J.J. (2003/2004) 'Topoclimatical factors and very small glaciers in Atlantic mountain of SW Europe: The Little Ice Age glacier advance in Picos de Europa (NW Spain)' *Zeitschrift für Gletscherkunde und Glazialgeologie* **39** pp115-125.

González Trueba, J.J. (2005) 'La Pequeña Edad del Hielo en los Picos de Europa (Cordillera Cantábrica, NO España) Análisis morfológicoy reconstrucción del avance glaciar histórico' *Revista Cuaternario and Geomorfologia* **19** pp79-94.

González Trueba, J.J., Martín Moreno, R., Martínez de Pisón, E. and Serrano, E. (2008) ''Little Ice Age' glaciation and current glaciers in the Iberian Peninsula' *The Holocene* **18** (4) pp551-568.

R e f e r e n c e s

González Trueba, J.J. and Serrano, E. (2005) 'Assessment of geomorphosites in natural protected areas: the Picos de Europa National Park (Spain)' *Géomorphologie, Relief, Processus, Environnement* **3** pp197-208.

González-Sampériz, P. (2004) 'Evolución paleoambiental del sector central de la Cuenca del Ebro durante el Pleistoceno superior y Holoceno' Instituto Pirenaico de Ecología-CSIC, Zaragoza pp210.

González-Sampériz, P., Utrilla, P., Mazo, C., Valero-Garcés, B., Sopena, M.C., Morellón, M., Sebastián, M., Moreno, A. and Martínez-Bea, M. (2009) 'Patterns of human occupation during the early Holocene in the Central Ebro Basin (NE Spain) in response to the 8.2 ka climatic event' *Quaternary Research* **71** (2) pp121-132.

Google Earth (2010) 'Google earth' Accessed at: http://earth.google.co.uk/

Goosse, H., Arzel, O., Luterbacher, J., Mann, M.E., Renssen, H., Riedwyl, N., Timmermann, A., Xoplaki, E. and Wanner, H. (2006) 'The origin of the European 'Medieval Warm Period'' *Climate of the Past* **2** pp99-113.

Gordon, N., McMahon, T.A. and Finlayson, B.L. (1992) Stream hydrology: An introduction for ecologists, Wiley, Chichester.

Gosse, J.C. and Phillips, F.M. (2001) 'Terrestrial in situ cosmogenic nuclides: theory and application' *Quaternary Science Reviews* **20** pp1475-1560.

Gray, A.B., Pasternack, G.B. and Watson, E.B. (2010) 'Hydrogen peroxide treatment effects on the particle size distribution of alluvial and marsh sediments' *The Holocene* **20** pp293-301.

Gregory, M.R. and Johnston, K.A. (1987) 'A nontoxic substitute for hazardous heavy liquids-aqueous sodiumpolytungstate ($3Na_{2}WO_{4}9WO_{3}H_{2}0$) solution' *New Zealand Journal of Geology and Geophysics* **30** pp317-320.

Greig-Smith, P. (1957) *Quantitative Plant Ecology*, Butterworths scientific publications, London.

Griffiths, H.I. and Holmes, J.A (2000) Non-marine ostracods and Quaternary palaeoenvironments, QRA technical Guide No.8 Quaternary Research Association, London.

Griffiths, H.I., Reed, J.M., Leng, M.J., Ryan, S. and Petkovski, S. (2002) 'The recent palaeoecology and conservation status of Balkan Lake Dojran' *Biological Conservation* **104** pp35-49.

Grimm, E.C. (1987) 'CONISS: A fortran 77 program for stratigraphically constrained cluster analysis by the method of incremental sum of squares' *Computers and Geoscience* **13** (1) pp13-35.

Grimm, E.C. (1991) TILLIA and TILLIA-GRAPH, Springfield, Illinois State Museum.

Grove, J.M. (1979) 'The glacial history of the Holocene' *Progress in Physical Geography* **3** pp1-54.

Grove, J.M. (1979) 'The glacial history of the Holocene' *Progress in Physical Geography* **3** pp1-54.

Grove, J.M. (1988) The Little Ice Age **Vol. I and II** Methuen, London.

References

Grove, J.M. (2001a) ‘The initiation of the “Little Ice Age” in regions round the North Atlantic’ *Climatic Change* **48** pp53-82.

Grove, A.T. (2001b) ’The ”Little Ice Age” and its geomorphological consequences in Mediterranean Europe’ *Climatic Change* **48** pp121-136.

Grove, J.M. and Gellatly, A.F. (1995) ‘Little Ice Age glacier fluctuations in the Pyrénées’ *Zeitschrift für Gletscherkunde und Glazialgeologie* **31** pp199-206.

Grudd, H. (1989) ‘Small glaciers as sensitive indicators of climatic fluctuations’ *Geografiska Annaler* **72A** (1) pp119-123.

Grunewald, K., Weber, J., Scheithauer, F. and Haubold, F. (2006) ‘Mikrogletscher im Piringebirge (Bulgarien)’ *Zeitschrift für Gletscherkunde und Glazialmorphologie* **39** pp99-114.

Grunewald, K., Haubold, F. & Gebel, M. (1999): Ökosystemforschung SüdWest-Bulgarien – Untersuchungen zur Struktur, Funktion und Dynamik der Landschaften im nördlichen Pirin und im Becken von Razlog. Dresdener Geographische Beiträge, H. 5.

Guichard, F., Carey, S., Arthur, M.A., Sigurdsson, H. and Arnold, M. (1993) ‘Tephra from the Minoan eruption of Santorini in sediments of the Black Sea’ *Science* **363** pp610-612.

Guiot, J., Pons, A., De Beaulieu, J.L. and Reille, M. (1989) ‘A 140,000-year continental climate reconstruction from two European pollen records’ *Nature* **338** pp309-313.

Guitérrez, E. (1989) ‘Dendroclimatological study of *Pinus sylvestris* L. in Southern Catalonia (Spain)’ *Tree-Ring Bulletin* **49** pp1-9.

Gündüz, M. and Özsoy, E. (2005) ‘Effects of the North Sea Caspian pattern on surface fluxes of Euro-Asian-Mediterranean seas’ *Geophysical Research Letters* **32** ppL21701(1-4).

Haeberli, W. and Hoelzle, M. (1993) ‘Fluctuations of glaciers 1985 – 1990’ (Vol VI) International Association of Hydrological Sciences (International Commission of Snow and Ice) United Nations Environmental Programme and United Nations Educational, Scientific and Cultural Organisation.

Hajdas, I. (2006) ‘Radiocarbon dating for Quaternary studies - New developments and perspectives’ *Quaternary International* **154** pp160-163.

Håkanson, L. and Jansson, M. (1983) Principles of lake sedimentology, Springer-Verlag, Berlin.

Hanson, T., Brooks, T.M., da Fonseca, G.A.B., Hoffmann, M., Lamoreux, J.F., Machlis, G., Mittermeier, C.G., Mittermeier, R.A. and Pilgrim, J.D (2008) ‘Warfare in Biodiversity Hotspots’ *Conservation Biology* **23** (3) pp578-587.

Harrison, S. P. and G. Digerfeldt (1993) ‘European lakes as paleohydrological and paleoclimatic indicators’ *Quaternary Science Reviews* **12** pp233-248.

Harrison, S.P., Prentice, C. and Guiot, J. (1993) ‘Climatic controls on Holocene lake-level changes in Europe’ *Climate Dynamics* **8** pp189-200.

R e f e r e n c e s

Harrison, S.P., Yu, G. and Tarasov, P.E. (1996) 'Late Quaternary lake-level record from Northern Eurasia' *Quaternary Research* **45** pp138-159.

Havinga, A.J. (1967) 'Palynology and pollen preservation' *Review of Palaeobotany and Palynology* **2** pp81-98.

Heiri, O., Lotter, A.F. and Lemcke, G. (2001) 'Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results' *Journal of Paleolimnology* **25** pp101-110.

Heller, F. and Tungsheng, L. (1984) 'Magnetism of Chinese loess deposits' *Geophysical Journal of the Royal Astronomical Society* **77** (1) pp125-141.

Hibbert, F.A. (1978) *The vegetation history of the Somerset Levels* In: Limbrey, S. and Evans, J.G. (eds.) CBA Research Report No.21 The effect of man on the landscape: The lowland zone, The Chameleon Press Ltd., London.

Higgitt, S.R., Oldfield, F. and Appleby, P.G. (1991) 'The record of land use change and soil erosion in the late Holocene sediments of the Petit Lac d'Annecy, Eastern France' *The Holocene* **1** (1) pp14-28.

Hildreth, W. and Fierstein, J. (2000) 'Katmai volcanic cluster and the great eruption of 1912' *Geological Society of America Bulletin* **112** pp1594-1620.

Hiruta, S.F., Hiruta, S.-I. And Mawatari, S.F. (2007) 'A new, interstitial species of Terrestricythere (Crustacea: Ostracoda) and its microdistribution at Orit Beach, Northeastern Sea of Japan' *Hydrobiologia* **585** pp43-56.

Hodell, D.A., Brenner, S., Curtis, J.H., Medina-González, Ildefonso-Chan Can, E., Albornaz-Pat, A. and Guilderson, T.P. (2005) 'Climate change on the Yucatan Peninsula during the Little Ice Age' *Quaternary Research* **63** (2) pp109-121.

Hoelzle, M., Haeberli, W., Dischl, M. and Peschke, W. (2003) 'Secular glacier mass balances derived from cumulative glacier length changes' *Global and Planetary Change* **36** pp295-306.

Holmes, J.A. (1996) 'Trace-element and stable-isotope geochemistry of non-marine ostracod shells in Quaternary palaeoenvironmental reconstruction' *Journal of Paleolimnology* **15** pp23-235.

Holmes, J.A. and Chivas, A.R. (2002) 'Ostracod shell chemistry – overview' *The Ostracoda: Applications in Quaternary Research Geophysical Monograph* **131** pp185-204.

Holmes, R.M. and Bassett, I.J. (1963) 'Effect of meterological events on Ragweed pollen count' *International Journal of Biometerology* **7** (1) pp27-34.

Holzhauser, H. Magny, M. and Zumbühl, H.J. (2005) 'Glacier and lake-level variations in West-central Europe over the last 3500 years' *The Holocene* **15** (6) pp789- 801.

Horne, D.J. (2007) 'A Mutual Temperature Range method for Quaternary palaeoclimatic analysis using European nonmarine Ostracoda' *Quaternary Science Reviews* **26** pp1398-1415.

R e f e r e n c e s

Horne, D.J., Cohen, A. and Martens, K. (2002) *Taxonomy, morphology and biology of Quaternary and Living Ostracoda* In: Holmes, J. and Chivas, A. (eds.): The Ostracoda: Applications in Quaternary Research, AGU Geophysical Monograph **Series 131** pp5-36.

Hose, L.D. and Pisarowicz, A. (1999) 'Cueva de Villa Luz, Tabasco, Mexico: Reconnaissance study of an active sulphur spring cave and ecosystem' *Journal of Cave and Karst Studies* **61** (1) pp13-21.
Hotchkiss, S.E., Calcote, R. and Lynch, E.A. (2007) 'Response of vegetation and fire to Little Ice Age climate change: regional continuity and landscape heterogeneity' *Landscape Ecology* **22** pp25–41.

Hrouda, F., Chlupáčová, M. and Chadima, M. (2009) 'The use of magnetic susceptibility of rocks in geological exploration (case histories study)' Terraplu publications, Accessed at: http://www.terraplus.ca/products/magsus/case-study-susceptibility.pdf (14.12.10).

Hu, S., Goddu, S.R., Appel, E. and Verosub, K. (2007) 'Fine-tuning of age integrating magnetostratigraphy, radiocarbon dating, and carbonate cyclicity: Example of lacustrine sediments from Heqing basin (Yunnan, China) covering the past 1 Myr' *Journal of Asian Earth Sciences* **30** (3-4) pp423-432.

Huang, D., Wang, Y., Masae, S., Liu, X., Song, B., Chen, J., Takahashi, S., Hori, Y. and Yamarnura, Y. (2007) 'Spatial Heterogeneity of Vegetation in China' *Tshinghua Science and Technology 12* (4) pp424-434.

Huber Verlag/Lokalna turistička organizacija Plav (2008) Montenegro: Prokletije hiking and biking, map 1:50,000, Huber Verlag, Belgrade.

Hughes, P.D. (2004) 'Quaternary Glaciation in the Pindus Mountains, Northwest Greece' Unpublished PhD thesis, University of Cambridge.

Hughes, P.D. (2007) 'Recent behaviour of the Debeli Namet glacier, Durmitor, Montenegro' *Earth Surface Processes and Landforms* **32** pp1593-1602.

Hughes, P.D. (2008) 'Response of a Montenegro glacier to extreme summer heat waves in 2003 and 2007' *Geografiska Annaler: Series A Physical Geograph*y **90** (4) pp259-267.

Hughes, P.D. (2009a) 'Twenty-first Century Glaciers and Climate in the Prokletije Mountains, Albania' *Arctic, Antarctic, and Alpine Research* **41** (4) pp455-459.

Hughes, P.D. (2009b) 'Little Ice Age glaciers in the Balkans: low altitude glaciation enabled by cooler temperatures and local topoclimatic controls' *Earth Surface Processes and Landforms* **35** (2) pp229-241.

Hughes, P.D. (2010) 'Little Ice Age glaciers in the Balkans: low altitude glaciation enabled by cooler temperatures and local topoclimatic controls' *Earth Surface Processes and Landforms* **35** pp229-241.

Hughes, P.D. and Braithwaite, R.J. (2008) 'Application of a degree-day model to reconstruct Pleistocene glacial climates' *Quaternary Research* **69** pp110–116.

Hughes, M.K. and Diaz, H.F. (1994) 'Was there a 'Medieval Warm Period', and if so, where and when?' *Climatic Change* **26** pp109-142.

References

Hughes, P.D., Gibbard, P.L. and Woodward, J.C. (2007) Geological controls on Pleistocene glaciation and cirque form in Greece' *Geomorphology* **88** pp242-253.

Hughes, P.D., Woodward, J.C. and Gibbard, P.L. (2006) 'Quaternary glacial history of the Mediterranean mountains' *Progress in Physical Geography* **30** (3) pp334-364.

Hughes, P.D., Woodward, J.C., van Calsteren, P.C., Thomas, L.E. and Adamson, K.R. (2010) 'Pleistocene ice caps on the coastal mountains of the Adriatic Sea' *Quaternary Science Reviews* **29** pp3690-3708.

Humphries, M.S., Kindness, A., Ellery, W.N., Hughes, J.C. and Benitez-Nelson, C.R. (2010) '^{137}Cs and ^{210}Pb derived sediment accumulation rates and their role in the long-term development of the Mkuze River floodplain, South Africa' *Geomorphology* **119** pp88-96.

Huntley, B. (1990) 'European vegetation history: palaeovegetation maps from pollen data - 13 000 yr BP to present' *Journal of Quaternary Science* **5** (2) pp130-122.

Huntley, B. (1993) *Rapid early-Holocene migration and high abundance of hazel (Corylus avellana **L.**): alternative hypotheses* In Chambers, F.M. (ed.) Climate change and human impact on the landscape, Chapman and Hall, London.

Hurrell, J.W. (1995) 'Decadal trends in the North Atlantic Oscillation: regional temperatures and precipitation' *Science* **269** (5224) pp676-670.

Hurrell, J.W. and van Loon, H. (1997) 'Decadal variations in climate associated with the North Atlantic Oscillation' *Climate Change* **36** pp301-326.

Hutchinson (1957) A treatise on limnology, volume 1: Geography, physics, and chemistry John Wiley and Sons, New York.

Ilus, E. and Saxén, R. (2005) 'Accumulation of Chernobyl-derived ^{137}Cs in bottom sediments of some Finnish lakes' *Journal of Environmental Radioactivity* **82** pp199-221.

Imhof, P. (2010) 'Glacier fluctuations in the Italian Mont Blanc massif from the Little Ice Age until the present: Historical reconstructions for the Miage, Brenva and Pré-de-Bard Glaciers' MSc thesis, University of Bern, Accessed at: http://www.climatestudies.unibe.ch/students/theses/msc/33.pdf (31.06.10).

Institute for Geological Exploration (1968) Gusinje geological sheet, no. K34-52 (Translated by Mijović-Pilić, D.).

IPCC (2007) *Summary for Policymakers* In: Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M.Tignor and H.L. Miller (eds.) Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, United Kingdom and New York.

ISSG (2006) 'Typha latifolia (aquatic plant)' Accessed at: http://www.issg.org/database/species/ecology.asp?si=895&fr=1&sts=&lang=EN (21.10.10).

References

Ito, E. (2002) *Mg/Ca, Sr/Ca, $\delta^{18}O$ and $\delta^{13}C$ Chemistry of Quaternary lacustrine Ostracode shells from the North American continental interior* In: Holmes, J. and Chivas, A. (eds.): The Ostracoda: Applications in Quaternary Research, AGU Geophysical Monograph **Series 131** pp5-36.

Iversen, J. (1944) '*Viscum*, *Hedera* and *Ilex* as climatic indicators: A contribution to the study of the Post-Glacial temperature climate' *Geologiska Föreningens i Stockholm Förhandlingar* **66** pp463-483.

Ivy-Ochs, S., Kerschner, H., Maisch, M., Christl, M., Kubik, P.W. and Schlüchter, C. (2009) 'Latest Pleistocene and Holocene glacier variations in the European Alps' *Quaternary Science Reviews* **28** pp2137-2149.

Jackson, S.T. (1990) 'Pollen source area and representation in small lakes of the NorthEastern United States' *Review of Palaeobotany and Palynology* **63** pp53-76.

Jalut, G., Amat, A.E., Bonnet, L., Gauquelin, T. and Fontugne, M. (2000) 'Holocene climatic changes in the Western Mediterranean, from South-East France to South-East Spain' *Palaeogeography, Palaeoclimatology, Palaeoecology* **160** pp255-290.

Jansenn, C.R. (1981) 'Contemporary pollen assemblage from the Vosges (France)' *Review of Palaeobotany and Palynology* **33** pp183-313.

Jenkins, D.G. and Shackleton, N. (1979) 'Parallel changes in species diversity and palaeotemperature in the Lower Miocene' *Nature* **278** (1) pp50-51.

Jennings, J.N. (1985) Karst Geomorphology, Basil Blackwell Ltd., Oxford.

Jensen, C., Kunzendorf, H. and Vorren, K.-D. (2002) 'Pollen deposition rates in peat and lake sediments from the *Pinus sylvestris* L. forest-line ecotone of Northern Norway' *Review of Palaeobotany and Palynology* **121** pp113-132.

Jex, C.N., Baker, A., Eden, J.E., Eastwood, W.J., Fairchild, I.J., Leng, M.J., Thomas, L. and Sloane, H.J. (*in press*) 'A 500 yr speleothem-derived reconstruction of late autumn–winter precipitation, North East Turkey' *Quaternary Research*.

Jimenez-Moreno, G., Fauquette, S. and Suc, J.-P. (2010) 'Miocene to Pliocene vegetation reconstruction and climate estimates in the Iberian Peninsula from pollen data' *Review of Palaeobotany and Palynology* **162** (3) pp403-415.

Joerin, U.E., Stocker, T.F. and Schlüchter, C. (2006) 'Multicentury glacier fluctuations in the Swiss Alps during the Holocene' *The Holocene* **16** (5) pp697-704.

Johnston, V.E., McDermott, F. and Tămaş, T. (2010) 'A radiocarbon dated bat guano deposit from N.W. Romania: Implications for the timing of the Little Ice Age and Medieval Climate Anomaly' *Palaeogeography, Palaeoclimatology, Palaeoecology* **291** pp217-227.

Joly, C., Barillé, L., Barreau, M., Mancheron, A. And Visset, L. (2007) 'Grain and annulus diameter as criteria for distinguishing pollen grains of cereals from wild grasses' *Review of Palaeobotany and Palynology* **146** pp221-233.

R e f e r e n c e s

Joly, C. and Visset, L. (2009) 'Evolution of vegetation landscapes since the Late Mesolithic on the French West Atlantic coast' *Review of Palaeobotany and Palynology* **154** pp124-179.

Jones, P.D. and Bradley, R.S. (1995) *Climatic variations in the longest instrumental records* In Bradley, R.S. and Jones, P.D. (eds.) Climate since AD 1500, Routledge pp246–68.

Jones, P.D. and Briffa, K.R. (2006) 'Unusual climate in Northwest Europe during the period 1730 to 1745 based on instrumental and documentary data' *Climate Change* **79** pp361-379.

Jones, P.D., Roberts, C.N., Leng, M.J. and Türkeş, M. (2006) 'A high-resolution late Holocene lake isotope record from Turkey and links to North Atlantic and monsoon climate' *Geology* **34** (5) pp361-364.

Joosten, H. and Klerk, P.D. (2002) 'What's in a name? Some thoughts on pollen classification, identification, and nomenclature in Quaternary palynology' *Review of Palaebotany and Palynology* **122** pp29-45.

Jordan, C.F., Fryer, G.E. and Hemmen, E.H. (1971) 'Size analysis of silt and clay by hydrophotometer' *Journal of Sedimentary Petrology* **41** pp489-496.

Jomelli, V., Delval, C., Grancher, D., Escande, S., Brunstein, D., Hetu, B., Filion, L. and Pech, P. (2007) 'Probabilistic analysis of recent snow avalanche activity and weather in the French Alps' *Cold Regions Science and Technology* **47** pp180-192.

Jorissen, F.J., Asioli, A., Borsetti, A.M., Capotondi, L., de Visser, J.P., Hilgen, F.J., Rohling, E.J., van der Borg, K., Vergnaud-Grazzini, C., Zachariasse, W.J. (1993) 'Late Quaternary central Mediterranean biochronology' *Marine Micropaleontology* **21** pp169-189.

Juliá, R., Burjachs, F., Dasí, M.J., Mezquita, F., Miracle, M.R., Roca, J.R., Seret, G. and Vicente, E. (1998) 'Meromixis origin and recent trophic evolution in the Spanish mountain lake La Cruz' *Aquatic Sciences* **60** pp279-299.

Julián, A. and Chueca, J. (1998) 'Le Petit Âge Glaciaire dans les Pyrénées Centrales Meridionales: estimation des paléotempératures á partir d'inférences géomorphologiques' *Sud-Ouest Europeen* **3** pp79–88.

Jull, A.J.T. (2007) *Dating techniques* In: Elias, S.A. (ed.) Encyclopedia of Quaternary Science, Elsevier Science, Amsterdam.

Karayigit, A.I., Akgun, F., Gayer, R.A. and Temel, A. (1999) 'Quality, palynology, and palaeoenvironmental interpretation of the Ilgin lignite, Turkey' *International Journal of Coal Geology* **38** pp219-236.

Karlén, W. (1988) 'Scandinavian glacial and climate fluctuation during the Holocene' *Quaternary Science Reviews* **7** pp190-209.

Kartografija Huber Verlag/Lokalna turistička organizacija Plav (2008) Montenegro: Prokletije, hiking and biking map 1:50 000.

Kelly, P.M., Leuschner, H.H., Briffa, K.R. and Harris, I.C. (2002) 'The climatic interpretation on pan-European signature years in oak ring-width series' *The Holocene* **12** pp689-694.

R e f e r e n c e s

Kidson, E.J. and Williams, G.L. (1969) ‘Concentrations of palynomorphs by use of sieves’ *Oklahoma Geology Notes* **29** pp117-119.

Kienast, F. and Kuhn, N. (1989) ‘Simulating forest succession along ecological gradients in Southern Central Europe’ *Plant Ecology* **79** (1-2) pp7-20.

Kilyeni, T.I. (1969) *The problems of Ostracod ecology in the Thames Estuary* In: Neale, J.W (ed.) The taxonomy, morphology and ecology of recent ostracoda, Oliver and Boyd Ltd., Edinburgh.

Kiss, A. (2007) ‘Factors affecting spatial and temporal distribution of Ostracoda assemblages in different macrophyte habitats of a shallow lake (Lake Fehér, Hungary)’ *Hydrobiologia* **585** pp89-98.

Knipping, M. 2008 ‘Early and Middle Pleistocene pollen assemblages of deep core drillings in the Northern Upper Rhine Graben, Germany’ *Netherlands Journal of Geosciences (Geologie en Mijnbouw)* **87** (1) pp51-65.

Koff, T., Punning, J.-M. and Kangur, M. (2000) ‘Impact of forest disturbance on the pollen influx in lake sediments during the last century’ *Review of Palaeobotany and Palynology* **111** pp19-29.

Kohnke, H. (1968) Soil Physics, McGraw-Hill, New York.

Koppen, W. (1923) Dle Klirnate der Erde, Walter de Gruyter, Berlin.

Köppen, W. (1936) *Das geographische System der Klimate*, In: Köppen, W. and Geiger, R. (eds.) Handbuch der Klimatologie, Bd. Gebr Bornträger I. Berlin, pp 1–44.

Korhola, A., Sorvari, S., Rautio, M., Appleby, P.G., Dearing, J.A., Hu, Y., Rose, N., Lami, A. and Camerson, N.G. (2002) ‘A multi-proxy analysis of climate impacts on the recent development of subarctic Lake Saanajärvi in Finnish Lapland’ *Journal of Paleolimnology* 28 (1) pp59-77.

Körner, C. and Paulsen, J. (2004) ‘A world-wide study of high altitude treeline temperatures’ *Journal of Biogeography* **31** pp713-732.

Kornicker, L.S. and Sohn, I.G. (1971) *Variability of ostracod eggs egested by fish and effect of digestive fluids on ostracode shells – ecologic and palaeoecologic implication* In: Oertli, H.J. (ed.) Paléoécologie Ostracodes *Bulletin du Centre du Recherches PAU-SNPA* **5** pp125-135.

Kostic, S. and Parker, G. (2003) ‘Physical and numerical modelling of deltaic sedimentation in lakes and reservoirs’ *Proceedings, XXX International Association of Hydraulic Research Congress, Thessaloniki, Greece* **August** pp24-29.

Koutsios, A., Kontopoulos, N., Kalisperi, D., Soupios, P. And Avramidis, P. (2010) ’Sedimentological and geophysical observations in the delta plain of Selinous River, Ancient Helike, Northern Peloponnesus Greece’ Bulletin of the Geological Society of Greece, Proceedings of the 12th International Congress, Petras, May 2010, pp654-662.

Kováts, N., Abdel-Hameid, N.-A., Kovács, K. and Paulovits, G. (2010) ‘Sensitivity of three unionid glochidia to elevated levels of copper, zinc and lead’ *Knowledge and Management of Aquatic Ecosystems* **399** pp04p2-04p8.

R e f e r e n c e s

Kranjc, A. (2009) ‘History of deforestation and reforestation in the Dinaric Karst’ *Journal of the Institute of Australian Geographers* **47** (1) pp15-23.

Kremenetski, K.V., Boettger, T., MacDonald, G.M., Vaschalova, T., Sulerzhitsky, L. and Hiller, A. (2004) ‘Medieval climate warming and aridity as indicated by multiproxy evidence from the Kola Peninsula, Russia’ *Palaeogeography, Palaeoclimatology, Palaeoecology* **209** pp113–125.

Krishnaswami, S., Benninger, L.K., Aller, R.C. and Von Damm, K.L. (1980) ‘Atmospherically-derived radionuclides as tracers of sediment mixing and accumulation in near-shore marine and lake sediments: Evidence from ^{7}Be, ^{210}Pb and ^{240}Pu’ *Earth and Planetary Science Letters* **47** pp307-318.

Krishnaswamy, S., Lal, D., Martin, J.M. and Meybeck, M. (1971) ’Geochronology of lake sediments’ *Earth and Planetary Science Letters* **11** pp407-414.

Kruss, P.D. and Hastenrath, S. (1987) ‘The role of radiation geometry in the climate response of Mount Kenya's glaciers, part I: Horizontal reference surfaces’ *International Journal of Climatology* **7** (5) pp493-505.

Kuhn, M. (1984) ‘Mass budget imbalances as criterion for a climatic classification of glaciers’ *Geografiska Annaler* **66A** pp229-238

Kuhn, M. (1989) *The response of the equilibrium line altitude and mass balance profiles to climate fluctuations* In: Oerlemans, J. (ed.) Glacier fluctuations and climate change, Kluwer, Germany pp407-417.

Kuhn, M. (1993) *Possible future contribution to sea level change from small glaciers* In: Warrick, R.A., Barrow, E.M. & Wigley, T.M.L. (eds.) Climate and Sea Level Change, Cambridge Univ. Press, Cambridge, pp134–143.

Kuhn, M., Schlosser, E. and Span, N. (1997) ‘Eastern alpine glacier activity and climatic records since 1860’ *Annals of Glaciology* **24** pp54–59.

Kukla, G., Heller, F., Ming, L.X., Chun, X.T., Sheng, L.T. and Sheng, A.Z. (1988) ‘Pleistocene climates in China dated by magnetic susceptibility’ *Geology* 16 pp811-814.

Külköylüoğlu, O. (2003) ‘Ecology of freshwater Ostracoda (Crustacea) from lakes and reservoirs in Bolu, Turkey’ *Journal of Freshwater Ecology* **18** (3) pp343-347.

Külköylüoğlu, O. and Dügel, M. (2004) ‘Ecology and spatiotemporal patterns of Ostracoda (Crustacea) from Lake Gölcück (Bolu, Turkey)’ *Archiv für Hydrobiologie* **160** (1) pp67-83.

Kullman, L. (1998) ‘The occurrence of thermophilous trees in the Scandes Mountains during the early Holocene: evidence for a diverse tree flora from macroscopic remains’ *Journal of Ecology* **86** pp421-428.

Kunaver, J. (1991) ‘Corrosion terraces as geoecological response to postglacial development of glaciokarst rock surface’ *Proceedings of the International Conference on Environmental Changes in Karst areas: Italy 1991* **13** pp325-331.

Kundzewicz, Z. W. and Parry, M. L. (2001) *Europe* In: McCarthy, J.J., Canziani, O.F., Leary, N.A., Dokken, D.J. and White, K.S. (eds.) Climate change 2001: impacts, adaptation and vulnerability.

R e f e r e n c e s

Contribution of Working Group II to the Third Assessment Report of the Intergovernmental Panel on Climate Change, Third Assessment Report, **vol. II**, Cambridge University Press, Cambridge.

Kutiel, H. and Benaroch, Y. (2002) 'North Sea-Caspian Pattern (NCP) – an upper level atmospheric teleconnection affecting the Eastern Mediterranean: Identification and definition' *Theoretical and Applied Climatology* **71** pp17-28.

Kutiel, H., Maheras, P. and Guika, S. (1996) 'Circulation indices over the Mediterranean and Europe and their relationship with rainfall conditions across the Mediterranean' *Theoretical and Applied Climatology* **54** pp125-138.

Kutiel, H., Maheras, P., Türkes, M. and Paz, S. (2002) 'North Sea – Caspian Pattern (NCP) – an upper level atmospheric teleconnection affecting the Eastern Mediterranean – implication on the regional climate' *Theoretical Applied Climatology* **72** pp173-192.

Kwiatkowski, A. and Lubliner-Mianowska, K. (1957) 'Badanie skladu chemicznego pylku II Badania blon pylkowich' *Acta Societatis Botanicorum Polonae* **26** pp501-514.

Lakhdar, R., Soussi, M., Ben Ismail, M.H. and M'Rabet, A. (2006) 'A Mediterranean Holocene restricted coastal laggon under arid climate: case of the sedimentary record of the Sabkha Boujmel (SE Tunisia)' *Palaeogeography, Palaeoclimatology, Palaeoecology* **241** pp177–191.

La Marche Jr., V.C and Hirschboeck, K.K. (1984) 'Frost rings in trees as records of major volcanic eruptions' *Nature* **307** (5946) pp121-126.

Lamb, H.H. (1965) 'The early medieval warm epoch and its sequel' *Palaeogeography, Palaeoclimatology, Palaeoecology* **1** pp13-37.

Lamb, H.H. (1977) Climate: Present, Past and Future (vol. II), Meuthuen, London.

Lamb, H.H. (1995) Climate, history and the modern world, Routledge, Oxford.

Lamb, P.J. and Peppler, R.A. (1987) 'North Atlantic Oscillation: Concept and an application' *Bulletin American Meteorological Society* **68** (10) pp1218-1225.

Lamentowicz, M., Milecka, K., Galka, M., Cedro, A., Pawlyta, J., Piotrowska, N., Lamentowicz, L. and van der Knaap, W. (2008) 'Climate and human induced hydrological change since AD 800 in an ombrotrophic mire in Pomerania (N Poland) tracked by testate amoebae, macro-fossils, pollen and tree rings of pine' *Boreas* **38** (2) pp214-229.

Lanci, L., Hirt, A.M., Lowrie, W., Lotter, A.F., Lemcke, G. and Strum, M. (1999) 'Mineral-magnetic record of Late Quaternary climatic changes in a high Alpine lake' *Earth and Planetary Science Letters* **170** pp49-59.

Lanci, L., Hirt, A.M., Lotter, A.F. and Sturm, M. (2001) 'A record of Holocene climate in the mineral magnetic record of Alpine lakes: Sagistalsee and Hinterburgsee' *Earth and Planetary Science Letters* **188** (1) pp29-44.

Lau, T.-C. and Stephenson, A.G. (1994) 'Effects of soil phosphorus on pollen production, pollen size, pollen phosphorus content, and the ability to sire seeds in Cucurbita pepo (Cucurbitaceae)' *Sexual Plant Reproduction* **7** pp215-220.

R e f e r e n c e s

Lawson, I.T., Al-Omari, S., Tzedakis, P.C., Bryant, C.L. and Christanis, K. (2005) 'Lateglacial and Holocene vegetation history at Nisi Fen and the Boras mountains, Northern Greece' *The Holocene* **15** (6) pp873-887.

Lawson, I., Frogley, M., Bryant, C., Preece, R. and Tzedakis, P. (2004) 'The Lateglacial and Holocene environmental history of the Ioannina basin, North-West Greece' *Quaternary Science Reviews* **23** pp1599-1625.

Leeman, A. and Niessen, F. (1994) 'Varve formation and the climatic record in an Alpine proglacial lake: calibrating annually- laminated sediments against hydrological and meteorological data' *The Holocene* **4** (1) pp1-8.

Lepš, J. and Hadincová, V. (1992) 'How Reliable Are Our Vegetation Analyses?' *Journal of Vegetation Science* **3** (1) pp119-124.

Leroyer, C., 1997. Homme, climat, végétation au tardi-et postglaciaire dans le bassin parisien : apports de l'étude palynologique des fonds de vallées. Thesis, Université de Paris I, France.

Le Roy Ladurie, E. (1971) Times of feast, times of famine, Allen and Unwin, London.

Lewin, J. and Woodward, J.C. (2009) *Karst Geomorphology and environmental change* In: Woodward, J.C. (ed.) The physical geography of the Mediterranean, Oxford University Press, Oxford.

Lewis, D.W. (1984) Practical Sedimentology, Hutchinson Ross, New York.

Li, Y., Wang, N., Morrill, C., Cheng, H., Long, H. and Zhao, Q. (2009) 'Environmental change implied by the relationship between pollen assemblages and grain-size in N.W. Chinese lake sediments since the Late Glacial' *Review of Palaeobotany and Palynology* **154** pp54-64.

Liaccos, L. G., and Mouloupoulos. (1967) 'Contributionto the identificationo f some range types of Quercus coccifera L.' North Greece Forest Research Center Research Bulletin **16** pp54 (in Greek with English summary)**.**

Lieffers, V.J. (1983) 'Growth of *Typha latifolia* in boreal forest habitats as measured by double sampling' *Aquatic Botany* **15** pp335-348.

Linderholm, H.W., Björklund, J.A., Seftigen, K., Gunnarson, B.E., Grudd, H., Jeong, J.-H., Drobyshev, I. and Liu, Y. (2010) 'Dendroclimatology in Fennoscandia – from past accomplishments to future potential' *Climate of the Past* **6** pp93-114.

Linick, T.W., Damon, P.E., Donahue, D.J. and Jull, A.F.T. (1989) 'Accelerator mass spectrometry: The new revolution in radiocarbon dating' *Quaternary International* **1** pp1-6.

Lionello, P., Malanotte-Rizzoli, P. and Boscolo, R. (eds., 2006) Mediterranean climate variability, *Developments in Earth and Environmental Sciences* **4** pp1-421.

Lippi, M.M., Guido, M., Menezzi, B.I., Bellini, C. and Montanari, C. (2007) 'The Massaciuccoli Holocene pollen sequence and the vegetation history of the coastal plains by the Mar Ligure (Tuscany and Liguria, Italy)' *Vegetation History and Archaeobotany* **16** pp267–277.

R e f e r e n c e s

Liquete, C., Arnau, P., Canals, M. and Colas, S. (2005) ‘Mediterranean river systems of Andalusia, Southern Spain, and associated deltas: A source to sink approach’ *Marine Geology* **222-223** pp471-495.

Livingstone, D.A. (1955) 'A lightweight piston sampler for lake deposits' *Ecology* **36** pp137-139.

Livingstone, D.A. (1967) 'The use of a filament tape in raising long cores from soft sediment' *Limnology and Oceanography* **12** pp346-348.

Ljubisavljević, K., Arribas, O., Džukic, G. and Carranza, S. (2007) ‘Genetic and morphological differentiation of Mosor rock lizards, *Dinarolacerta mosorensis* (Kolombatović, 1886), with the description of a new species from the Prokletije Mountain Massif (Montenegro) (Squamata: Lacertidae)’ *Zootaxa* **1613** pp1-22.

Llasat, M.-C. and Rodriguez, R. (1997) ‘Towards a regionalization of extreme rainfall events in the Mediterranean area’ *FRIEND '97 — Regional Hydrology: Concepts and Models for Sustainable Water Resource Management (Proceedings of the Postojna, Slovenia, Conference, September-October 1997)* **IAHS Publication No. 246** pp.215-222.

Lobo, I., Mozeto, A.A. and Aravena, R. (2001) ‘Paleohydrological investigation of Infernão Lake, Moji-Guaçu River watershed, São Paulo, Brazil’ *Journal of Paleolimnology* **26** pp119-129.

Lockwood, J.G. (2001) ‘Abrupt and sudden climatic transitions and fluctuations: A review’ *International Journal of Climatology* **21** pp1153–1179.

Lohmann, U., Sausen, R., Bengtsson, L., Cubasch, U., Perlwitz, J. and Roeckner, E. (1993) ‘The Köppen climate classification as a diagnostic tool for general circulation models’ *Climate Research* **3** pp177-193.

Longmore (McCallan), M.E., O'Leary, B.M. and Rose, C.W. (1983) ‘Caesium-137 profiles in the sediments of a partial-meromictic lake on Great Sandy Island (Fraser Island), Queensland, Australia’ *Hydrobiologia* **103** pp21-27.

Longmore, M.E., Torgersen, T., O’Leary, B.M. and Luly, J.G. (1986) ‘Cesium-137 redistribution in the sediments of the playa, Lake Tyrrell, Northwestern Victoria. I. Stratigraphy and cesium-137 mobility in the upper sediments’ *Palaeogeography, Palaeoclimatology, Palaeoecology* **54** pp181-195.

López-Bustins, J.A. (2007) ‘The Western Mediterranean Oscillation and Rainfall in the Catalan Pyrenees’ PhD Thesis, Department of Physical Geography and Regional Geographical Analysis, University of Barcelona, Catalonia.

Lotter, A.F., Appleby, B.G., Bindler, R., Dearing, J.A., Grytnes, J.A., Hofmann, W., Kamenik, C., Lami, A., Livingstone, D.M., Ohlendorf, C., Rose, N. and Sturm, M. (2002) ‘The sediment record of the past 200 years in a Swiss high-alpine lake: Hagelseewli (2339 m a.s.l.)’ *Journal of Paleolimnology* **28** (1) pp111-127.

Lotter, A.F., Heiri, O., Hofmann, W., van der Knaap, W.O., van Leeuwen, J.F.N, Walker, I.R. and Wick, L. (2006) ‚Holocene timber-line dynamics at Bachalpsee, a lake at 2265 m a.s.l. in the Northern Swiss Alps’ *Vegetation History and Archaeobotany* **15** pp295-307.

R e f e r e n c e s

Lowe, J.J. (1992) 'Lateglacial and early Holocene lake sediments from the Northern Apennines, Italy - pollen stratigraphy and radiocarbon dating' *Boreas* **21** pp193-208.

Lowe, J.J., Lowe, S., Fowler, A.J., Hedges, R.E.M. and Austin, T.J.F. (1988), 'Comparison of accelerator and radiometric radiocarbon measurements obtained from Late Devensian Lateglacial lake sediments from Llyn Gwernan, North Wales, UK' *Boreas* **17** pp355-369.

Lowe, J.J. and Walker, M.J.C. (1984) Reconstructing Quaternary environments, Addison Wesley Longman Ltd., Essex.

Lowe, J.J. and Watson, C. (1993) 'Lateglacial and early Holocene pollen stratigraphy of the Northern Apennines, Italy' *Quaternary Science Reviews* **12** pp727-738.

Lovvorn, M.B., Frison, G.C. and Tieszen, L.L. (2001) 'Paleoclimate and Amerindians: Evidence from stable isotopes and atmospheric circulation' *Proceedings of The National Academy of Science U.S.A.* **98** (5) pp2485-2490.

Luckman, B.H. and Osborn, G.D. (1979) 'Holocene glacier fluctuations in the Middle Canadian Rocky Mountains' *Quaternary Research* **11** pp52-77.

Lundqvist, G. (1925) 'Utvecklinshistoriska insjöstudier I Sydsverige' *Sveriges Geologiska Undersökning C* **330** pp1-29.

Lundqvist, G. (1927) 'Bodenablagerungen und Entwicklungstypen der Seen' *Binnergewässer* **2** pp1-122.

Luque, J.A., Juliá, R., Riera, S., Marqués, M.A., López-Sáez, J.A. and Mezquita, F. (2004) 'Requesta sedimentológica a los cambios ambientales de épocas históricas en el sur de la Peninsula Ibérica: La secuencia de la Laguna Grande de Archidona (Málaga)' *Geotemas* **6** pp113-116.

Luterbacher, J., Dietrich, D., Xoplaki, E., Grosjean, M. and Wanner, H. (2004) 'Variability, trends, and extremes since 1500 european seasonal and annual temperature' *Science* **303** pp1499-1503.

Luterbacher, J., Rickli, R., Xoplaki, E., Tinguely, C., Beck, C., Pfister, C., and Wanner, H. (2001) 'The Late Maunder Minimum (1675–1715) - a key period for studying decadal scale climatic change in Europe' *Climate Change* **49** pp441-462.

Luterbacher, J., Schmutz, C., Gyalistras, D., Xoplaki, E. and Wanner, H. (1999) 'Reconstruction of monthly NAO and EU indices back to AD 1675' *Geophysical Research Letters* **26** (17) pp2745-2748.

Luterbacher, J., Xoplaki, E., Dietrich, D., Jones, P.D., Davies, T.D., Portis, D., Gonazlez-Rouco, J.F., von Storch, H., Gyalistras, D., Casty, C. and Wanner, H. (2002) 'Extending North Atlantic Oscillation reconstruction back to 1500' *Atmospheric Science Letters* **2** (1-4) pp114-124.

Luterbacher, J., Xoplaki, E., Dietrich, D., Jones, P.D., Davies, T.D., Portis, D., Gonazlez-Rouco, J.F., von Storch, H., Gyalistras, D., Casty, C. and Wanner, H. (2002) 'Extending North Atlantic Oscillation reconstruction back to 1500' *Atmospheric Science Letters* **2** (1-4) pp114-124.

R e f e r e n c e s

Maas, G.S. and Macklin, M.G. (2002) 'The impact of recent climate change on flooding and sediment supply within a Mediterranean Mountains catchment, Southwestern Crete, Greece' *Earth Surface Processes and Landforms* **27** pp1087-1105.

Machlus, M., Enzel, Y., Goldstein, S.L., Marco, S. and Stein, M. (2000) 'Reconstructing low levels of Lake Lisan by correlating fan-delta and lacustrine deposits' *Quaternary International* **73/74** pp137-144.

Mackenzie, A.B. (2000) 'Environmental radioactivity: experience from the 20th century - trends and issues for the 21st century' *The Science of the Total Environment* **249** pp313-329.

Mackereth, F.J.H. (1958) 'A portable core sampler for lake deposits' *Limnology and Oceanography* **3** pp181-191.

Macklin, M.G., Lewin, J. and Woodward, J.C. (1995) *Quaternary fluvial systems in the Mediterranean basin* In: Lewin, J., Macklin, M.G. and Woodward, J.C. (eds.) Mediterranean Quaternary River Environments Balkema, Rotterdam, pp1-25.

Macklin, M.G., Benito, G., Gregory, K.J., Johnstone, E., Lewin, J., Michczyńska, D.J., Soja, R., Starkel, L. And Thorndycraft, V.R. (2006) 'Past hydrological events reflected in the Holocene fluvial record of Europe' *Catena* **66** pp145-154.

Macklin, M.G. and Woodward, J.C. (2009) *River systems and environmental change* In: Woodward, J.C. (ed.) The physical geography of the Mediterranean, Oxford University Press, Oxford, pp319-352.

Macklin, M.G., Tooth, S., Brewer, P.A., Noble, P.L. and Duller, G.A.T. (2010) 'Holocene flooding and river development in a Mediterranean steepland catchment: The Anapodaris Gorge, South central Crete, Greece' *Global and Planetary Change* **70** pp35-52.

Magaš, D. (2002) 'Natural-Geographic characteristics of the Boka Kotorska Area as the Basis of Development' *Geoadria* **7** (1) pp51-81.

Magny, M. (2004) 'Holocene climate variability as reflected by mid-European lake-level fluctuations and its probable impact on prehistoric human settlements' *Quaternary International* **113** pp65-79.

Magny, M., Miramont, C. and Siva, O. (2002) 'Assessment of the impact of climate and anthropogenic factors on Holocene Mediterranean vegetation in Europe on the basis of palaeohydrological records' *Palaeogeography, Palaeoclimatology, Palaeoecology* **186** pp47-59.

Magny, M., Mouthon, J. And Ruffaldi, P. (1995) 'Late Holocene level fluctuations of the Lake Ilay in Jura, France: sediment and mollusc evidence and climatic implications' *Journal of Paleolimnology* **13** pp219-229.

Maher, B.A. (1986) 'Characterisation of soils by mineral magnetic measurements' *Physics of the Earth and Planetary Interiors* **42** pp76-92.

Maher, B.A. and Thompson, R. (1991) 'Mineral magnetic record of the Chinese loess and paleosols' *Geology* **19** pp3-6.

R e f e r e n c e s

Mann, M.E. (2002) 'Little Ice Age Vol I: The Earth system: physical and chemical dimensions of global environmental change' In MacCracken, M. and Perry, J.S., (eds.) Encyclopaedia of global environmental change, Wiley & Sons Ltd, Chichester, pp504–509.

Margielewski, W., Krąpiec, M., Valde-Nowak, P., Zernitskaya, V. (2010) 'A Neolithic yew bow in the Polish Carpathians: Evidence of the impact of human activity on mountainous palaeoenvironment from the Kamiennik landslide peat bog' *Catena* **80** pp141-153.

Marinova, E. and Atanassova, J. (2006) 'Anthropogenic impact on vegetation and environment during the Bronze Age in the area of Lake Durankulak, NE Bulgaria:
Pollen, microscopic charcoal, non-pollen palynomorphs and plant macrofossils' *Review of Palaeobotany and Palynology* **141** pp165-178.

Marion, C., Dufois, F., Arnaud, M. and Vella, C. (2010) 'In situ record of sedimentary processes near the Rhŏne River mouth during winter events (Gulf of Lions, Mediterranean Sea)' *Continental Shelf Research* **30** pp1095-1107.

Marquer, L., Pomel, S., Abichou, A., Schulz, E., Kaniewski, D. and Van Campo, E. (2008) 'Late Holocene high resolution palaeoclimatic reconstruction inferred from Sebkha Mhabeul, Southeast Tunisia' *Quaternary Research* **70** pp240-250.

Martínez de Pisón, E. and Parra, A. (1988) 'Los glaciares actuales del Pirineo español' En La Nieve en el Pirineo espanñol, Madrid, M.O.P.U. pp 29-98.

Martin-Puertas, C., Jiménez-Espejo, F., Martínez-Ruiz, F., Nieto-Moreno, V., Rodrigo, M., Mata, M.P. and Valero-Garcés, B.L. (2010) 'Late Holocene climate variability in the SouthWestern Mediterranean region: an integrated marine and terrestrial geochemical approach' *Climate of the Past* **6** pp807-816.

Martin-Puertas, C., Valero-Garcés, B.L., Mata, P., González-Sampériz, P., Bao, R. and Stefanova, V. (2008) 'Arid and humid phases in Southern Spain during the last 4000 years: the Zoñar Lake record' *The Holocene* **18** (6) pp907-621.

Martín-Puertas, C., Valero-Garcés, B.L., Brauer, A., Mata, M.P., Delgado-Huertas, A. and Dulski, P. (2009) 'The Iberian–Roman Humid Period (2600–1600 cal yr BP) in the Zoñar Lake varve record (Andalucía, Southern Spain)' *Quaternary Research*, **71** (2) pp108-120.

Mass, G.S. and Macklin, M.G. (2002) 'The impact of recent climate change on flooding and sediment supply within a Mediterranean mountain catchment, SouthWestern Crete, Greece' *Earth Surface Processes and Landforms* **27** (10) pp1087-1105.

Matsuda, I. (2004) *River Morphology and channel processes* In: Dooge, J.C.I. (ed.) Fresh Surface Water, Encyclopedia of Life Support Systems (EOLSS), Developed under the Auspices of the UNESCO, EOLSS Publishers, Oxford ,UK (http://www.eolss.net) .

Matthes, F.E. (1939) 'Report of comité on Glaciers' *National Research Council. American Geographical Union* **20** pp518-523.

References

Matthews, J.A. (2005) ''Little Ice Age' glacier variations in Jotunheimen, Southern Norway: a study in regionally controlled lichenometric dating of recessional moraines with implications for climate and lichen growth rates' *The Holocene* **15** (1) pp1-19.

Matthews, J.A. and Briffa, K.R. (2005) 'The 'Little Ice Age': re-evaluation of an evolving concept' *Geografiska Annaler* **87A** pp17–36.

Matthews, J.A. and Shakesby, R.A. (1984) 'The status of the 'Little Ice Age' in Southern Norway: relative-age dating of Neoglacial moraines with Schmidt hammer and lichenometry' *Boreas* **13** pp333-346.

Mauquoy, D., van Geel, G., Blaauw, M. and van der Plicht, J. (2002) 'Evidence from Northwest European bogs shows 'Little Ice Age' climatic changes driven by variations in solar activity' *The Holocene* **12** (1) pp1-6.

Mazier, F., Galop, D., Gaillard, M.J., Rendu, C., Cugny, C., Legaz, A., Peyron, O. and Buttler, A. (2009) **'**Multidisciplinary approach to reconstructing local pastoral activities: an example from the Pyrenean Mountains (Pays Basque)' *The Holocene* **19** (2) pp171-188.

McCarthy, D.P. (2003) 'Estimating Lichenometric Ages by direct and indirect measurement of radial growth: A case study of *Rhizocarpon agg.* at the Illecillewaet Glacier, British Columbia' *Arctic, Antarctic, and Alpine Research* **35** (2) pp203-213.

McCarthy, D.P. (2007) *Lichenometry* In: Elias, S.A. (ed.) Encyclopedia of Quaternary Science, Elsevier Science, Amsterdam**.**

McCarthy, D.P. and Smith, D.J. (1995) 'Growth curves for Calcium-tolerant lichens in the Canadian Rocky Mountains' *Arctic and Alpine Research* **27** (3) pp290-297.

McCave, L.N., Bryant, R.J., Cook, H.F. and Coughanowr, C.A. (1986) Evalutaion of a laser diffraction size analyser for use with natural sediments' *Journal of Sedimentary Petrology* **56** pp561-564.

McGowan, S., Grauert, M. and Andersonm N.J. (2008) 'A Late Holocene record of landscape degradation from Heygsvatn, the Faroe Islands' *Palaeogeography, Palaeoclimatology, Palaeoecology* **264** pp11-24.

McNeill, J.R. (1992) *The mountains of the Mediterranean world: an environmental history*, Cambridge University Press, Cambridge.

Médail, F. and Diadema, K. (2006) 'Mediterranean plant biodiversity and human impact: macro and micro-regional approaches' *Annales de Géographie* **651** pp618-640.

Médail, F. and Diadema, K. (2009) 'Glacial refugia influence plant diversity patterns in the Mediterranean Basin' *Journal of Biogeography* **36** pp1333-1345.

Médail, F. and Quézel, P. (1997) 'Hot-spots analysis for conservation of plant biodiversity in the Mediterranean Basin' *Annals of the Missouri Botanical Garden* **84** (1) pp112-127.

Médail, F. and Quézel, P. (1999) 'Biodiversity Hotspots in the Mediterranean Basin: Setting Global Conservation Priorities' *Conservation Biology* **13** (6) pp1510-1513.

R e f e r e n c e s

Meisch, C. (2000) Freshwater Ostracoda of Western and Central Europe Spektrum, Heidelberg.

Menzel, R.G., Schiebe, F.R. and McHenry, J.R. (1985) 'Sediment Properties and Deposition in Lake Atoka, Oklahoma' *Proceedings of the Oklahoma Academy of Science* **65** pp45-49.

Messerli, B. (1980) 'Mountain glaciers in the Mediterranean area and in Africa' *World Glacier Inventory - Inventaire mondial des Glaciers* (Proceedings of the Riederalp Workshop, Actes de l'Atelier de Riederalp) **126** pp197-211.

Meyer, T.G. and Gross, P.L. (1985) 'Fatal systemic fluorosis due to hydrofluoric acid burns' *Annals of Emergency Medcine* **149** pp113-117.

Meyer, A. and Meyer, E.I. (2000) 'Discharge regime and the effect of drying on macroinvertebrate communities in a temporary karst stream in East Westphalia (Germany)' *Aquatic Science* **62** pp261-231.

Meyer, A., Meyer, E.I. and Meyer, C. (2003) 'Lotic communities of two small temporary karstic stream systems (East Westphalia, Germany) along a longitudinal gradient of hydrological intermittency' *Limnologica* **33** pp271-279.

Mezquita, F., Sanz-Brau, A. and Miracle, M.R. (1996) 'New data on freshwater ostracoda assemblages (Crustacea, Ostracoda) from Cunca (central Spain' *Bulletin de la Société des naturalists luxembourgeois* **97** pp239-247.

Michelier, P.L. (1887) 'Rapport sur les variations des glaciers de Pyrenees' *Annales du bureau Central Métérologiques de France* pp1-235.

Middeldorp, A. A. (1982) 'Pollen concentration as a basis for indirect dating and quantifying net organic and fungal production in a peat bog ecosystem' *Review of Palaeobotany and Palynology* **37** pp225-282.

Milivojević, M., Menković, L. and Ćalić, J. (2008) 'Pleistocene glacial relief of the central part of Mt. Prokletije (Albanian Alps)' *Quaternary International* **190** pp112-122.

Miller, C.D. (1969) 'Chronology of Neoglacial Moraines in the Dome Peak Area, North Cascade Range, Washington' *Arctic and Alpine Research* **1** (1) pp49-65.

Miras, Y., Ejarque, A., Orengo, H., Mora, S.R., Palet, J.M. and Poirand, A. (2010) 'Prehistoric impact on landscape and vegetation at high altitudes: An integrated palaeoecological and archaeological approach in the Eastern Pyrenees (Perafita valley, Andorra)' *Plant Biosystems* **144** (4) pp924-939.

Mischke, S. and Holmes, J.A. (2008) 'Application of lacustrine and marginal Ostracoda to palaeoenvironmental reconstruction' *Palaeogeography, Palaeoclimatology, Palaeoecology* **264** pp211-212.

Mittermeier, R.A., Myers, N., Thomsen, J.B., da Fonseca, G.A.B. and Oliveri, S. (1998) 'Biodiversity hotspots and major tropical wilderness areas: approaches to setting conservation priorities' *Conservation Biology* 12 (3) pp516-520.

R e f e r e n c e s

Moberg, A., Sonechkin, D.M., Holmgren, K., Datsenko, N.M. and Karlén, W. (2005) 'Highly variable Northern Hemisphere temperatures reconstructed from low- and high-resolution proxy data' *Nature* **433** pp613-617.

Mohammed, M.U., Bonnefille, R. and Johnson, T.C. (1995) 'Pollen isotopic records in Late Holocene sediments from Lake Turkana, Kenya' *Palaeogeography, Palaeoclimatology and Palaeoecology* **119** pp371-383.

Moore, D. (1966) 'Deltaic sedimentation' *Earth Science Reviews* **1** pp87-104.

Moore, P. D. and Webb, J.A. (1978) An illustrated guide to pollen analysis, Hodder and Stroughton, London.

Moore, P. D., Webb, J. A. and Collinson, M. E. (1991) Pollen Analysis (Second Edition), Blackwell Scientific Publications, Oxford.

Morellón, M., Valero-Garcés, B., Moreno, A., González-Sampériz, P., Mata, P., Romero, O., Maestro, M. and Navas, A. (2008) 'Holocene palaeohydrology and climate variability in NorthEastern Spain: The sedimentary record of Lake Estanya (Pre-Pyrenean range)' *Quaternary International* **181** (1) pp15-31.

Moreton, S.G., Rosqvist, G.C., Davies, S.J. and Bentley, M.J. (2004) 'Radiocarbon reservoir ages from freshwater lakes, South Georgia, Sub-Antarctic: Modern analogues from particulate organic matter and surface sediments' *Radiocarbon* **46** (2) pp621-626.

Morley, M.W., Woodward, J.C. and Whallon, R. (2006) 'Crvena Stijena: Unravelling a deep cave sediment record in the upland karst of Montenegro' Geoarchaeology 2006 Conference 10th - 13th September, 2006, Exeter University, UK.

Morley, M.W. and Woodward, J.C. (2011) 'The Campanian Ignimbrite (Y5) tephra at Crvena Stijena Rockshelter, Montenegro' *Quaternary Research* ***in press***.

Mottershead, D.N. and White, I.D. (1972) 'The lichenometric dating of glacier recession - Tunsbergdal, Southern Norway' *Geografiska Annaler* **54A** pp47-52.

Mourguiart, P. and Montenegro, M.E. (2002) *Climate change in the Lake Titicaca area: evidence from Ostracod ecology* In: Holmes, J. and Chivas, A. (eds.): The Ostracoda: Applications in Quaternary Research, AGU Geophysical Monograph **Series 131** pp5-36.

Mudie, P.J. and Byrne, R. (1980) 'Pollen evidence for historic sedimentation rates in California coastal marshes' *Estuarine and Coastal Marine Science* **10** pp305-316.

Mullins, C.E. (1977) 'Magnetic susceptibility of the soil and its significance in Soil Science: A review' *Journal of Soil Science* **28** pp223-246.

Munsell color (2000) Munsell soil color charts, revised washable edition, x-rite, Grand Rapids, Michigan.

Murray, M.R. (2002) 'Is laser particle size determination possible for carbonate-rich lake sediments?' *Journal of Paleolimnology* **27** pp173-183.

R e f e r e n c e s

Nesje, A. and Dahl, S.O. (2000) Glaciers and Environmental Change: Key issues in environmental change series, Oxford University Press, London.

NOAA (2008) 'NOAA palaeoclimatology: Tree ring' Accessed at: http://www.ncdc.noaa.gov/paleo/treering.html (05.05.09)

Noren, A.J., Bierman, P.R., Steig, E.J., Lini, A. And Southon, J. (2002) 'Millennial-scale storminess variability in the NorthEastern United States during the Holocene epoch' *Nature* **419** pp821-824.

Oerlemans, J. (2001) Glacier fluctuations and climate change, Kluwer, Germany

Oertli, H.J. (1971) *The aspects of Ostracode faunas: A possible new tool in petroleum sedimentology* In: Oertli, H.J. (ed.) Paléoécologie Ostracodes Bulletin, Centre de Recherches Pau-SNPA **5** pp137-151.

Ohmura, A. (2004) *Cryosphere during the twentieth century* In: Sparling, J.Y. and Hawkesworth, C.J. (eds.) The State of the Planet: Frontiers and Challenges in Geophysics, Washington DC, American Geophysical Union pp239-257.

Ohmura, A., Kasser, P. and Funk, M. (1992) 'Climate at the equilibrium line of glaciers' *Journal od Glaciology* **38** (130) pp397-411.

Oldfield, F. (1988) 'Magnetic and element analysis of recent lake sediments from the Highlands of Papua New Guinea' *Journal of Biogeography* **15** pp529-553.

Oldfield, F. (1996) 'The PALICLAS Project: synthesis and overview' *Journal Memorie dell'Istituto Italiano di Idrobiologia* **55** pp329-357.

Oldfield, F., Appleby, P.G. and Battarbee, R.W. (1978) 'Alternative ^{210}Pb dating: results from the New Guinea Highlands and Lough Erne' *Nature* **271** pp339-342.

Oldfield, F., Asioli, A., Accorsi, C.A., Mercuri, A.M., Juggins, S., Langone, L., Rolph, T., Trincardi, F., Wolff, G., Gibbs, Z., Vigliotti, L., Frignani, M., van der Post, K. and Branch, N. (2003) 'A high resolution late Holocene palaeo environmental record from the central Adriatic Sea' *Quaternary Science Reviews* **22** (2-4) pp319-342.

Olsson, I.U. (1986) *Radiometric dating* In: Berglund, B.E. Handbook of Holocene palaeoecology and palaeohydrology, John Wiley and Sons, Chichester.

O'Reilly, J., Vintró, L.L., Mitchell, P.I., Donohue, I., Leira, M., Hobbs, W. and Irvine, K. (2010) '^{210}Pb-dating of a lake sediment core from Lough Carra (Co. Mayo, Western Ireland): use of paleolimnological data for chronology validation below the ^{210}Pb dating horizon' *Journal of Environmental Radioactivity* **102** (5) pp495-499.

Osborn, G.D. and Taylor, J. (1975) 'Lichenometry on calcareous substrates in the Canadian Rockies' *Quaternary Research* **5** pp111-120.

O'Sullivan, P.E. (1994) 'The natural history of Slapton Ley national nature reserve' *Field Studies* **8** (2) pp403-449.

References

Pallás, R., Rodés, A., Braucher, R., Carcaillet, J., Ortuño, M., Bordonau, J., Bourlés, D., Vilaplana, J.M., Masana, E. and Santanach, P. (2006) 'Late Pleistocene and Holocene glaciation in the Pyrenees: a critical review and new evidence from 10 Be exposures ages, South-central Pyrenees' *Quaternary Science Reviews* **25** pp2937–2963.

Palmentola, G., Baboçi, K., Gruda, G.J. and Zito, G. (1995) 'Short communication: A note on rock glaciers in the Albanian Alps' *Permafrost and Periglacial Processes* **6** pp251-257.

Panek, T., Hradecký, J., Smolková, V., Šilhán, K., Minár, J. and Zernitskaya, V. (2010) 'The largest prehistoric landslide in NorthWestern Slovakia: Chronological constraints of the Kykula long-runout landslide and related dammed lakes' *Geomorphology* **120** (3-4) pp233-247.

Parris, A.S., Bierman, P.R., Noren, A.J., Prins, M.A. and Lini, A. (2010) 'Holocene paleostorms identified by particle size signatures in lake sediments from the NorthEastern United States' *Journal of Paleolimnology* **43** pp29–49.

Parry, M.L. and Carter, T.R. (1985) 'The effect of climatic variations on agricultural risk' *Climatic Change* **7** pp95-110.

Pasumot (1797) Voyages physiques dans les Pyrenees en 1787 et 1789.

Pavšek, M. (2004) 'The Skuta glacier' *Geografski obzornik* **51** pp11-17.

Pavšek, M. (2007) 'Ledenik pod skuto kot pokazatelj podnebnih sprememb v Slovenskem delu Alp' *Dela* **28** pp207-219.

Pederson, D.C., Peteet, D.M., Kurdyla, D. and Guilderson, T.P. (2005) 'Medieval Warming, Little Ice Age, and European impact on the environment during the last millennium in the lower Hudson Valley, New York, USA' *Quaternary Research* **63** (3) pp238-249.

Penck, A. (1897) Die Picos de Europa und das kantabrische Gebirge Geographische Zeitschrift Leipzig pp278–81.

Pennington, W., Cambray, R.S. and Fisher, E.M. (1973) 'Observations on lake sediments using fallout ^{137}Cs as a tracer' Nature 242 pp324-326.

Pensa, M., Salminen, H. and Jalkanen, R. (2005) 'A 250-year-long height-increment chronology for Pinus sylvestris at the Northern coniferous timberline: A novel tool for reconstructing past summer temperatures?' *Dendrochronologia* **22** pp75–81.

Pérez, L., Lorenschat, J., Bugja, R., Brenner, M., Scharf, B. and Schwalb, A. (2010) 'Distribution, diversity and ecology of modern freshwater ostracodes (Crustacea), and hydrochemical characteristics of Lago Petén Itzá, Guatemala' *Journal of Limnology* **69** (1) pp146-159.

Perry, D.A. (1978) 'An Estimate of the Effective Range of Pollen Dispersal in Lodgepole Pine *(Pinus contorts* Dougl.)' *Annals of Botany* **42** pp1001-1002.

Petit, R.J., Aguinagalde, I., de Beaulieu, J.-L., Bittkau, C., Brewer, S., Cheddadi, R., Ennos, R., Fineschi, S., Grivet, D., Lascoux, M., Mohanty, A., Müller-Starck, G., Demesure-Musch, B., Palmé, A., Pedro Martín, J., Rendell, S. and Vendramin, G.G. (2003) 'Glacial refugia: Hotspots but not melting post of genetic diversity' *Science* **300** pp1563-1565.

R e f e r e n c e s

Pfister, C. (1980) 'The Little Ice Age: Thermal and wetness indices for Central Europe' *Journal of Interdisciplinary History* **10** (4) pp665-696.

Pfister, C. (1995) *Monthly Temperature and Precipitation in Central Europe 1525–1979: Quantifying Documentary Evidence on Weather and its Effects* In: Bradley, R.A. and Jones, P.D. (eds.) Climate Since A.D. 1500, Routledge, London.

Pfister, C. (1994) *'Switzerland: The Time of Icy Winters and Chilly Springs* In: Frenzel, B., Pfister, C. and Gläser, B. (eds.) Climatic Trends and Anomalies in Europe 1675–1715, G. Fischer, Stuttgart, pp205-224.

Pfister, C. (1999) Wetternachhersage: 500 Jahre Klimavariationen Klimavariationen und Naturkatastrophen 1496–1995, Haupt-Verlag, Bern.

Pharo, C.H. and Carmack, E.C. (1979) 'Sedimentation processes in a short residence-time intermontane lake, Kamploops Lake, British Colombia' *Sedimentology* **26** pp523-541.

Pichard, G. (1995) 'Les crues sur le bas Rhône de 1500 à nos jours. Pour une histoire hydro-climatique' Méditerrane (3-4) pp105–116.

Pinglot, J.F. and Pourchet, M. (1995) 'Radioactivity measurements applied to glaciers and lake sediments' *The Science of the Total Environment* **173/174** pp211-223.

Pini, R., Ravazzi, C. and Donegana, M. (2009) 'Pollen stratigraphy, vegetation and climate history of the last 215 ka in the Azzano Decimo core (plain of Friuli, North-Eastern Italy)' *Quaternary Science Reviews* **28** (13-14) pp1268-1290.

Piva, A., Asioli, A., Trincardi, F., Schneider, R.R. and Vigliotti, L. (2008) 'Late-Holocene climate variability in the Adriatic Sea (Central Mediterranean)' *The Holocene* **18** (1) pp153-167.

Poesen, J.W.A. and Hooke, J.M. (1997) 'Erosion, flooding and channel management in Mediterranean environments of Southern Europe' *Progress in Physical Geography* **21** (2) pp157-199.

Pohl, F. (1937) 'Die Pollenerzeugung der Windbluter' Botanisch Centralblatt 56A pp365-470.

Pokorny, V. (1965) *Some palaeoecological problems in marine ostracode faunas, demonstrated on the Upper Cretaceous ostracodes of Bohemia, Czechslovakia* In: Puri, H.S. (ed.) Ostracods as ecological and palaeoecological indicators, Symposium on Ostracods, Zoological station, Naples (June).

Polack, B., Maarleveld, G.C. and Nota, D.J.G. (1962) 'Palynological and sedimentary petrological data of a section in ice-pushed deposits (Southern Veluwe, Netherlands)' *Geologie en Mijnbouw* **41** pp333-350.

Polunin, O. (1969) Flowers of Europe: A field guide, Oxford University Press, Oxford.

Polunin, O. (1980) Flowers of Greece and the Balkans: A field guide, Oxford University Press, Oxford.

R e f e r e n c e s

Pons, A. and Reille, M. (1988) ‘The Holocene and upper Pleistocene pollen record from Padul (Granada, Spain): a new study’ *Palaeogeography, Palaoeclimatology and Palaeoecology* **66** pp243-263.

Popa, I. and Kern, Z. (2009) ‘Long-term summer temperature reconstruction inferred from tree-ring records from the Eastern Carpathians’ *Climate Dynamics* **32** pp1107-1117.

Porter, S.C. (1986) ‘Pattern and forcing of Northern Hemisphere glacier variations during the last Millennium’ *Quaternary Research* **26** pp27–48.

de Prado, C. (1860) ‘Sur l'existence de la faune primordiale dans la chaîne cantabrique’ *PEP de Verneuil, Imprimerie de L. Martinet.*

Poska, A., Saarse, L. and Veski, S. (2004) ‘Reflections of pre- and early-agrarian human impact in the pollen diagrams of Estonia’ *Palaeogeography, Palaeoclimatology, Palaeoecology* **209** pp37-50.

Power, J. and Cooper, A. (1995) ‘Vegetation and land use change in North-Eastern Ireland’ *Landscape and Urban Planning* **31** pp195-203.

Prentice, I.C. (1985) ‘Pollen representation, source area, and basin size: toward a unified theory of pollen analysis’ *Quaternary Research* **23** pp76-86.

Prentice, C., Guiot, J., Huntley, B., Jolly, D., and Cheddadi, R. (1996) ‘Reconstructing biomes from palaeoecological data: a general method and its application to European pollen data at 0 and 6 ka’ *Climate Dynamics* **12** pp185-194.

Provansal, M., Vella, C., Arnaud-Fassetta, G., Sabatier, F. and Maillet, G. (2003) ‘Role of fluvial sediment inputs in the mobility of the Rhône delta coast (France)’ *Géomorphologie: relief, processus, environnement* **4** pp271-282.

Pruysers, P.A., de Lange, G.J. and Middelburg, J.L. (1991) ‘Geochemistry of Eastern Mediterranean sediments: Primary sediment composition and diagenetic alterations’ *Marine Geology* **100** pp137-154.

Punning, J.-M. and Koff, T. (1997) ‘The landscape factor in the formation of pollen records in lake sediments’ *Journal of Paleolimnology* **18** pp33-44.

Punning, J.-M., Koff, T., Sakson, M. And Terasmaa, J. (2004) ‘Human impact on the ecosystem of Lake Ruusmäe (Southern Estonia) traced in the sediments’ Polish *Journal of Ecology* **52** (3) pp285-299.

Punning, J.M., Terasmaa, J. and Vaasmam, T. (2006) ‘The impact of lake-level fluctuations on the sediment composition’ *Water, Air, and Soil Pollution: Focus* **6** pp151-157.

Pye, K. and Blott, S.J. (2004) ‘Particle size analysis of sediments, soils and related particulate materials for forensic purposes using laser granulometry’ *Forensic Science International* **144** pp19–27.

R e f e r e n c e s

Quigyu, G., Baotian, P., Na, L., Qiong, L., Jundi, Z., Hongshan, G. and Jia, L. (2010) 'A warming interval during the MIS 5a/4 transition in two high-resolution loess sections from China' *Journal of Asian Earth Sciences* **38** (6) pp255-261.

Quincey, D.J., Luckman, A. and Benn, D. (2009) 'Quantification of Everest region glacier velocities between 1992 and 2002, using satellite radar interferometry and feature tracking' *Journal of Glaciology* **55** (192) pp596-606.

Rabatel, A., Francou, B., Jomelli, V., Naveau, P. and Grancher, P. (2008) 'A chronology of the Little Ice Age in the tropical Andes of Bolivia (16°S) and its implications for climate reconstruction' *Quaternary Research* **70** (2) pp198-212.

Ranković, S., Radičević, D. and Soković-Ilić, G. (1981) Basic characteristics of the precipitation distribution in Yugoslavia - Contribution to the maps of climate Atlas **Note 2** Federal Hydrometeorological Institute of Yugoslavia, Belgrade, Yugoslavia pp226.

Rebolledo, L., Sepúlveda, J., Lange, C.B., Pantoja, S., Bertrand, S., Hughen, K. and Figueroa, D. (2008) 'Late Holocene marine productivity changes in Northern Patagonia-Chile inferred from a multi-proxy analysis of Jacaf channel sediments' *Estuarine, Coastal and Shelf Science* **80** (3) pp314-322.

Reed, J.M., Leng, M.J., Ryan, S., Black, S., Altinsacli, S. and Griffiths, H.I. (2008) 'Recent habitat degradation in karstic Lake Uluabat, Western Turkey: A coupled limnological-palaeolimnological approach' *Biological Conservation* **141** pp2765-2783.

Reeves, J.M., Chivas, A.R., Garcia, A. and De Deckker, P. (2007) 'Palaeoenvironmental change in the Gulf of Carpentaria (Australia) since the last interglacial based on Ostracode' *Palaeogeography, Palaeoclimatology, Palaeoecology* **246** pp163-187.

Reille, M. (1992) Pollen et spores d'Europe et d'Afrique du nord, Laboratoire de botanique historique et palynologie, Marseille.

Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C.J.H., Blackwell, P.G. and Buck, C.E. (2004) 'IntCal04 Terrestrial Radiocarbon Age Calibration, 0-26 Cal Kyr BP' *Radiocarbon* **46** pp1029-1058.

Reinemann, S.A., Porinchu, D.F., Bloom, A.M., Mark, B.G. and Box, J.E. (2009) 'A multi-proxy paleolimnological reconstruction of Holocene climate conditions in the Great Basin, United States' *Quaternary Research* **72** pp347-358.

Revel-Rolland, M., Arnaud, F., Chapron, E., Desmet, M., Givelet, N., Alibert, C. and McCulloch, M. (2005) 'Sr and Nd isotopes as tracers of clastic sources in Lake Le Bourget sediment (NW Alps, France) during the Little Ice Age: Palaeohydrology implications' *Chemical Geology* **224** pp183-200.

Revelli, P. (1911) 'Le fronti di 7 ghiacciai del versante italiano del Monte Bianco nel 1910' *Rivista del Club Alpino Italiano* **XXX** pp254-258.

Reynolds, C.S. (1980) 'Phytoplankton assemblages and their periodicity in stratifying lake systems' *Holarctic Ecology* **3** (3) pp141-159.

R e f e r e n c e s

Riera, S., Wansard, G. and Julià, R. (2004) ‘2000-year environmental history of a karstic lake in the Mediterranean Pre-Pyrenees: the Estanya lakes (Spain)’ *Catena* **55** pp293–324.

Rigozo, N.R., Echer, E., Vieira, L.E.A. and Nordemann, D.J.R. (2001) ‘Reconstruction of Wolf sunspot numbers on the basis of spectral characteristics and estimates of associated radio flux and solar wind parameters for the last millennium’ *Solar Physics* **203** pp179-191.

Rind, D., Lean, J. and Healy, R. (1999) ‘Simulated time-dependent climate response to solar radiative forcing since 1600’ *Journal of Geophysical Research* **104** (D2) pp1973-1990.

Ritchie, J.C. and McHenry, J.R. (1990) ‘Application of radioactive fallout Cesium-137 for measuring soil erosion and sediment accumulation rates and patterns: A review’ *Journal of Environmental Quality* **19** (2) pp.215-233.

Ritchie, J. C., McHenry, J. R. and Gill, A. C. (1973) ‘Dating recent reservoir sediments’ *Limnology and Oceanography* **18** pp254-263.

Ritchie, J. C., Sparberry, J. A. and McHenry, J. R. (1974) ‘Estimating soil erosion from the redistribution of fallout ^{137}Cs’ *Soil Science Society of America* **38** pp137-139.

Robbins, J.A., Krezoski, J.R. and Mozley, S.C. (1977) ‘Radioactivity in sediments of the great lakes: post-depositional redistribution by deposit-feeding organisms’ *Earth and Planetary Science Letters* **36** pp325-333.

Robbins, J. A., Edgington, D. N., and Kemp, A. L. W. (1978) ‘Comparative $^{21°}$Pb, ^{137}Cs, and pollen geochronologies of sediments from lakes Ontario and Erie’ *Quaternary Research* **10** pp256-278.

Roberts, N., Stevenson, A.C., Davis, B. and Cheddadi, S. (2004) ’Holocene climate, environment and cultural change in the circum-Mediterranean region’ *Conference on Climate Variability through Europe and Africa****,*** *Past Climate Variability through Europe and Africa* **6** pp343-362.

Robinson, G.W. (1922) ‘Note on the mechanical analysis of humus soils’ *Journal of Agricultural Science* **XII** pp287-291.

Rodrigo, F.S. and Barriendos, M. (2008) ‘Reconstruction of seasonal and annual rainfall variability in the Iberian peninsula (16th–20th centuries) from documentary data’ *Global and Planetary Change* **63** pp243-257.

Rodrigo, F.S., Esteban-Parra, M.J., Vázquez, P.D. and Castro-Diez, Y. (2000) ‘Rainfall variability in Southern Spain on decadal to centennial time scales’ *International Journal of Climatology* **20** pp721-732.

Rodrigo, F.S., Parra, M.J.E. and Diez, Y.C. (1995) Reconstruction of total annual rainfall in Andalusia (Southern Spain) during the 16th and 17th Centuries from documentary sources’ *Theoretical and Applied Climatology* **52** pp207-218.

Romera, G.G., Carrión, J.S., Pausas, J.G., Sevilla-Callejo, M., Lamb, H.F., Fernández, S. and Burjachs, F. (2010) ‘Holocene fire activity and vegetation response in South- Eastern Iberia’ *Quaternary Science Reviews* **29** pp1082-1092.

R e f e r e n c e s

Roucoux, K.H., Tzedakis, P.C., Frogley, M.R., Lawson, I.T. and Preece, R.C (2008) 'Vegetation history of the marine isotope stage 7 interglacial complex at Ioannina, NW Greece' *Quaternary Science Reviews* **27** (13-14) pp1378-1395.

Rovelli, A. (1984) 'Seismic *Q* for the lithosphere of the Montenegro region (Yugoslavia)" frequency, depth and time windowing effects' *Physics of the Earth and Planetary Interiors* 34 pp159-172.

Royall, D. (2001) 'Use of mineral magnetic measurements to investigate soil erosion and sediment delivery in a small agricultural catchment in limestone terrain' *Catena* **46** pp15-34.

Ruiz, F., González-Regalado, M.L., Muñoz, J.M., Pendón, J.G., Rodríguez-Ramírez, A., Cáceres, L. and Vidal, R. (2003) 'Population age structure techniques and ostracods: Applications in coastal hydrodynamics and palaeoenvironmental analysis' *Palaeogeography, Palaeoclimatology, Palaeoecology* **199** pp51-69.

Russell, H. (1908) Souvenirs d'un montaignard. Pau (Republished 1979, Slatkine, Genève).

Rutka, M.A. and Eyles, N. (1989) 'Ostracod faunas in a large Pleistocene ice-dammed lake, Southern Ontario, Canada' *Palaeogeography, Palaeoclimatology, Palaeoecology* **73** pp61-76.

Saarse, L., Niinemets, E., Poska, A. and Veski, S. (2010) 'Is there a relationship between crop farming and the Alnus decline in the Eastern Baltic region?' *Vegetation History and Archaeobotany* **19** pp17-28.

Sarmaja-Korjonen, K. (2003) 'Contemporaneous Alnus decline and the beginning of Iron Age cultivation in pollen stratigraphies from Southern Finland' *Vegetation History and Archaeobotany* **12** pp49-59.

Sarmaja-Korjonen, K., Seppänen, A. and Bennike, O. (2006) 'Pediastrum algae from the classic late glacial Bølling Sø site, Denmark: Response of aquatic biota to climate change' *Review of Palaeobotany and Palynology* **138** pp95-107.

Saint-Saud, A.A. (1893) 'Les Picos de Europa (Monts Cantabriques).Etude Orographique' *Annales Club Alpine Français* **38**.

Schaetzl, R.J. and Anderson, S. (2005) Soils: Genesis and Geomorphology, Cambridge University Press, Cambridge.

Schilman, B., Bar-Matthews, M., Almogi-Labin, A. and Luz, B. (2001) 'Global climate instability reflected by Eastern Mediterranean marine records during the late Holocene' *Palaeogeography, Palaeoclimatology, Palaeoecology* **176** pp157-176.

Schilman, B., Ayalon, A., Bar-Matthews, M., Kagan, E.J. and Almogi-Labin, A. (2002) 'Sea-land paleoclimate correlation in the Eastern Mediterranean region during the late Holocene' *Israel Journal of Earth Sciences* **51** (3-4) pp181-190 (abstract only).

Schimmel, J. and Granstrom, A. (1996) 'Fire Severity and Vegetation Response in the Boreal Swedish Forest' *Ecology* **77** (5) pp1436-1450.

References

Schönweise (1987) Information sources: Paleoclimate, Landolt-Börnstein, New Series V/4c1 pp95-96.

Schrader, F. (1894) 'Sur l'étendue des glaciers des Pyrenees' *Annuaire Club Alpine francais* **21** pp403-423.
Schrader, F. (1936) 'Sur 1'étendue des glaciers des Pyrenees 1894' *Biology Bulletin* **V70** pp201–21.

Schulte, L. (2002) 'Climatic and human influence on river systems and glacier fluctuations in Southeast Spain since the Last Glacial Maximum' *Quaternary International* **93-94** pp85-100.

Schwab, M.J., Neumann, F., Litt, T., Negendank, J.F.W., Stein, M. (2004) 'Holocene palaeoecology of the Golan Heights (Near East): investigation of lacustrine sediments from Birkat Ram crater lake' *Quaternary Science Reviews* **23** pp1723-1731.

Schweingruber, F.H. (1985) 'Dendro-ecological zones in the coniferous forests of Europe' *Dendrochronologia* **3** pp67-75.

Schweingruber, F.H. (1988) Tree rings: basics and applications of dendrochronology, Dordrecht, Reidel.

Schweingruber, F.H. (2008) 'Jahorina, PCAB - ITRDB YUGO004'
DOC/NOAA/NESDIS/NCDC (National Climatic Data Center, NESDIS, NOAA, Boulder, CO 80305, USA, Accessed at: http://www.ncdc.noaa.gov/paleo/metadata/noaa-tree-4444.html (05.05.09).

Schweingruber, F.H., Briffa, K.R. and Jones, P.D. (1991) 'Yearly maps of summer temperatures in Western Europe from A.D. 1750 to 1975 and Western North America from 1600 to 1982: Results of a radiodensitometrical study on tree rings' *Vegetation* **92** pp5-71.

Sergerstrom, U. (1991) 'Soil pollen analysis - An application for tracing ancient arable patches' *Journal of Archaeological Science* **18** pp165-175.

Serrano, E. and Martínez de Pisón, E. (1994) 'Geomorfología y evolución glaciar en el Pirineo aragonés oriental' In Martí, C.E., and García, J.M. (eds.) El glaciarismo surpirenaico, Nuevas aportaciones, Geoforma ediciones pp33–64.

Serrat, D. and Ventura, J. (1993) 'Glaciers of Europe – Glaciers of the Pyrenees Spain and France' *U.S. Geological Survey Professional Paper* **1386-E-2** ppE49-E61.

Sevenning, J.-C. (2002) 'A review of natural vegetation openness in North-Western Europe' *Biological Conservation* **104** pp133-148.

Sharp, M.J. (1985) 'Sharp, Sedimentation and stratigraphy at Eyjabakkajokull - an Icelandic surging glacier' *Quaternary Research* **24** pp268–284.

Sheppard, P.R. (2010) 'Dendroclimatology: extracting climate from trees' *Climate Change* **Focus article 1** pp343-352.

Shogenova, A. (1999) 'The influence of dolomitization on the magnetic properties of Lower Palaeozoic carbonate rocks in Estonia' *Geological Society, London, Special Publication* **151** pp167-180.

References

Shukhorukov, F.V., Gavshin, V.M., Malikova, I.N., Kovalev, S.I., Malikov, I.Y. and Romashkin, P.A. (2000) 'Cesium-137 in the environment of the Altay region (Russia)' *Water, Air, and Soil Pollution* **118** pp395-406.

Shu, G. and Collins, M. (2001) 'The use of grain size sediment trends in marine sediment dynamics: A review' *Chinese Journal of Oceanology and limnology* **19** (3) pp265-271.

Shuman, B. (2003) 'Controls on loss-on-ignition variation in cores from two shallow lakes in the Northeastern United States' *Journal of Paleolimnology* **30** pp371-385.

Sifrer M. (1963) 'New findings about the glaciation of Triglav' *Geografski zbornik* **8** pp157-210.

Sikorski, J. and Bluszcz, A. (2003) 'Testing applicability of ^{210}Pb method to date sediments of human-made Lake Kozlowa Góra' *Geochronometria* **22** pp63-66.

Simopoulos, S. (1989) 'Soil Sampling and ^{137}Cs Analysis of the Chernobyl Fallout in Greece' *Applied Radiation and Isotopes* **40** (7) pp607-613.

Sjögren, P., van der Knaap, W.O., Kaplan, J.O., van Leeuwen, J.F.N and Ammann, B. (2008) 'A pilot study on pollen representation of mountain valley vegetation in the central Alps' *Review of Palaeobotany and Palynology* **149** pp208–218.

Smirnov, N.N. (1964) 'On the quantity of allochthonous pollen and spores received by the Rybinsk Reservoir' *Hydrobiologia* **24** (1-2) pp421-429.

Smith, A.J. (1993) 'Lacustrine ostracodes as paleohydrochemical indicators in lakes of the North-central United States' *Journal of Paleolimnology* **8** pp121-134.

Smol, J.P. (2002) Pollution of lakes and rivers–a paleoenvironmental perspective, Arnold, London.

Sohar, K. and Meidla, T. (2010) 'Changes in the Early Holocene lacustrine environment inferred from the subfossil ostracod record in the Varangu section, Northern Estonia' *Estonian Journal of Earth Sciences* **59** (3) pp195-206.

Storti, F. and Balsamo, F. (2009) 'Particle size distributions by laser diffraction – Part 1: Sensitivity of granular matter strength to analytical operating procedures' *Solid Earth Discuss* **1** pp93–141.

Stothers, R. B. (1984) 'The great Tambora eruption in 1815 and its aftermath' *Science* **224** (4654) pp1191-1198.

Sousa, A. and Garcia-Murillo, P. (2003) 'Changes in the wetlands of Andalusia (Doñana National Park, SW Spain) at the end of the Little Ice Age' *Climatic Change* **58** pp193-217.

Sperazza, M., Moore, J.N. and Hendrix, M.S. (2004) 'High resolution particle size analysis of naturally occurring very fine-grained sediment through laser diffractometry' *Journal of Sedimentary Research* **74** (5) pp736-743.

Srivastava, S.K. (1994) 'Evolution of Cretaceous phytogeoprovinces, continents and climate' *Review of Palaeobotany and Palynology* **82** pp197-224.

R e f e r e n c e s

Stefanova, I., Atanassova, J., Delcheva, M. and Wright, H.E. (2006a) ‘Chronological framework for the Lateglacial pollen and macrofossil sequence in the Pirin Mountains, Bulgaria: Lake Besbog and Lake Kremensko-5’ *The Holocene* **16** (6) pp877-892.

Stefanova, V., Lazarova, M. and Wright, H.E. (2006b) ‘Elevational gradients during the Late-Glacial/Holocene vegetational transition in Southern Bulgaria’ *Vegetation History and Archaeobotany* **15** pp333-343.

Stevens, L. R., Wright, H.E. and Ito, E. (2001) ‘Proposed changes in seasonality of climate during the Lateglacial and Holocene at Lake Zeribar, Iran’ *The Holocene* **11** pp747-755.

Stine, S. (1990) ‘Late Holocene fluctuations of Mono Lake, Eastern California’ *Palaeogeography, Palaeoclimatology, Palaeoecology* **78** pp 333-381.

Stotter, M. (1846) Die Gletscher des Vernagtthales in Tirol und ihre Geschichte, Wagner, Innsbruck.

Stuiver, M. and Braziunas, T.F. (1995) *Evidence of solar variations* In: Bradley, R.S. and Jones P.D. (eds.) Climate since A.D. 1500, Routledge, Chapman and Hall, London.

Stuiver, M. and Reimer, P.J. (1993) ‘Extended ^{14}C database and revised CALIB radiocarbon calibration program’ *Radiocarbon* **35** pp215-230.

Stuiver, M., Reimer, P.J. and Braziunas, T.F. (1998) ‘High-precision radiocarbon age calibration for terrestrial and marine samples’ *Radiocarbon* **40** pp1127-1151.

Sturm, M. (1975) Depositional and erosional sedimentary features in a turbidity current controlled basin (Lake Brienz) Proceedings of the IXth International Congress of Sedimentology, Nice.

Subally, D. and Quézel, P. (2002) ‘Glacial or interglacial: Artemisia, a plant indicator with dual responses’ *Review of Palaeobotany and Palynology* **120** pp123-130.

Sugita, S. (1993) ‘A model of pollen source area for an entire lake surface’ *Quaternary Research* **39** pp239-244.

Sutherland, T.F., Lane, P.M., Amos, C.L. and Downing, J. (2000) ‘The calibration of optical backscatter sensors for suspended sediment of varying darkness levels’ *Marine Geology* **162** pp587-597.

Sutton, 0.G. (1947) ‘The problem of diffusion in the lower atmosphere’ *Quarterly Journal of the Royal Meteorological Society* **73** pp257-276.

Syvitski, J.P.M., Leblanc, K.W.G., Asprey, K.W. (1991) *Interlaboratory, interinstrument calibration experiment* In: Syvitski, J.P.M. (ed.) Principles, Methods and Application of Particle Size Analysis, Cambridge University Press, Cambridge.

Taberlet, P. and Cheddadi, R. (2002) ‘Quaternary Refugia and Persistence of Biodiversity’ *Science* **297** pp2009-2010.

Tanarro, L. M., Hoelzle, M., García, A., Ramos, M., Gruber, S., Gómez, A., Montserrat, P. and Palacios, D. (2001) 'Permafrost distribution modelling in the mountains of the Mediterranean:

Corral del Veleta, Sierra Nevada, Spain' *Norsk Geografisk Tidsskrift (Norwegian Journal of Geography)* **55** (4) pp253-260.

Tanner, P.A., Pan, S.M., Mao, S.Y. and Yu, K.N. (2000) 'γ-Ray spectrometric and α-counting method comparison for the determination of Pb-210 in estuarine sediments' *Applied Spectroscopy* **54** (10) pp1443-1446.

Tauber, J. (1965) 'Differential pollen dispersion and the interpretation of pollen diagrams' *Danmarks Geologiske Undersøgelse Series II* **89** pp1-89.

Tauber, J. (1967) 'Investigations of the mode of pollen transfer in forested areas' *Review of Palaeobotany and Palynology* **3** (1-4) pp277-286.

Tchillingarian, G. (1952) 'Study of dispersing agents' *Journal of Sedimentary Petrology* **22** (4) pp229-233.

Tessier, L., Guibal, F. and Schweingruber, F.H. (1997) 'Research strategies in dendroecology and dendroclimatology in mountain environments' *Climate Change* **36** pp499-517.

Theocharopoulos, M., Georgiadis, T., Dimitrllos, G., Chochliouros and Tiniakou, A. (2006) 'Vegetation types with *Cladium mariscus (Cyperaceae)* in Greece' *Willdenowia* **36** pp247-256.

Thompson, R., Battarbee, R.W., O'Sullivan, P.E. and Oldfield, F. (1975) 'Magnetic susceptibility of lake sediments' *Limnology and Oceanography* **20** (5) pp687-698.

Thompson, R. and Morton, D.J. (1979) 'Magnetic susceptibility and particle-size distribution in Recent sediments of the Loch Lomond drainage basin, Scotland' *Journal of Sedimentary Research* **49** (3) pp801-811.

Thompson, R. and Oldfield, F. (1986) Environmental Magnetism, Allen & Unwin, London.

Thompson Davis, P. (1988) 'Holocene glacier fluctuations in the American cordillera' *Quaternary Science Reviews* **7** pp129-157.

Thorndycraft, V.R., Benito, G., Walling, D.E., Sopeña, A., Sánchez-Moya, Y., Rico, M. and Casas, A. (2005) 'Caesium-137 dating applied to slackwater flood deposits of the Llobregat River, NE Spain' *Catena* **59** pp305-318.

Thornes, J., López-Bermúdez and Woodward, J. (2009) *Hydrology, river regimes, and sediment yield* In: Wooward, J. (ed.) The Physical Geography of the Mediterranean, Oxford university press, London.

Thorp, J.H. and Covich, A.P. (2009) Ecology and Classification of North American Freshwater Invertebrates, 3rd Edition, Academic Press, New York.

Tipping, R. (1986) 'A late-Devensian pollen site in Cowal, South-West Scotland' *Scottish Journal of Geology* **22** pp27-40.

Tipping, R. (1989) 'Lomond re-advance in the Awe valley and adjacent palynological evidence for the extent of the Loch Lomond Readvance in the Awe valley and adjacent areas, SE Highlands' *Scottish Journal of Geology* **25** pp325-337.

R e f e r e n c e s

Tobin M. J. and Schell W. R. (1988) 'Recent developments sedimentation modelling and the statistical reliability of ^{210}Pb dating method' Material of Trace Models in Lakes Conference, The Master University Canada, Ontario pp1-23.

Tonkov, S., Panovska, H., Possnert, G. and Bozilova, E. (2002) 'The Holocene vegetation history Northern Pirin Mountain, SouthWestern Bulgaria: pollen analysis and radiocarbon dating of a core from Lake Ribno Banderishko' *The Holocene* **12** (2) pp201-210.

Tonkov, S., Possnert, G. and Bozilova, E. (2006) 'The lateglacial vegetation and radiocarbon dating of Lake Trilistnika, Rila Moutains (Bulgaria)' *Vegetation History and Archaeobotany* **16** pp15-22.

Tošić, I. (2004) 'Spatial and temporal variability of winter and summer precipitation over Serbia and Montenegro' *Theoretical and Applied Climatology* **77** pp47-56.

Touchan, R., Akkemik, U., Hughes, M.K. and Erkan, N. *(*2007) 'May–June precipitation reconstruction of SouthWestern Anatolia, Turkey during the last 900 years from tree rings' *Quaternary Research* **68** pp196-202.

Trigo, R.M., Osborn, T.J. and Corte-Real, J.M. (2002) 'The North Atlantic Oscillation influence on Europe: climate impacts and associated physical mechanisms' *Climate Research* **20** pp9-17.

Tröels-Smith, J. (1955) 'Karakterisering af løse jordarter' *Danmarks Geologiske Undersøgelse Series IV* **3** (10) pp73.

Trouet, V., Esper, J., Graham, N.E., Baker, A., Scourse, J.D. and Frank, D.C. (2009) 'Persistent positive North Atlantic Oscillation Mode dominated the Medieval climate anomaly' *Science* **324** pp78-80.

Trutat, E. (1875) *'Les glaciers de la Maladeta et le Pic de Posets' Annuaire de Club Alpine Français* **12** pp440.

Tsonis, A.A., Swanson, K.L., Sugihara, G. and Tsonis, P.A. (2010) 'Climate change and the demise of Minoan civilization' *Climate of the Past* **6** pp525-530.

Tucker, M.E. (2003) Sedimentary Rocks in the Field (Geological Field Guide), 4th edition, Wiley-Blackwell, London.

Turner, J. and Hodgson, J. (1981) 'Studies in the vegetational history of the Northern Pennines.II. An atypical pollen diagram from Pow Hill, Co. Durham' *Journal of Ecology* **69** pp171-188.

Tutiempo (2009) Climate data for Kolašin, Serbia Montenegro, Accessed: http://www.tutiempo.net/en/Climate/KOLASIN/134650.htm (14.09.09).

Tzedakis, P.C. (1993) 'Long-term tree populations in Northwest Greece through multiple Quaternary climatic cycles' *Nature* **364** pp437-440.

Tzedakis, P.C. (1999) 'The last climatic cycle at Kopais, central Greece' *Journal of the Geological Society* **156** pp425-434.

R e f e r e n c e s

Tzedakis, P.C., Andrieu, V., de Beaulieu, J.L., Crowhurst, S., Follieri, M., Hooghiemstra, H., Magri, D., Reille, M., Sadori, L., Shackleton, N.J. and Wijmstra, T.A. (1997) 'Comparison of terrestrial and marine records of changing climate of the last 500,000 years' *Earth and Planetary Science Letters* **150** pp171-176.

Tzedakis, P.C., Lawson, I.T., Frogley, M.R., Hewitt, G.M. and Preece, R.C. (2002) 'Buffered Tree Population Changes in a Quaternary Refugium: Evolutionary Implications' *Science* **297** pp2044-2047.

Tzedakis, P.C., Frogley, M.R. and Heaton, T.H.E. (2003) 'Last Interglacial conditions in Southern Europe: evidence from Ioannina, Northwest Greece' *Global and Planetary Change* **36** pp157–170.

Tzedakis, P.C., Frogley, M.R., Lawson, I.T., Preece, R.C., Cacho, I. and de Abreu, L. (2004a) 'Ecological thresholds and patterns of millennial-scale climate variability: The response of vegetation in Greece during the last glacial period' *Geology* **32** (2) pp109-112.

Tzedakis, P.C., Roucoux, K.H., de Abreu, L. and Shackleton, N.J. (2004b) 'The Duration of forest stages in Southern Europe and interglacial climate variability' *Science* **306** pp2231-2235.

Unkašević, M. and Radinović, Đ. (2000) 'Statistical analysis of daily maximum and monthly precipitation at Belgrade' *Theoretical and Applied Climatology* **66** pp241-249.

van den Brink, L.M. and Janssen, C.R. (1985) 'The effect of human activities during cultural phases on the development of montane vegetation in the Serra de Estrela, Portugal' *Review of Palaeobotany and Palynology* **44** pp193-215.

Valero-Garcés, B.L., González-Sampériz, P., Navas, A., Machín, J., Mata, P., Delgado-Huertas, A., Bao, R., Moreno, A., Carrión, J.S., Schwalb, A. and González-Barrios, A. (2006) 'Human impact since medieval times and recent ecological restoration in a Mediterranean lake: the Laguna Zoñar, Southern Spain' *Journal of Paleolimnology* **35** pp441-465.

Vallot, J. (1887) 'Oscillations des glaciers Pyrenees' *Études Pyrénéennes* **3** pp1-6.

van der Linden, M. and van Geel, B. (2006) 'Late Holocene climate change and human impact recorded in a South Swedish ombrotrophic peat bog' *Palaeogeography, Palaeoclimatology, Palaeoecology* **240** pp649-667.

van der Linden, M., Vickery, E., Charman, D.J. and van Geel, B. (2008) 'Effects of human impact and climate change during the last 350 years recorded in a Swedish raised bog deposit' *Palaeogeography, Palaeoclimatology, Palaeoecology* **262** pp1–31.

van Welden , A., Beck, C., Reyan, J.-L., *et al.*, Bushati, S., Koci, R., Jouanne, F. and Mugnier, J.-L. (2008) 'The last 500 year of sedimentation in Shkodra Lake (Albania/Montenegro): paleoenvironmental evolution and potential for paleoseismicity studies' *Journal of Paleolimnology* **40** pp619-633.

van der Wiel, A.M. and Wijmstra, T.A. (1997) 'Palynology of the 112.8-197.8 m interval of the core Tenaghi Philippon III, Middle Pleistocene of Macedonia' *Review of Palaeobotany and Palynology* **52** pp89-117.

R e f e r e n c e s

Vannière, B., Bossuet, G., Walter-Simonnet, A.-V., Gauthier, E., Barral, P., Petit, C., Buatier, M. and Daubigney, A. (2003) 'Land use change, soil erosion and alluvial dynamic in the lower Doubs Valley over the 1st millennium AD (Neublans, Jura, France)' *Journal of Archaeological Science* **30** (10) pp1283-1299.

Vannier, J. and Abe, K. (1995) 'Size, body plan and respiration in the Ostracoda' *Palaeontology* **38** (4) pp843-873.

Velez, M.I., Wille, M., Hooghiemstra, H., Metcalfe, S., Vandenberghe, J. and van der Borg, K. (2001) 'Late Holocene environmental history of Southern Chocó regions, Pacifica Colombia; sediment, diatom and pollen analysis of core El Caimito' *Palaeogeography, Palaeoclimatology, Palaeoecology* **173** pp197-214.

Velo, E., Paparisto, A., Bongiorno, G., Di Muccio, T., Khoury, C., Bino, S., Gramiccia, M., Gradoni, L. and Maroli, M. (2005) 'Entomological and parasitological study on phlebotomine sandflies in central and Northern Albania' *Parasite* **12** pp45-49.

Velusček, A. (1999) *Mlajša kamena in bakrena doba* In: Aubelj, B. (ed.) Zakladi tisočletij, Ljubljana, Modrijan, pp52-75.

Veski, S. (1994) 'Stratigraphy of Holocene *Pediastrum* taxa from the sediments of Lake Maardu North Estonia' *Proceedings of the Estonian Academy of Sciences, Geology* **43** (1) pp46-54.

Vidic, N.J., Singer, M.J. and Verosub, K.L. (2004) 'Duration dependence of magnetic susceptibility enhancement in the Chinese loess–palaeosols of the past 620 ky' *Palaeogeography, Palaeoclimatology, Palaeoecology* **211** (3-4) pp271-288.

Vincent, C.(2002) 'Influence of climate change over the 20th Century on four French glacier mass balances' *Journal of Geophysical Research* **107 ppACL1-12.**

Viti, M., D'Onza, F., Mantovani, E., Albarello, D. and Cenni, N. (2003) 'Post-seismic relaxation and earthquake triggering in the Southern Adriatic region' *Geophysical Journal International* **153** pp645-657.

Vivian, R. (1975) Les Glaciers des Alpes Occidentales, Allier, Grenoble.

von Post, L. (1916) 'Om skogstradpolleni sydsvenskat orvmosslagerfdljder' *Geologiska Foreningens I Stockholm Forhandlingar* **38** pp384-94.

Waas, M. (1995) *Proposed Northern course of the Rother: A sedimentological and miscrofaunal investigation* In: Eddison, J. (ed.) Romney Marsh: the debatable ground, **Monograph 41**, Oxford University Committee for Archaeology, Oxford, pp51-77.

Waele, J.G., Martina, M.L.V., Sanna, L., Cabras, S. and Cossu, Q.A. (2010) 'Flash flood hydrology in karstic terrain: Flumineddu Canyon, central-East Sardinia' *Geomorphology* **120** pp162-173.

Wagner, B., Sulpizio, R., Zanchetta, G., Wulf, S., Wessels, M., Daut, G. and Nowaczyk, N. (2008) 'The last 40 ka tephrostratigraphic record of Lake Ohrid, Albania and Macedonia: A very distal archive for ash dispersal from Italian volcanoes' *Journal of Volcanology and Geothermal Research* **177** pp71-80.

R e f e r e n c e s

Wagner, S. and Zorita, E. (2005) 'The influence of volcanic, solar and CO2 forcing on the temperatures in the Dalton Minimum (1790–1830): a model study)' *Climate Dynamics* **25** pp205-218.

Walker, G.T. and Bliss, E.W. (1932) 'World weather' *Memorials of the Royal Meteorological Society* **4** (36) pp53-84.

Wang,H., Hong, Y., Lin, O., Hong, B., Zhu, Y., Wang, Y. and Xu, H. (2010) 'Response of humification degree to monsoon climate during the Holocene from the Hongyuan peat bog, Eastern Tibetan Plateau' *Palaeogeography, Palaeoclimatology, Palaeoecology* **286** pp171-177.

Wanner, H., Holzhauser, H., Pfister, C. and Zumbuhl, H.J. (2000) 'Interannual to century scale climate variability in the European Alps' *Erdkunde* **54** pp62-69.

Wantzen, K.M., Junk, W.J. and Rothhaupt, K.O. (2008) 'An extension of the floodpulse concept (FPC) for lakes' *Hydrobiologia* **613** pp151-170.

Watson, C.S. (1996) 'The vegetational history of the Northern Apennines, Italy: Information from three new sequences and a review of regional vegetational change' *Journal of Biogeography* **23** (6) pp805-841.

Webb, T. (2001) *Paleoecology* In: Levin, S.A. (ed.) Encyclopedia of Biodiversity Vol. 4 Academic Press, New York.

Weckström, K., Weckström, J., Yliniemi, L.-M. and Korhola, A. (2010) 'The ecology of Pediastrum (Chlorophyceae) in subarctic lakes and their potential as paleobioindicators' *Journal of Paleolimnology* **43** pp61-73

Weiss, E.L. and Frock, H.N. (1976) 'Rapid Analysis of Particle Size Distributions by Laser Light Scattering' *Powder Technology* **14** pp287-293.

Weiss, H., Courty, M.-A., Wetterstrom, W., Guichard, F., Senior, L., Meadow, R. and Curnow, A. (1993) 'The Genesis and Collapse of Third Millennium North Mesopotamian Civilization' *Science* **261** pp995-1004.

Wentworth, C.K. (1922) 'A scale of grade and class terms for clastic sediments' The *Journal of Geology* **30** (5) pp377-392.

West, R.G. (1956) 'The Quaternary deposits at Hoxne, Suffolk' *Philosophical Transactions of the Royal Society of London Series B* **239** 665 pp265-356.

West, R.G. (1964) 'Inter-Relations of Ecology and Quaternary Palaeobotany' *Journal of Animal Ecology* **33** pp47-57.

Whatley, R. (1988) *Ostracoda and palaeogeography* In: De Deckker, P., Colin, J.-P. and Peypouquet, J.-P. (eds.) Ostracoda in the Earth Science, Elsevier, Amsterdam pp103-123.

Wick, L., Lemcke, G. and Sturm, M. (2003) **'**Evidence of Lateglacial and Holocene climatic change and human impact in Eastern Anatolia: high-resolution pollen, charcoal, isotopic and geochemical records from the laminated sediments of Lake Van, Turkey' *The Holocene* **13** (5) pp665-675.

Wijmstra, T.A. (1969) ‘Palynology of the first 30 metres of a 120-m deep section in Northern Greece’ *Acta Botanica Neerlandica* **18** (4) pp511-527.

Wijmstra, T.A. and Smit, A. (1976) ‘Palynology of the middle part (30–78 metres) of the 120 m deep section in Northern Greece (Macedonia)’ *Acta Botanica Neerlandica* **25** (4) pp297– 312.

Williams, H.F.L. and Roberts, M.C. (1990) ‘Two Middle Holocene marker beds in vertically accreted floodplain deposits, Lower Fraser River, British Columbia’ *Géographie physique et Quaternaire* **44** (1) pp27-32.

Williams, T., Thouveny, N. and Creer, K.M. (1996) ‘Palaeoclimatic significance of the 300 ka mineral magnetic record from the sediments of Lac du Bouchet, France’ *Quaternary Science Reviews* **15** (2-3) pp223-235.

Williamson, D., Thouveny, N., Hillaire-Marcel, C., Mondeguer, A., Taieb, M., Tiercelin, J.-J. and Vincens, A. (1991) ‘Chronological potential of palaeomagnetic oscillations recorded in late Quaternary sediments from Lake Tanganyika’ *Quaternary Science Reviews* **10** pp351-361.

Willis, K.J. and Bennett, K.D. (1994) ‘The Neolithic transition - fact or fiction? Palaeoecological evidence from the Balkans’ *The Holocene* **4** pp326-330.

Wilmhurst, J.M. and McGlone, M.S. (2005a) ‘Origin of pollen and spores in surface lake sediments: Comparison of modern palynomorph assemblages in moss cushions, surface soils and surface lake sediments’ *Review of Palaeobotany and Palynology* **136** pp1-15.

Wilmhurst, J.M. and McGlone, M.S. (2005b) ‘Corroded pollen and spores as indicators of changing lake sediment sources and catchment disturbance’ *Journal of Paleolimnology* **34** pp503–517.

Wilson, R.J.S., Esper, J. and Luckman, B.H. (2004) ‘Utilising historical tree-ring data for dendroclimatology: A case study from the Bavarian Forest, German’ *Dendrochronologia* **21** (2) pp53-68.

Winkler, S. (2003) 'A new interpretation of the date of the ‘Little Ice Age’ glacier maximum at Svartisen and Okstindan, Northern Norway' *The Holocene* **13** pp83-95.

Winkler, S. and Matthews, J.A. (2010) ‘Observations on terminal moraine-ridge formation during recent advances of Southern Norwegian glaciers’ *Geomorphology* **116** pp87-106.

Wohlfarth, B. (1996) ‘The chronology of the last termination: A review of radiocarbon-dated, high-resolution terrestrial stratigraphies’ *Quaternary Science Reviews* **15** pp267-284.

Wood, F.B. (1988) ‘Alpine Glacier Trends, 1960s to 1980s’ *Arctic and Alpine Research* **20** (4) pp404-413.

Wright Jr., H.E. (1967) ‘The use of surface samples in Quaternary pollen analysis’ *Review of Palaeobotany and Palynology* **2** pp321-330.

Wulf, S., Kraml, M., Kuhn, T., Schwarz, M., Inthron, M., Keller, J., Kuscu, I. and Halbach, P. (2002) ‘Marine tephra from the Cape Riva eruption (22 ka) of Santorini in the Sea of Marmara’ *Marine Geology* **183** pp131-141.

References

Wyrwoll, K.-H. and Miller, G.H. (2001) 'Initiation of the Australian summer monsoon 14,000 years ago' *Quaternary International* **83-85** pp119-128.

Xiang, L., Lu, X.X., Higgitt, D.L. and Wang, S.M. (2002) 'Recent lake sedimentation in the middle and lower Yangtze basin inferred from ^{137}Cs and ^{210}Pb measurements' *Journal of Asian Earth Sciences* **21** pp77-86.

Xoplaki, E., Maheras, P. and Luterbacher, J. (2001) 'Variability of climate in meridional Balkans during the periods 1675-1715 and 1780-1830 and its impact on human life' *Climate Change* **48** pp581-615.

Yey, S.-W. and Kim, C.-H. (2010) 'Recent warming in the Yellow/East China Sea during winter and the associated atmospheric circulation' *Continental Shelf Research* **30** pp1428-1434.

Zabenskie, S., Peros, M. and Gajewski, K. (2006) 'The use of heavy liquid in the separation of pollen from Arctic lake sediments' *Canadian Association of Palynologists Newsletter* **29** (2) pp5-7.

Zahno, C., Akcar, N., Yavuz, V., Kubik, P.W. and Schlüchter, C. (2009) 'Surface exposure dating of Late Pleistocene glaciations at the Dedegöl Mountains (Lake Beyşehir, SW Turkey)' *Journal of Quaternary Science* **24** (8) pp1016-1028.

Zetzsche, F. (1932) *Kork und cuticularsubstanzen* In: Klein, G. (ed.) Handbuch der Pflanzen-analyse, Springer-Verlag, Berlin.

Zhao, Y. and Sun, Q.F. (2010) 'Reliability of pollen concentration as the indicator of effective moisture in arid and semi-arid regions of China' *Journal of Arid Environments* **74** pp423–427.

Zielinski, G.A. and Germani, M.S. (1998) 'New ice-core evidence challenges the 1620sBC age for the Thera Santorini (Minoan) Eruption' *Journal of Archaeological Science* **25** pp279-289.

APPENDIX I

Method used to create the dendrochronology record for Jahorina and summer temperatures in the Southern European region, Figure 2.3 (A,B), Chap. 2.

Figure 2.3 A

The dendrochronological data for Jahorina, Bosnia Herzegovina (166km North West of Lake Plav), presented in Section 2.2.2 Dendroclimatological records from Europe, formed part of a larger body of work published by Schweingruber *et al.* (1988; 1991). De-trended maximum ring density data for up to 23 samples of *Picea abies* at the site was accessed at the NOAA website - http://www.ncdc.noaa.gov/paleo/metadata/noaa-tree-4444.html. This was inputted into excel, and a 10 year running mean was then applied to the data to provide an assessment of the general trends represented. However, as the data covered a 245 year period (AD 1736 - 1981) a 10 year running mean could not be used for all of the data. Therefore, for the years AD 1977 – 1981 a declining mean was used, for example:

- AD 1971 for the 10 year running mean the data from AD1967-1976 was averaged
- In contrast the mean for AD 1980 was reduced to averaging the years AD 1979 – 1981
- The mean for the final date AD 1981 used the data entry recorded for this year.

Figure 2.3 B

The Southern European data used formed the basis of Briffa *et al.* (2001), but the associated data (referenced as Briffa *et al.*, 2002) was accessed via the NOAA website - ftp://ftp.ncdc.noaa.gov/pub/data/paleo/treering/reconstructions/n_hem_temp/briffa2001jgr2.txt. A similar method to that presented above was used to produce graph B, but the locations of the chronologies used within the data set provided by Briffa *et al.* (2002) can be seen in Figure 1.

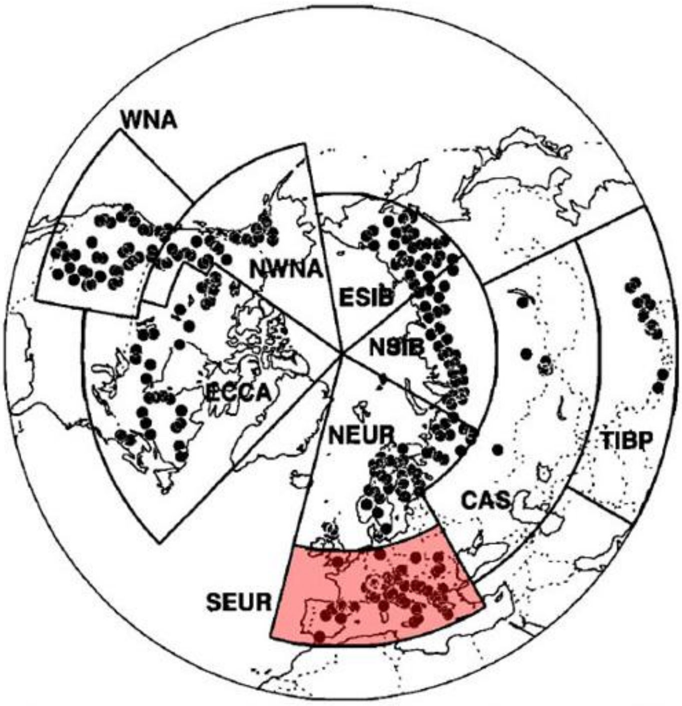
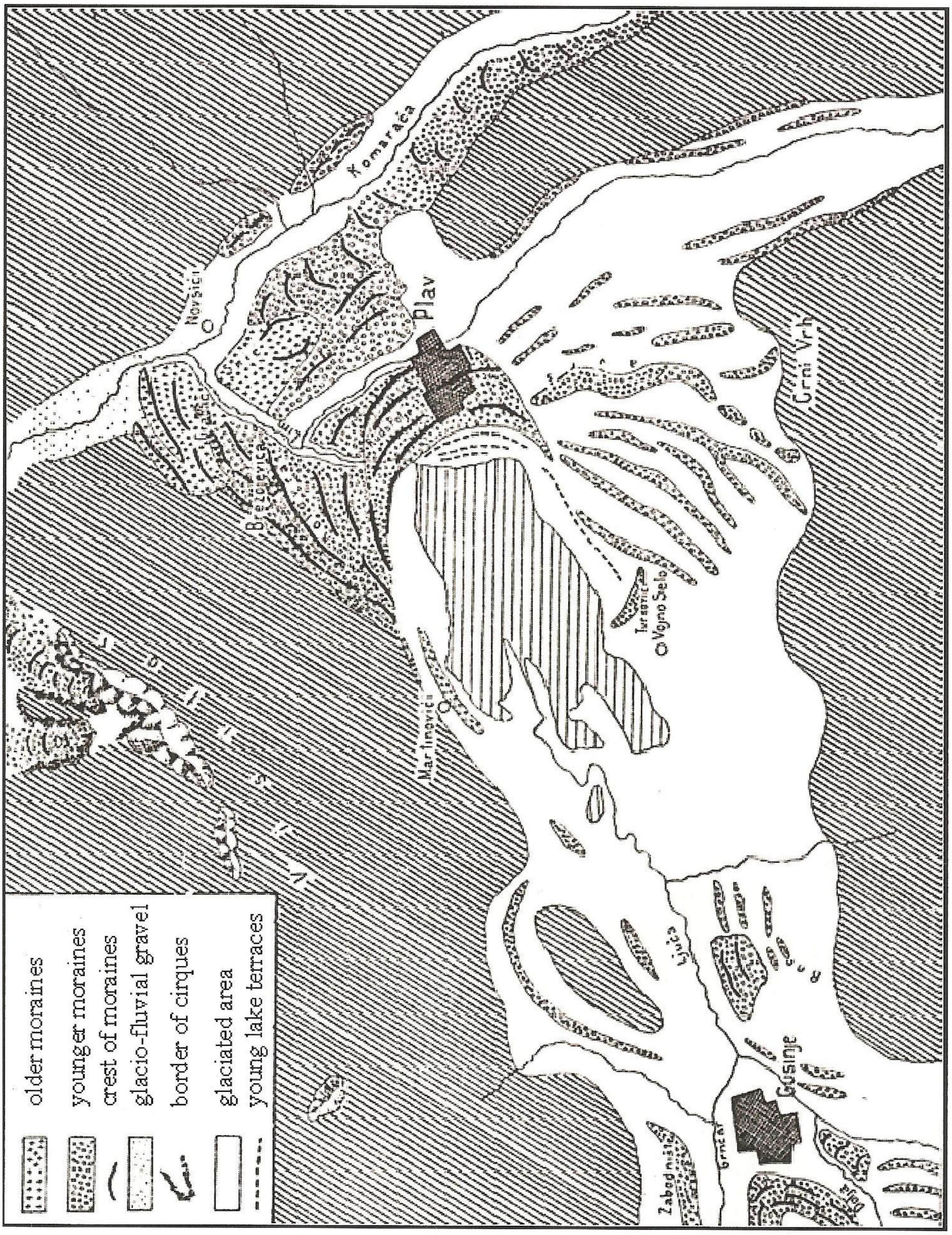


Figure 1: The locations of tree ring density chronologies used by Briffa *et al.* (2001) to complete northern hemisphere summer (Apr. - Sep.) temperature reconstructions. Data from the Southern European region (SEUR), highlighted in red, was used to produce Figure 2.3 B.

APPENDIX II

Original Cvijić map, published AD 1913.



APPENDIX III

Tröels-Smith Classification

Class	**Symbol**	**Composition**	**Description**
Turfa	Tb^{0-4}	T. bryophytica	- Mosses +/- humous
	Tl	T. lignose	- Stumps, roots/intertwined rootlets of ligneous plants
	Th	T. herbacea	- Roots, intertwined rootlets of herbaceous plants
Detritus	Dl^{0-4}	D. lignosus	- Fragments of ligneous plants larger than 2 mm
	Dh	D. herbosus	- Fragments of herbaceous plants larger than 2 mm
	Dg	D. granosus	- Fragments of Dl, Dh with animal fossils 0.1 mm to 2 mm
Limnus	Ld^{0-4}	L. detrituosus	- Plant and animal fragments +/- humous
	Lso	L. siliceous organogenes	- Diatoms, siliceous skeletons, particles less than 1 mm
	Lc	L. calcareus	
	Lf	L. ferrugineus	- Iron oxide
Argilla	As^{0-4}	Clay	-Particles < 2 µm
	Ag	Silt	- Particles 2 – 63 µm
Grana	Ga^{0-4}	Fine sand	- Particles 63 – 250 µm
	Gs	Medium to coarse sand	- Particles 250 µm - 2 mm
	Gg	Gravel	- Particles > 2mm
Substantia humosa	Sh^{0-4}	Humous substance	- Disintegrated organic substances, precipitated humic acids/staining

Adapted from Birks and Briks (1980).

APPENDIX IV

Pollen preparations

Kidson and Williams (1969) sodium diphosphate ($Na_4P_2O_7$) preparation.

Suspend sediments in $Na_4P_2O_7$three to five times and centrifuge off the finer sediments after each wash. Decant sample on the final suspension and pass through a coarse sieve to remove sand fractions and/or large organics. Then follow the standard Faegri and Iversen (1975) preparation, but following acetolysis the pass sample through a fine 7 µm mesh to concentrate the pollen grains and spores. Wash samples from the mesh and apply standard dehydration and mounting techniques applied.

Problems associated with the preparation:

Cwynar *et al.* (1979) tested this method and noted that by sieving using a 7µm mesh 0.4% pollen grains were lost and that large grains such as Cyperaceae and *Picea* were absent from filtrates. Furthermore, the inherent difficulty in using such fine mesh during sample preparation is that it can easily become clogged, making the method less efficient. The samples tested by Cwynar *et al.* (1979) included marine and deep freshwater lake clays, late-glacial clays and sub-arctic lake sediments. The greatest concern in using this preparation is the mesh size with Cwynar *et al.* (1979) stating that pollen types such as *Picea* and Cyperaceae are absent from samples, rendering such a small mesh useless for samples that may contain grains larger than 7 µm i.e. likely to be far more than the 0.4% quoted by Cwynar *et al.* (1979) for most sites.

Sodium Polytungstate (SPT) pollen preparation method.

A 0.6 ml volume sample of sediment was measured and placed in a large numbered centrifuge tube. Approximately 5 ml 10%v/v Hydrochloric Acid was added, followed by a single *Lycopodium* tablet. After the reaction had settled approximately 10 ml 10% v/v Potassium Hydroxide was added before placing the samples into a water bath for 40 minutes, stirring occasionally. The samples were removed from the water-bath and centrifuged 4500rpm for 3 minutes. Each sampled was agitated before being washed through a 125 micron sieve down a funnel and into a clean, large centrifuge tube. The sieve was then rinsed through with distilled water to maximise the amount of pollen being washed into the second tube. Distilled water was added to each sample to balance before centrifuging.

SPT was used to separate the pollen from minerogenic material, following Zabenskie and Gajewski (2006). 3 ml of SPT with specific gravity of ***c.*** 1.95 g/cm^3 was added to each the sample and agitated using a vortex mixer. The centrifuge tubes were then balanced using deionised water and **c**entrifuge at 1800rpm for 10 minutes. The supernatant was decanted into clean labelled centrifuge tubes and this step repeated twice to ensure any remaining pollen grains were removed from the sample. Each centrifuge tube was then filled with as much deionised water as possible, to ensure the density of the SPT/water solution decreases and the pollen sinks when centrifuged. To produce a solid pollen pellet samples were centrifuged at 4500rpm once more.

Acetolysis was then performed to remove any remaining organic material, by adding was 5ml glacial acetic before centrifuging. The samples were agitated, before adding 5ml of acetolysis mixture (1 part Concentrated Sulphuric Acid, 9 parts Acetic Anhydride), and placed in a water bath for 2 minutes. The samples were then centrifuged before Glacial Acetic Acid, followed by 10% Potassium Hydroxide were added to neutralise the acetolysis mixture and acid. Samples were then washed in distilled water prior to dehydration by Ethanol. The final stage was the addition of 1 ml tert-Butyl alcohol and silicone oil. Each sample was covered with a small cotton wool stopper to allow the remaining tert-Butyl alcohol to evaporate off overnight.

APPENDIX V

Standard Faegri and Iversen (1975) pollen preparation

Samples were measured by volume displacement using a 10ml measuring cylinder and then placed in a large centrifuge tube. To remove calcium carbonate material from the samples an excess of dilute hydrochloric acid (10% HCl) was added and each sample was spiked using a lycopodium exotic marker grain (13,400 spores/tablet). Removal of any humic acids was completed with the addition of 10ml of potassium hydroxide (10% KOH) and placing samples in a water bath at 80^{o}C for 40 minutes. Removal of large organic or mineral fragments was ensured by passing samples through a 125µm sieve. Each sample was then centrifuged and the supernatant decanted until it became clear. Before HF could be added to the samples, they were mixed thoroughly with an excess of dilute HCl. Between 6ml and 8ml of HF (40%) was added to each sample and heated in a water bath at 80^{o}C for a minimum of 60 minutes. The samples were then centrifuge and the supernatant decanted, before a further 6ml to 8ml of HF was added. Samples were then left in cold HF for at least 2 days to ensure all siliceous material was removed. Once this time had elapsed samples were centrifuged and the supernatant decanted, if any siliceous material remained, felt by running a polythene stirring rod across the base of the test tube, a further treatment of hot HF was completed. 5ml dilute HCl removed any of the remaining HF before samples were centrifuged and the supernatant decanted. A glacial acetic acid wash ensured each sample was dehydrated prior to acetolysis and any remaining cellulose was removed by acetolysis. 5ml of acetolysis mixture (9 parts acetic anhydride, 1 part concentrated sulphuric acid) was added to each sample and heated in a water bath at 80^{o}C for 2 minutes, stirring frequently, before being centrifuged and decanted. Samples were then washed, firstly in 5ml glacial acetic acid to remove any remaining acetolysis mixture and then 5ml KOH to neutralize the acid, samples were centrifuged and decanted after each addition of acid or alkali. An excess of de-ionised water was then used to wash the samples before alcohol dehydration began by washing in 5ml Ethanol. Each sample was then agitated into suspension using 1ml of tert-butyl alcohol (TBA) and a vortex mixer, before being transferred into a labelled glass sample tube with a Pasteur pipette. The samples were then centrifuged and excess TBA was decanted off before adding silicone oil to preserve the pollen. Any remaining TBA was allowed to evaporate off, as small pieces of cotton wool were used as stoppers for around 24hrs.

APPENDIX VI

Method used for radiocarbon analysis, following Beta Analytic Inc.

1. Sample was gently dispersed in de-ionised water and then washed in hot HCL and NaOH consecutively to remove carbonates and secondary organic acids, with a final HCl wash to neutralize the NaOH.

2. The AMS results were then produced from the reduction of sample carbon to graphite (100% carbon), alongside standards and background analyses.

3. The graphite produced from the sample was then detected for ^{14}C content in one of nine accelerator mass spectrometers (Beta Analytic Inc. *pers. comm.*).

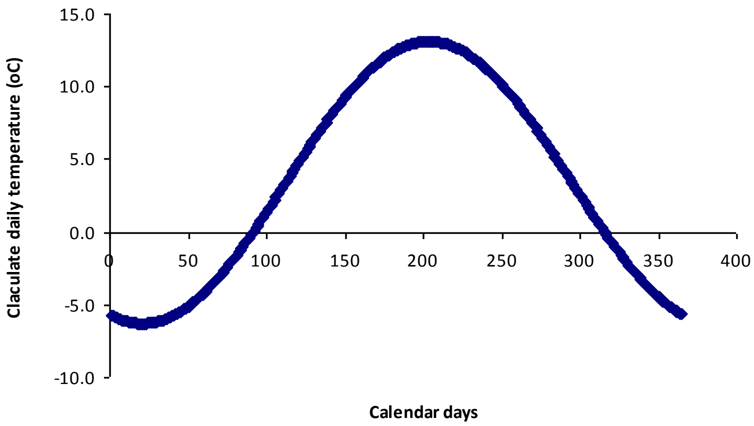
APPENDIX VII

Data used for Degree day modelling.

Full monthly climate data for Kolašin AD 1973 - 2008

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sept	Oct	Nov	Dec
1973				4.30	10.30	16.50	13.50		17.40		-2.60	
1974		1.00	8.90	6.50	9.20		17.50		15.20		5.00	1.40
1975		-5.50	2.20	5.80	14.00	14.90	17.70	16.60	16.90	7.80	2.80	0.70
1976	-0.80	-0.50	1.00	7.60	12.10	14.50	16.40	12.00	11.80	9.60	3.50	1.00
1977	3.50	3.70	5.30	8.50	12.60	14.60	18.50	16.90	12.20	8.70	5.00	-2.20
1978	0.50	1.10	3.90	6.10	10.20	16.10	17.20	16.60	11.20	8.90	1.80	1.70
1979	-1.80	1.40	5.10	5.30	13.40	16.80	16.50	14.80	13.00	9.00	4.90	2.80
1980	-3.80	0.20	2.50	4.80	9.70	15.30	17.00	17.30	13.60	9.20	5.20	-2.20
1981	-6.10	-1.40	5.00	7.20	11.80	16.20	16.40	14.70	13.90	9.70	-0.60	-0.60
1982	0.00	-2.10	1.30	5.90	18.20	14.00		16.60	15.70	11.40	2.20	2.30
1983	-2.70											
1984	no data	no data	no data	no data	no data	no data	no data	no data	no data	no data	no data	no data
1985												2.90
1986	0.60	-1.90			17.20			17.70	11.30	7.50		-1.80
1987	2.00			5.60	11.60							
1988	-3.50		1.30		15.30	20.30	21.60	19.00	14.60	10.40	-1.90	-2.60
1989	-1.50	3.00	6.10	8.40	9.80	13.90	17.10	16.80	13.60	7.50	1.10	0.70
1990	-1.50	2.30	5.60	7.30	13.60	17.40	17.60	17.50	13.00	9.60	6.80	-2.20
1991	-3.60	-2.20	6.60	6.10	8.30	16.30	17.20	18.00	14.50	8.80	4.90	-3.90
1992	-1.50	-0.10	2.70	6.60	13.50	15.20	17.30	21.20	13.80	10.50	5.80	-3.20
1993	-0.50	-1.80	1.00	7.90	13.00	16.30	18.20	18.90	13.80	11.30	4.30	2.70
1994	1.00	0.40	6.10	7.30	13.60	15.70	18.60	18.20	15.70	9.90	5.30	0.00
1995	-2.90	4.00	3.10	5.30	11.90	16.00	19.70	16.90	11.80	9.30	1.30	3.30
1996	0.40	-1.80	2.30	6.90	13.90	16.50	17.00	17.00	11.50	8.40	5.70	0.50
1997	1.30	0.90	3.30	3.00	12.90	17.50	17.40	16.40	13.50	6.50	5.30	0.40
1998	-0.50	0.90	2.00	8.10	12.00	18.00	18.90	17.60	13.60	10.10	6.30	-3.70
1999	-1.40	-1.30	3.60	8.20	15.30	16.00	19.10	19.50		13.50	3.60	0.80
2000	-5.30	-1.10	3.00	10.40						6.10	7.40	0.70
2001		1.50	4.90	6.90	13.70	14.90	18.90	20.20	13.30	10.90	2.20	-4.10
2002	-2.30	3.40	5.70	7.80	14.20	17.20	18.70	17.80	10.70	8.90	6.20	1.50
2003	0.20	-4.80	2.00	6.00	15.50	19.10	18.90	19.50	12.80	8.10	6.40	-0.70
2004	-2.00	1.80	3.00	9.10	10.30	16.60	18.10	17.60	14.10	11.60	1.90	1.20
2005	-2.40	-4.10	1.30	7.40	13.20	16.10	18.20	16.90	14.30	8.90	5.30	0.00
2006	-4.20	-1.50	1.90	9.20	13.10	15.70	18.60	17.10	14.20	10.70	2.90	-0.30
2007	1.90	3.20	5.40	10.30	14.00	18.00	20.00	19.50	12.00	8.60	1.40	-1.30
2008	0.00	1.50	4.30	8.20	13.80	17.50	18.60	19.30	13.10	10.40	5.30	5.60
Average	**-1.2**	**0.0**	**3.7**	**7.0**	**12.9**	**16.3**	**17.9**	**18.1**	**14.0**	**9.7**	**3.8**	**0.0**
Annual range		**19.4**										

AD 1832 Moraine: SINE CURVE 1
Amplitude of annual range (1/2 annual range) = **9.7°C**
Mean annual temperature = **3.42 °C**
Annual melt (mm w.e.) = **7291**



*Full data for '**SINE CURVE 1'** above and degree day factor of 4mm day^{-1} K^{-1}, following Braithwaite et al. 2006.*

	Day	Daily temperature mean	Degree-day factor	Daily melt
January 1st	1	-5.7	4	0.0
	2	-5.8	4	0.0
	3	-5.8	4	0.0
	4	-5.9	4	0.0
	5	-5.9	4	0.0
	6	-6.0	4	0.0
	7	-6.0	4	0.0
	8	-6.0	4	0.0
	9	-6.1	4	0.0
	10	-6.1	4	0.0
	11	-6.1	4	0.0
	12	-6.2	4	0.0
	13	-6.2	4	0.0
	14	-6.2	4	0.0
	15	-6.2	4	0.0
	16	-6.2	4	0.0
	17	-6.3	4	0.0
	18	-6.3	4	0.0
	19	-6.3	4	0.0
	20	-6.3	4	0.0
	21	-6.3	4	0.0
	22	-6.3	4	0.0
	23	-6.3	4	0.0
	24	-6.3	4	0.0
	25	-6.3	4	0.0
	26	-6.2	4	0.0
	27	-6.2	4	0.0
	28	-6.2	4	0.0
	29	-6.2	4	0.0
	30	-6.2	4	0.0
	31	-6.1	4	0.0
	32	-6.1	4	0.0
	33	-6.1	4	0.0
	34	-6.0	4	0.0
	35	-6.0	4	0.0
	36	-6.0	4	0.0
	37	-5.9	4	0.0
	38	-5.9	4	0.0
	39	-5.8	4	0.0
	40	-5.8	4	0.0
	41	-5.7	4	0.0
	42	-5.6	4	0.0
	43	-5.6	4	0.0
	44	-5.5	4	0.0
	45	-5.5	4	0.0
	46	-5.4	4	0.0
	47	-5.3	4	0.0
	48	-5.2	4	0.0
	49	-5.2	4	0.0
	50	-5.1	4	0.0
	51	-5.0	4	0.0
	52	-4.9	4	0.0
	53	-4.8	4	0.0
	54	-4.7	4	0.0
	55	-4.7	4	0.0
	56	-4.6	4	0.0
	57	-4.5	4	0.0
	58	-4.4	4	0.0
	59	-4.3	4	0.0
	60	-4.2	4	0.0

APPENDIX VII

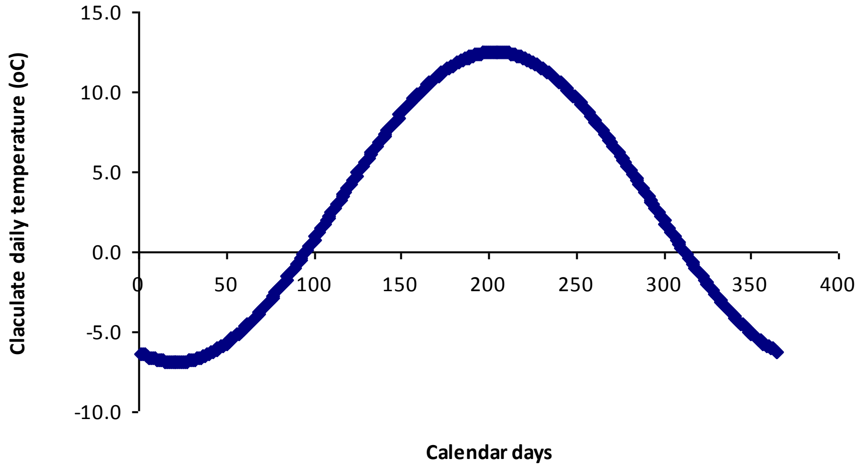
Day	Daily temperature mean	Degree-day factor	Daily melt	Day	Daily temperature mean	Degree-day factor	Daily melt
61	-4.1	4	0.0	140	7.9	4	31.6
62	-4.0	4	0.0	141	8.0	4	32.1
63	-3.8	4	0.0	142	8.2	4	32.7
64	-3.7	4	0.0	143	8.3	4	33.3
65	-3.6	4	0.0	144	8.5	4	33.9
66	-3.5	4	0.0	145	8.6	4	34.4
67	-3.4	4	0.0	146	8.8	4	35.0
68	-3.3	4	0.0	147	8.9	4	35.6
69	-3.1	4	0.0	148	9.0	4	36.1
70	-3.0	4	0.0	149	9.2	4	36.7
71	-2.9	4	0.0	150	9.3	4	37.2
72	-2.8	4	0.0	151	9.4	4	37.7
73	-2.6	4	0.0	152	9.6	4	38.2
74	-2.5	4	0.0	153	9.7	4	38.7
75	-2.4	4	0.0	154	9.8	4	39.3
76	-2.2	4	0.0	155	9.9	4	39.8
77	-2.1	4	0.0	156	10.1	4	40.2
78	-2.0	4	0.0	157	10.2	4	40.7
79	-1.8	4	0.0	158	10.3	4	41.2
80	-1.7	4	0.0	159	10.4	4	41.7
81	-1.5	4	0.0	160	10.5	4	42.1
82	-1.4	4	0.0	161	10.6	4	42.6
83	-1.2	4	0.0	162	10.8	4	43.0
84	-1.1	4	0.0	163	10.9	4	43.4
85	-1.0	4	0.0	164	11.0	4	43.9
86	-0.8	4	0.0	165	11.1	4	44.3
87	-0.7	4	0.0	166	11.2	4	44.7
88	-0.5	4	0.0	167	11.3	4	45.1
89	-0.3	4	0.0	168	11.4	4	45.5
90	-0.2	4	0.0	169	11.5	4	45.9
91	0.0	4	0.0	170	11.6	4	46.2
92	0.1	4	0.5	171	11.6	4	46.6
93	0.3	4	1.1	172	11.7	4	46.9
94	0.4	4	1.7	173	11.8	4	47.3
95	0.6	4	2.4	174	11.9	4	47.6
96	0.8	4	3.0	175	12.0	4	47.9
97	0.9	4	3.7	176	12.1	4	48.2
98	1.1	4	4.3	177	12.1	4	48.5
99	1.2	4	5.0	178	12.2	4	48.8
100	1.4	4	5.6	179	12.3	4	49.1
101	1.6	4	6.3	180	12.3	4	49.4
102	1.7	4	6.9	181	12.4	4	49.6
103	1.9	4	7.6	182	12.5	4	49.9
104	2.1	4	8.2	183	12.5	4	50.1
105	2.2	4	8.9	184	12.6	4	50.3
106	2.4	4	9.6	185	12.6	4	50.5
107	2.6	4	10.2	186	12.7	4	50.7
108	2.7	4	10.9	187	12.7	4	50.9
109	2.9	4	11.6	188	12.8	4	51.1
110	3.1	4	12.2	189	12.8	4	51.3
111	3.2	4	12.9	190	12.9	4	51.4
112	3.4	4	13.6	191	12.9	4	51.6
113	3.6	4	14.2	192	12.9	4	51.7
114	3.7	4	14.9	193	13.0	4	51.9
115	3.9	4	15.6	194	13.0	4	52.0
116	4.1	4	16.2	195	13.0	4	52.1
117	4.2	4	16.9	196	13.0	4	52.2
118	4.4	4	17.6	197	13.1	4	52.2
119	4.6	4	18.2	198	13.1	4	52.3
120	4.7	4	18.9	199	13.1	4	52.4
121	4.9	4	19.5	200	13.1	4	52.4
122	5.1	4	20.2	201	13.1	4	52.4
123	5.2	4	20.9	202	13.1	4	52.5
124	5.4	4	21.5	203	13.1	4	52.5
125	5.5	4	22.2	204	13.1	4	52.5
126	5.7	4	22.8	205	13.1	4	52.5
127	5.9	4	23.5	206	13.1	4	52.4
128	6.0	4	24.1	207	13.1	4	52.4
129	6.2	4	24.8	208	13.1	4	52.4
130	6.3	4	25.4	209	13.1	4	52.3
131	6.5	4	26.0	210	13.1	4	52.2
132	6.7	4	26.7	211	13.0	4	52.2
133	6.8	4	27.3	212	13.0	4	52.1
134	7.0	4	27.9	213	13.0	4	52.0
135	7.1	4	28.5	214	13.0	4	51.8
136	7.3	4	29.1	215	12.9	4	51.7
137	7.4	4	29.8	216	12.9	4	51.6
138	7.6	4	30.4	217	12.9	4	51.4

APPENDIX VII

Day	Daily temperature mean	Degree-day factor	Daily melt
219	12.8	4	51.1
220	12.7	4	50.9
221	12.7	4	50.7
222	12.6	4	50.5
223	12.6	4	50.3
224	12.5	4	50.1
225	12.5	4	49.8
226	12.4	4	49.6
227	12.3	4	49.3
228	12.3	4	49.1
229	12.2	4	48.8
230	12.1	4	48.5
231	12.1	4	48.2
232	12.0	4	47.9
233	11.9	4	47.6
234	11.8	4	47.2
235	11.7	4	46.9
236	11.6	4	46.6
237	11.6	4	46.2
238	11.5	4	45.8
239	11.4	4	45.5
240	11.3	4	45.1
241	11.2	4	44.7
242	11.1	4	44.3
243	11.0	4	43.8
244	10.9	4	43.4
245	10.7	4	43.0
246	10.6	4	42.5
247	10.5	4	42.1
248	10.4	4	41.6
249	10.3	4	41.2
250	10.2	4	40.7
251	10.1	4	40.2
252	9.9	4	39.7
253	9.8	4	39.2
254	9.7	4	38.7
255	9.6	4	38.2
256	9.4	4	37.7
257	9.3	4	37.2
258	9.2	4	36.6
259	9.0	4	36.1
260	8.9	4	35.5
261	8.7	4	35.0
262	8.6	4	34.4
263	8.5	4	33.8
264	8.3	4	33.3
265	8.2	4	32.7
266	8.0	4	32.1
267	7.9	4	31.5
268	7.7	4	30.9
269	7.6	4	30.3
270	7.4	4	29.7
271	7.3	4	29.1
272	7.1	4	28.5
273	7.0	4	27.9
274	6.8	4	27.2
275	6.7	4	26.6
276	6.5	4	26.0
277	6.3	4	25.4
278	6.2	4	24.7
279	6.0	4	24.1
280	5.9	4	23.4
281	5.7	4	22.8
282	5.5	4	22.1
283	5.4	4	21.5
284	5.2	4	20.8
285	5.0	4	20.2
286	4.9	4	19.5
287	4.7	4	18.8
288	4.5	4	18.2
289	4.4	4	17.5
290	4.2	4	16.9
291	4.0	4	16.2
292	3.9	4	15.5
293	3.7	4	14.9
294	3.5	4	14.2
295	3.4	4	13.5
296	3.2	4	12.9
297	3.0	4	12.2

Day	Daily temperature mean	Degree-day factor	Daily melt
298	2.9	4	11.5
299	2.7	4	10.9
300	2.5	4	10.2
301	2.4	4	9.5
302	2.2	4	8.9
303	2.1	4	8.2
304	1.9	4	7.5
305	1.7	4	6.9
306	1.6	4	6.2
307	1.4	4	5.6
308	1.2	4	4.9
309	1.1	4	4.3
310	0.9	4	3.6
311	0.7	4	3.0
312	0.6	4	2.3
313	0.4	4	1.7
314	0.3	4	1.1
315	0.1	4	0.4
316	0.0	4	0.0
317	-0.2	4	0.0
318	-0.4	4	0.0
319	-0.5	4	0.0
320	-0.7	4	0.0
321	-0.8	4	0.0
322	-1.0	4	0.0
323	-1.1	4	0.0
324	-1.3	4	0.0
325	-1.4	4	0.0
326	-1.5	4	0.0
327	-1.7	4	0.0
328	-1.8	4	0.0
329	-2.0	4	0.0
330	-2.1	4	0.0
331	-2.2	4	0.0
332	-2.4	4	0.0
333	-2.5	4	0.0
334	-2.6	4	0.0
335	-2.8	4	0.0
336	-2.9	4	0.0
337	-3.0	4	0.0
338	-3.1	4	0.0
339	-3.3	4	0.0
340	-3.4	4	0.0
341	-3.5	4	0.0
342	-3.6	4	0.0
343	-3.7	4	0.0
344	-3.8	4	0.0
345	-4.0	4	0.0
346	-4.1	4	0.0
347	-4.2	4	0.0
348	-4.3	4	0.0
349	-4.4	4	0.0
350	-4.5	4	0.0
351	-4.6	4	0.0
352	-4.7	4	0.0
353	-4.8	4	0.0
354	-4.8	4	0.0
355	-4.9	4	0.0
356	-5.0	4	0.0
357	-5.1	4	0.0
358	-5.2	4	0.0
359	-5.2	4	0.0
360	-5.3	4	0.0
361	-5.4	4	0.0
362	-5.5	4	0.0
363	-5.5	4	0.0
364	-5.6	4	0.0
December 31st 365	-5.7	4	0.0

AD 1907 Moraine: SINE CURVE 2
Amplitude of annual range (1/2 annual range) = **9.7°C**
Mean annual temperature = **2.82 °C**
Annual melt (mm w.e.) = **6762**



*Full data for '**SINE CURVE 2'** above and degree day factor of 4mm day^{-1} K^{-1}, following Braithwaite et al. 2006.*

	Day	Daily temperature mean	Degree-day factor	Daily melt
January 1st	1	-6.3	4	0.0
	2	-6.4	4	0.0
	3	-6.4	4	0.0
	4	-6.5	4	0.0
	5	-6.5	4	0.0
	6	-6.6	4	0.0
	7	-6.6	4	0.0
	8	-6.6	4	0.0
	9	-6.7	4	0.0
	10	-6.7	4	0.0
	11	-6.7	4	0.0
	12	-6.8	4	0.0
	13	-6.8	4	0.0
	14	-6.8	4	0.0
	15	-6.8	4	0.0
	16	-6.8	4	0.0
	17	-6.9	4	0.0
	18	-6.9	4	0.0
	19	-6.9	4	0.0
	20	-6.9	4	0.0
	21	-6.9	4	0.0
	22	-6.9	4	0.0
	23	-6.9	4	0.0
	24	-6.9	4	0.0
	25	-6.9	4	0.0
	26	-6.8	4	0.0
	27	-6.8	4	0.0
	28	-6.8	4	0.0
	29	-6.8	4	0.0
	30	-6.8	4	0.0

Day	Daily temperature mean	Degree-day factor	Daily melt
31	-6.7	4	0.0
32	-6.7	4	0.0
33	-6.7	4	0.0
34	-6.6	4	0.0
35	-6.6	4	0.0
36	-6.6	4	0.0
37	-6.5	4	0.0
38	-6.5	4	0.0
39	-6.4	4	0.0
40	-6.4	4	0.0
41	-6.3	4	0.0
42	-6.2	4	0.0
43	-6.2	4	0.0
44	-6.1	4	0.0
45	-6.1	4	0.0
46	-6.0	4	0.0
47	-5.9	4	0.0
48	-5.8	4	0.0
49	-5.8	4	0.0
50	-5.7	4	0.0
51	-5.6	4	0.0
52	-5.5	4	0.0
53	-5.4	4	0.0
54	-5.3	4	0.0
55	-5.3	4	0.0
56	-5.2	4	0.0
57	-5.1	4	0.0
58	-5.0	4	0.0
59	-4.9	4	0.0
60	-4.8	4	0.0
61	-4.7	4	0.0
62	-4.6	4	0.0

APPENDIX VII

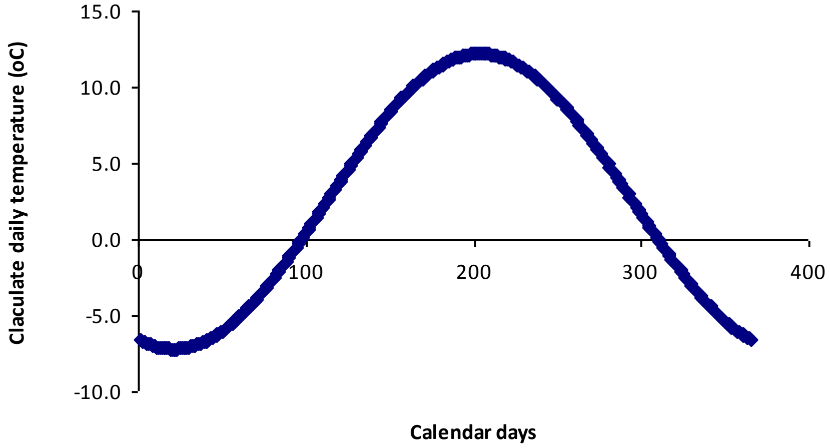
Day	Daily temperature mean	Degree-day factor	Daily melt
63	-4.4	4	0.0
64	-4.3	4	0.0
65	-4.2	4	0.0
66	-4.1	4	0.0
67	-4.0	4	0.0
68	-3.9	4	0.0
69	-3.7	4	0.0
70	-3.6	4	0.0
71	-3.5	4	0.0
72	-3.4	4	0.0
73	-3.2	4	0.0
74	-3.1	4	0.0
75	-3.0	4	0.0
76	-2.8	4	0.0
77	-2.7	4	0.0
78	-2.6	4	0.0
79	-2.4	4	0.0
80	-2.3	4	0.0
81	-2.1	4	0.0
82	-2.0	4	0.0
83	-1.8	4	0.0
84	-1.7	4	0.0
85	-1.6	4	0.0
86	-1.4	4	0.0
87	-1.3	4	0.0
88	-1.1	4	0.0
89	-0.9	4	0.0
90	-0.8	4	0.0
91	-0.6	4	0.0
92	-0.5	4	0.0
93	-0.3	4	0.0
94	-0.2	4	0.0
95	0.0	4	0.0
96	0.2	4	0.6
97	0.3	4	1.3
98	0.5	4	1.9
99	0.6	4	2.6
100	0.8	4	3.2
101	1.0	4	3.9
102	1.1	4	4.5
103	1.3	4	5.2
104	1.5	4	5.8
105	1.6	4	6.5
106	1.8	4	7.2
107	2.0	4	7.8
108	2.1	4	8.5
109	2.3	4	9.2
110	2.5	4	9.8
111	2.6	4	10.5
112	2.8	4	11.2
113	3.0	4	11.8
114	3.1	4	12.5
115	3.3	4	13.2
116	3.5	4	13.8
117	3.6	4	14.5
118	3.8	4	15.2
119	4.0	4	15.8
120	4.1	4	16.5
121	4.3	4	17.1
122	4.5	4	17.8
123	4.6	4	18.5
124	4.8	4	19.1
125	4.9	4	19.8
126	5.1	4	20.4
127	5.3	4	21.1
128	5.4	4	21.7
129	5.6	4	22.4
130	5.7	4	23.0
131	5.9	4	23.6
132	6.1	4	24.3
133	6.2	4	24.9
134	6.4	4	25.5
135	6.5	4	26.1
136	6.7	4	26.7
137	6.8	4	27.4
138	7.0	4	28.0
139	7.1	4	28.6
140	7.3	4	29.2
141	7.4	4	29.7
142	7.6	4	30.3
143	7.7	4	30.9
144	7.9	4	31.5
145	8.0	4	32.0
146	8.2	4	32.6
147	8.3	4	33.2
148	8.4	4	33.7
149	8.6	4	34.3
150	8.7	4	34.8
151	8.8	4	35.3
152	9.0	4	35.8
153	9.1	4	36.3
154	9.2	4	36.9
155	9.3	4	37.4
156	9.5	4	37.8
157	9.6	4	38.3
158	9.7	4	38.8
159	9.8	4	39.3
160	9.9	4	39.7
161	10.0	4	40.2
162	10.2	4	40.6
163	10.3	4	41.0
164	10.4	4	41.5
165	10.5	4	41.9
166	10.6	4	42.3
167	10.7	4	42.7
168	10.8	4	43.1
169	10.9	4	43.5
170	11.0	4	43.8
171	11.0	4	44.2
172	11.1	4	44.5
173	11.2	4	44.9
174	11.3	4	45.2
175	11.4	4	45.5
176	11.5	4	45.8
177	11.5	4	46.1
178	11.6	4	46.4
179	11.7	4	46.7
180	11.7	4	47.0
181	11.8	4	47.2
182	11.9	4	47.5
183	11.9	4	47.7
184	12.0	4	47.9
185	12.0	4	48.1
186	12.1	4	48.3
187	12.1	4	48.5
188	12.2	4	48.7
189	12.2	4	48.9
190	12.3	4	49.0
191	12.3	4	49.2
192	12.3	4	49.3
193	12.4	4	49.5
194	12.4	4	49.6
195	12.4	4	49.7
196	12.4	4	49.8
197	12.5	4	49.8
198	12.5	4	49.9
199	12.5	4	50.0
200	12.5	4	50.0
201	12.5	4	50.0
202	12.5	4	50.1
203	12.5	4	50.1
204	12.5	4	50.1
205	12.5	4	50.1
206	12.5	4	50.0
207	12.5	4	50.0
208	12.5	4	50.0
209	12.5	4	49.9
210	12.5	4	49.8
211	12.4	4	49.8
212	12.4	4	49.7
213	12.4	4	49.6
214	12.4	4	49.4
215	12.3	4	49.3
216	12.3	4	49.2
217	12.3	4	49.0
218	12.2	4	48.9

APPENDIX VII

Day	Daily temperature mean	Degree-day factor	Daily melt
219	12.2	4	48.7
220	12.1	4	48.5
221	12.1	4	48.3
222	12.0	4	48.1
223	12.0	4	47.9
224	11.9	4	47.7
225	11.9	4	47.4
226	11.8	4	47.2
227	11.7	4	46.9
228	11.7	4	46.7
229	11.6	4	46.4
230	11.5	4	46.1
231	11.5	4	45.8
232	11.4	4	45.5
233	11.3	4	45.2
234	11.2	4	44.8
235	11.1	4	44.5
236	11.0	4	44.2
237	11.0	4	43.8
238	10.9	4	43.4
239	10.8	4	43.1
240	10.7	4	42.7
241	10.6	4	42.3
242	10.5	4	41.9
243	10.4	4	41.4
244	10.3	4	41.0
245	10.1	4	40.6
246	10.0	4	40.1
247	9.9	4	39.7
248	9.8	4	39.2
249	9.7	4	38.8
250	9.6	4	38.3
251	9.5	4	37.8
252	9.3	4	37.3
253	9.2	4	36.8
254	9.1	4	36.3
255	9.0	4	35.8
256	8.8	4	35.3
257	8.7	4	34.8
258	8.6	4	34.2
259	8.4	4	33.7
260	8.3	4	33.1
261	8.1	4	32.6
262	8.0	4	32.0
263	7.9	4	31.4
264	7.7	4	30.9
265	7.6	4	30.3
266	7.4	4	29.7
267	7.3	4	29.1
268	7.1	4	28.5
269	7.0	4	27.9
270	6.8	4	27.3
271	6.7	4	26.7
272	6.5	4	26.1
273	6.4	4	25.5
274	6.2	4	24.8
275	6.1	4	24.2
276	5.9	4	23.6
277	5.7	4	23.0
278	5.6	4	22.3
279	5.4	4	21.7
280	5.3	4	21.0
281	5.1	4	20.4
282	4.9	4	19.7
283	4.8	4	19.1
284	4.6	4	18.4
285	4.4	4	17.8
286	4.3	4	17.1
287	4.1	4	16.4
288	3.9	4	15.8
289	3.8	4	15.1
290	3.6	4	14.5
291	3.4	4	13.8
292	3.3	4	13.1
293	3.1	4	12.5

Day	Daily temperature mean	Degree-day factor	Daily melt
294	2.9	4	11.8
295	2.8	4	11.1
296	2.6	4	10.5
297	2.4	4	9.8
298	2.3	4	9.1
299	2.1	4	8.5
300	1.9	4	7.8
301	1.8	4	7.1
302	1.6	4	6.5
303	1.5	4	5.8
304	1.3	4	5.1
305	1.1	4	4.5
306	1.0	4	3.8
307	0.8	4	3.2
308	0.6	4	2.5
309	0.5	4	1.9
310	0.3	4	1.2
311	0.1	4	0.6
312	0.0	4	0.0
313	-0.2	4	0.0
314	-0.3	4	0.0
315	-0.5	4	0.0
316	-0.6	4	0.0
317	-0.8	4	0.0
318	-1.0	4	0.0
319	-1.1	4	0.0
320	-1.3	4	0.0
321	-1.4	4	0.0
322	-1.6	4	0.0
323	-1.7	4	0.0
324	-1.9	4	0.0
325	-2.0	4	0.0
326	-2.1	4	0.0
327	-2.3	4	0.0
328	-2.4	4	0.0
329	-2.6	4	0.0
330	-2.7	4	0.0
331	-2.8	4	0.0
332	-3.0	4	0.0
333	-3.1	4	0.0
334	-3.2	4	0.0
335	-3.4	4	0.0
336	-3.5	4	0.0
337	-3.6	4	0.0
338	-3.7	4	0.0
339	-3.9	4	0.0
340	-4.0	4	0.0
341	-4.1	4	0.0
342	-4.2	4	0.0
343	-4.3	4	0.0
344	-4.4	4	0.0
345	-4.6	4	0.0
346	-4.7	4	0.0
347	-4.8	4	0.0
348	-4.9	4	0.0
349	-5.0	4	0.0
350	-5.1	4	0.0
351	-5.2	4	0.0
352	-5.3	4	0.0
353	-5.4	4	0.0
354	-5.4	4	0.0
355	-5.5	4	0.0
356	-5.6	4	0.0
357	-5.7	4	0.0
358	-5.8	4	0.0
359	-5.8	4	0.0
360	-5.9	4	0.0
361	-6.0	4	0.0
362	-6.1	4	0.0
363	-6.1	4	0.0
364	-6.2	4	0.0
December 31st 365	-6.3	4	0.0

AD 1975 and 2009 Moraine: SINE CURVE 3
Amplitude of annual range (1/2 annual range) = **9.7°C**
Mean annual temperature = **2.52 °C**
Annual melt (mm w.e.) = **6504**



*Full data for '**SINE CURVE 3**' above and degree day factor of 4mm day^{-1} K^{-1}, following Braithwaite et al. 2006.*

	Day	Daily temperature mean	Degree-day factor	Daily melt
January 1st	1	-6.6	4	0.0
	2	-6.7	4	0.0
	3	-6.7	4	0.0
	4	-6.8	4	0.0
	5	-6.8	4	0.0
	6	-6.9	4	0.0
	7	-6.9	4	0.0
	8	-6.9	4	0.0
	9	-7.0	4	0.0
	10	-7.0	4	0.0
	11	-7.0	4	0.0
	12	-7.1	4	0.0
	13	-7.1	4	0.0
	14	-7.1	4	0.0
	15	-7.1	4	0.0
	16	-7.1	4	0.0
	17	-7.2	4	0.0
	18	-7.2	4	0.0
	19	-7.2	4	0.0
	20	-7.2	4	0.0
	21	-7.2	4	0.0
	22	-7.2	4	0.0
	23	-7.2	4	0.0
	24	-7.2	4	0.0
	25	-7.2	4	0.0
	26	-7.1	4	0.0
	27	-7.1	4	0.0
	28	-7.1	4	0.0
	29	-7.1	4	0.0
	30	-7.1	4	0.0

Day	Daily temperature mean	Degree-day factor	Daily melt
31	-7.0	4	0.0
32	-7.0	4	0.0
33	-7.0	4	0.0
34	-6.9	4	0.0
35	-6.9	4	0.0
36	-6.9	4	0.0
37	-6.8	4	0.0
38	-6.8	4	0.0
39	-6.7	4	0.0
40	-6.7	4	0.0
41	-6.6	4	0.0
42	-6.5	4	0.0
43	-6.5	4	0.0
44	-6.4	4	0.0
45	-6.4	4	0.0
46	-6.3	4	0.0
47	-6.2	4	0.0
48	-6.1	4	0.0
49	-6.1	4	0.0
50	-6.0	4	0.0
51	-5.9	4	0.0
52	-5.8	4	0.0
53	-5.7	4	0.0
54	-5.6	4	0.0
55	-5.6	4	0.0
56	-5.5	4	0.0
57	-5.4	4	0.0
58	-5.3	4	0.0
59	-5.2	4	0.0
60	-5.1	4	0.0

APPENDIX VII

Day	Daily temperature mean	Degree-day factor	Daily melt
61	-5.0	4	0.0
62	-4.9	4	0.0
63	-4.7	4	0.0
64	-4.6	4	0.0
65	-4.5	4	0.0
66	-4.4	4	0.0
67	-4.3	4	0.0
68	-4.2	4	0.0
69	-4.0	4	0.0
70	-3.9	4	0.0
71	-3.8	4	0.0
72	-3.7	4	0.0
73	-3.5	4	0.0
74	-3.4	4	0.0
75	-3.3	4	0.0
76	-3.1	4	0.0
77	-3.0	4	0.0
78	-2.9	4	0.0
79	-2.7	4	0.0
80	-2.6	4	0.0
81	-2.4	4	0.0
82	-2.3	4	0.0
83	-2.1	4	0.0
84	-2.0	4	0.0
85	-1.9	4	0.0
86	-1.7	4	0.0
87	-1.6	4	0.0
88	-1.4	4	0.0
89	-1.2	4	0.0
90	-1.1	4	0.0
91	-0.9	4	0.0
92	-0.8	4	0.0
93	-0.6	4	0.0
94	-0.5	4	0.0
95	-0.3	4	0.0
96	-0.1	4	0.0
97	0.0	4	0.1
98	0.2	4	0.7
99	0.3	4	1.4
100	0.5	4	2.0
101	0.7	4	2.7
102	0.8	4	3.3
103	1.0	4	4.0
104	1.2	4	4.6
105	1.3	4	5.3
106	1.5	4	6.0
107	1.7	4	6.6
108	1.8	4	7.3
109	2.0	4	8.0
110	2.2	4	8.6
111	2.3	4	9.3
112	2.5	4	10.0
113	2.7	4	10.6
114	2.8	4	11.3
115	3.0	4	12.0
116	3.2	4	12.6
117	3.3	4	13.3
118	3.5	4	14.0
119	3.7	4	14.6
120	3.8	4	15.3
121	4.0	4	15.9
122	4.2	4	16.6
123	4.3	4	17.3
124	4.5	4	17.9
125	4.6	4	18.6
126	4.8	4	19.2
127	5.0	4	19.9
128	5.1	4	20.5
129	5.3	4	21.2
130	5.4	4	21.8
131	5.6	4	22.4
132	5.8	4	23.1
133	5.9	4	23.7
134	6.1	4	24.3
135	6.2	4	24.9
136	6.4	4	25.5
137	6.5	4	26.2
138	6.7	4	26.8
139	6.8	4	27.4
140	7.0	4	28.0
141	7.1	4	28.5
142	7.3	4	29.1
143	7.4	4	29.7
144	7.6	4	30.3
145	7.7	4	30.8
146	7.9	4	31.4
147	8.0	4	32.0
148	8.1	4	32.5
149	8.3	4	33.1
150	8.4	4	33.6
151	8.5	4	34.1
152	8.7	4	34.6
153	8.8	4	35.1
154	8.9	4	35.7
155	9.0	4	36.2
156	9.2	4	36.6
157	9.3	4	37.1
158	9.4	4	37.6
159	9.5	4	38.1
160	9.6	4	38.5
161	9.7	4	39.0
162	9.9	4	39.4
163	10.0	4	39.8
164	10.1	4	40.3
165	10.2	4	40.7
166	10.3	4	41.1
167	10.4	4	41.5
168	10.5	4	41.9
169	10.6	4	42.3
170	10.7	4	42.6
171	10.7	4	43.0
172	10.8	4	43.3
173	10.9	4	43.7
174	11.0	4	44.0
175	11.1	4	44.3
176	11.2	4	44.6
177	11.2	4	44.9
178	11.3	4	45.2
179	11.4	4	45.5
180	11.4	4	45.8
181	11.5	4	46.0
182	11.6	4	46.3
183	11.6	4	46.5
184	11.7	4	46.7
185	11.7	4	46.9
186	11.8	4	47.1
187	11.8	4	47.3
188	11.9	4	47.5
189	11.9	4	47.7
190	12.0	4	47.8
191	12.0	4	48.0
192	12.0	4	48.1
193	12.1	4	48.3
194	12.1	4	48.4
195	12.1	4	48.5
196	12.1	4	48.6
197	12.2	4	48.6
198	12.2	4	48.7
199	12.2	4	48.8
200	12.2	4	48.8
201	12.2	4	48.8
202	12.2	4	48.9
203	12.2	4	48.9
204	12.2	4	48.9
205	12.2	4	48.9
206	12.2	4	48.8
207	12.2	4	48.8
208	12.2	4	48.8
209	12.2	4	48.7
210	12.2	4	48.6
211	12.1	4	48.6
212	12.1	4	48.5
213	12.1	4	48.4
214	12.1	4	48.2
215	12.0	4	48.1
216	12.0	4	48.0

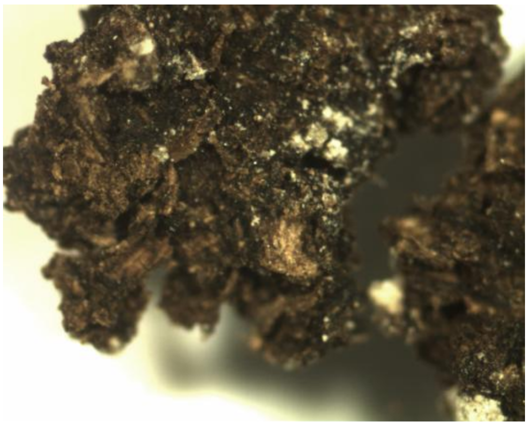
APPENDIX VII

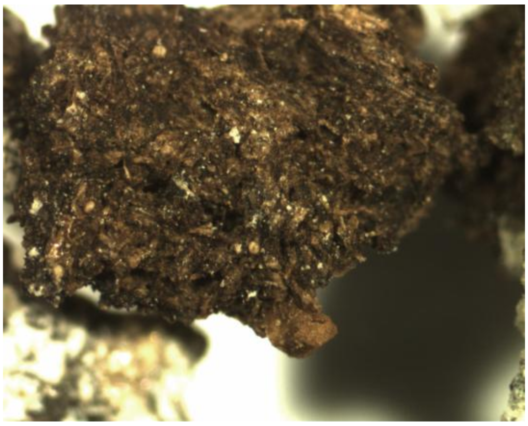
Day	Daily temperature mean	Degree-day factor	Daily melt
217	12.0	4	47.8
218	11.9	4	47.7
219	11.9	4	47.5
220	11.8	4	47.3
221	11.8	4	47.1
222	11.7	4	46.9
223	11.7	4	46.7
224	11.6	4	46.5
225	11.6	4	46.2
226	11.5	4	46.0
227	11.4	4	45.7
228	11.4	4	45.5
229	11.3	4	45.2
230	11.2	4	44.9
231	11.2	4	44.6
232	11.1	4	44.3
233	11.0	4	44.0
234	10.9	4	43.6
235	10.8	4	43.3
236	10.7	4	43.0
237	10.7	4	42.6
238	10.6	4	42.2
239	10.5	4	41.9
240	10.4	4	41.5
241	10.3	4	41.1
242	10.2	4	40.7
243	10.1	4	40.2
244	10.0	4	39.8
245	9.8	4	39.4
246	9.7	4	38.9
247	9.6	4	38.5
248	9.5	4	38.0
249	9.4	4	37.6
250	9.3	4	37.1
251	9.2	4	36.6
252	9.0	4	36.1
253	8.9	4	35.6
254	8.8	4	35.1
255	8.7	4	34.6
256	8.5	4	34.1
257	8.4	4	33.6
258	8.3	4	33.0
259	8.1	4	32.5
260	8.0	4	31.9
261	7.8	4	31.4
262	7.7	4	30.8
263	7.6	4	30.2
264	7.4	4	29.7
265	7.3	4	29.1
266	7.1	4	28.5
267	7.0	4	27.9
268	6.8	4	27.3
269	6.7	4	26.7
270	6.5	4	26.1
271	6.4	4	25.5
272	6.2	4	24.9
273	6.1	4	24.3
274	5.9	4	23.6
275	5.8	4	23.0
276	5.6	4	22.4
277	5.4	4	21.8
278	5.3	4	21.1
279	5.1	4	20.5
280	5.0	4	19.8
281	4.8	4	19.2
282	4.6	4	18.5
283	4.5	4	17.9
284	4.3	4	17.2
285	4.1	4	16.6
286	4.0	4	15.9
287	3.8	4	15.2
288	3.6	4	14.6
289	3.5	4	13.9
290	3.3	4	13.3
291	3.1	4	12.6
292	3.0	4	11.9
293	2.8	4	11.3
294	2.6	4	10.6

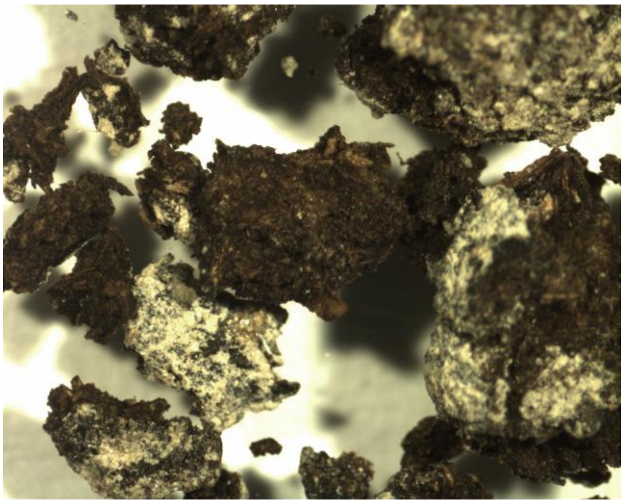
	Day	Daily temperature mean	Degree-day factor	Daily melt
	295	2.5	4	9.9
	296	2.3	4	9.3
	297	2.1	4	8.6
	298	2.0	4	7.9
	299	1.8	4	7.3
	300	1.6	4	6.6
	301	1.5	4	5.9
	302	1.3	4	5.3
	303	1.2	4	4.6
	304	1.0	4	3.9
	305	0.8	4	3.3
	306	0.7	4	2.6
	307	0.5	4	2.0
	308	0.3	4	1.3
	309	0.2	4	0.7
	310	0.0	4	0.0
	311	-0.2	4	0.0
	312	-0.3	4	0.0
	313	-0.5	4	0.0
	314	-0.6	4	0.0
	315	-0.8	4	0.0
	316	-0.9	4	0.0
	317	-1.1	4	0.0
	318	-1.3	4	0.0
	319	-1.4	4	0.0
	320	-1.6	4	0.0
	321	-1.7	4	0.0
	322	-1.9	4	0.0
	323	-2.0	4	0.0
	324	-2.2	4	0.0
	325	-2.3	4	0.0
	326	-2.4	4	0.0
	327	-2.6	4	0.0
	328	-2.7	4	0.0
	329	-2.9	4	0.0
	330	-3.0	4	0.0
	331	-3.1	4	0.0
	332	-3.3	4	0.0
	333	-3.4	4	0.0
	334	-3.5	4	0.0
	335	-3.7	4	0.0
	336	-3.8	4	0.0
	337	-3.9	4	0.0
	338	-4.0	4	0.0
	339	-4.2	4	0.0
	340	-4.3	4	0.0
	341	-4.4	4	0.0
	342	-4.5	4	0.0
	343	-4.6	4	0.0
	344	-4.7	4	0.0
	345	-4.9	4	0.0
	346	-5.0	4	0.0
	347	-5.1	4	0.0
	348	-5.2	4	0.0
	349	-5.3	4	0.0
	350	-5.4	4	0.0
	351	-5.5	4	0.0
	352	-5.6	4	0.0
	353	-5.7	4	0.0
	354	-5.7	4	0.0
	355	-5.8	4	0.0
	356	-5.9	4	0.0
	357	-6.0	4	0.0
	358	-6.1	4	0.0
	359	-6.1	4	0.0
	360	-6.2	4	0.0
	361	-6.3	4	0.0
	362	-6.4	4	0.0
	363	-6.4	4	0.0
	364	-6.5	4	0.0
December 31st	365	-6.6	4	0.0

APPENDIX VIII

Terrestrial plant macrofossil from core BJC1, Lake C at Buni i Jezerces in Montenegro/Albania.







APPENDIX IX

Stratigraphic survey results for cores B1 – B6. Results are taken from field descriptions along the Lake Plav transect.

Boundary recorded as: VS – very sharp, over < 0.5mm; S – sharp, between 0.5 – 1 mm; G – gradual, between 1 – 2 mm; VG - very gradual, between 2 mm – 1 cm; D – diffuse boundary, >1 cm (after Birks and Birks, 1980) against the stratigraphic unit below.

Colour follows Munsell colour following Munsell soil chart (2000) and Sutherland *et al.* (2000).

Lake Plav, Core 1 (B1) stratigraphy:

Depth (cm)	Description and composition	Boundary	Colour
0 – 39	Greyish brown clay silt with intertwined rootlets, high organic content $Ag^2 Th^1 As^1$	VG	
39 - 67	Dark greyish brown to olive clay silt with intertwined rootlets and fine sand $As^2Ag^2 Ga^+ Th^+$		
67 – 100	Dark grey silty clay loam and fine sand $As^3Ag^1 Ga^+$	S	
100 – 225	Possibly one unit with laminations as follows (grey highlighted)		
100 – 107	Dark grey clay loam with silt and fine sand $As^3Ag^1 Ga^+$		
107 – 107.5	Dark grey silty clay with soily remains $As^3Ag^1 Sh^+$		
107.5 – 112	Dark grey fine to medium sand with silt $Ga^3 Ag^1$		
112 – 116.5	Brownish grey clay silt with some fine sand $As^2 Ag^2 Th^+ Ga^+$		
116.5 – 117.5	Brownish grey clay silt with some fine sand $As^2 Ag^2 Th^+ Ga^+$		
117.5 – 120.5	Brownish grey fine sand with clay silt $Ga^2 As^1 Ag^1$		
120.5 – 124	Dark grey silt and fine to medium sand Ag^2Ga^2		
124 – 124.5	Brownish grey fine sand with clay silt $Ga^2 As^1 Ag^1$		
124.5 – 140.5	Dark grey fine to medium sand with silt $Ga^3 Ag^1$		
140 – 142	Dark grey silty clay As^3Ag^1		
142 – 152	Brownish grey fine sand with clay silt $Ga^2 As^1 Ag^1$	S	
152 – 156.5	White/cream mottled clay silt with sand		

	$Ag^3\ As^1$		
156.5 – 157.5	Brownish grey fine sand with clay silt $Ga^2\ As^1\ Ag^1$		
157.5 – 162	Brownish grey sandy loam $Ga^2\ As^1\ Ag^1$		
162 – 163.5	As above, but with organic content $Ga^2\ Sh^1\ Ag^1$		
163.5 – 168	Brownish grey fine sand with clay silt $Ga^2\ As^1\ Ag^1$		
168 – 169	Dark grey clay, silt and sand mix $Ag^2\ Ga^1\ As^1$		
169 – 173	Brownish grey fine sand with clay silt $Ga^2\ As^1\ Ag^1$		
173 – 183	Dark grey medium sand with clay $Ga^3\ As^1$		
183 – 184.5	Dark grey silty clay with fine sand $As^3\ Ag^1\ Ga^+$		
184.5 – 186	Fine sand with silt and woody fragments $Ga^3\ Tl^1\ Ag^+$		
186 – 188.5	Fine sand with silt $Ga^3\ Ag^+$		
188.5 – 207	Fine sand with clay silt $Ga^2\ As^1\ Ag^1$		
207 – 210	As above with soily remains $Ga^2\ As^1\ Ag^1\ Sh^+$		
210 – 223	Fine sand with clay silt $Ga^2\ As^1\ Ag^1$		
223 – 225	As above with soily remains $Ga^2\ As^1\ Ag^1\ Sh^+$		
225 – 238	Medium fine sand Ga^4	S/G	
238 – 250	Medium sand including coarse sand grains $Ga^3\ Gs^1$		
250 – 350	Dark grey medium sand Ga^4		

Lake Plav, Core 2 (B2) stratigraphy:

Depth (cm)	Description and composition	Boundary	Colour
0 – 21.5	Dark grey clay silt with stems, roots and orange mottling $Ag^2\ As^1\ Th^1$	G	
21.5 - 31	Brownish grey with orange mottling, clay silt $Ag^3\ As^1$	G	
31 – 60.5	Dark grey silty clay, less mottled than previous unit, with fine rootlets and increasing siccitas with depth $As^2\ Ag^1\ Th^1$		
60.5 - 64	Dark grey clay:silt with small wood fragments	S	

	As^2 Ag^2 Dl^+		
64 – 109	Dark grey clay:silt with horizontally bedded Phragmites As^2 Ag^2 Th^+		
109 – 117	Grey clay:silt with less organic fragments than above unit As^2 Ag^2 Sh^+	VG	
117 – 126.5	Dark grey clay:silt with dark rootlets As^2 Ag^2 Th^+	D	
126.5 – 128.5	Grey clay:silt with less organic fragments As^2 Ag^2 Sh^+	S	
128.5 – 132	Dark grey clay:silt with horizontally bedded Phragmites As^2 Ag^2 Th^+	S	
132 – 150	Grey clay:silt with some organic fragments As^2 Ag^2 Sh^+		
150 – 187.5	Grey clay:silt with white mottling at the top of the unit, becoming darker to the base As^2 Ag^2	D	
187.5 – 200	Dark grey clay:silt As^2 Ag^2	G	
200 – 232	Coarse sand including white and pale yellow sand fragments Gs^4		
232 – 234.5	Transition to medium/fine sand Ga^4		
234.5 – 350	Medium sand with wood sample taken at 237 – 238cm Ga^4 Dl^+		
350 – 378	Coarse sand Gs^4		
378 – 383	5mm dark grit unit Gg^4		
383 – 622	Coarse sand with wood sampled at 545cm Gs^4 Dl^+		

Lake Plav, Core 3 (B3) stratigraphy:

Depth (cm)	Description and composition	Boundary	Colour
0 – 40	Brown grey clay silt with intertwined roots, some woody Ag^3 As^1 Th^+ Tl^+	VG	
40 – 63	Brown grey clay silt with organic horizons (~2mm) 50 – 63cm Ag^3 As^1 Sh^+		
63 – 67	Grey silty loam Ag^2 As^1 Ga^1		

67 – 77	Medium sand with some finer fragments and silt Gs4 Ag^+		
77 – 80	Grey silty loam Ag^2 As^1 Ga^1	S	
80 – 95	Medium sand with some finer fragments and silt Ga^4 Ag^+		
95 – 100	Dark grey coarse sand with silt Gs^4 Ag^+		
100 – 124	Medium sand with silt Ag^+ Ga^4 and silty loam organic horizons at: 100 – 102cm Sh^4 As^+ Ag^+ Ga^+ 109.5–111cm, 115.5–119cm, 123.5–124cm $Ag^2As^1Ga^1Sh^+$		
124 – 145	Silty loam with clay Ag^2 As^1 Ga^1 Sh^+ With organic remains (Dh) at 139cm		
145 – 149	Grey silty clay As^4 Ag^+		
149 – 154	Olive grey silty clay As^3 Ag^1		
154 – 204	Clay:silt with occasional sand layers, organic remains and staining at 182 – 204cm As^2 Ag^2 Ga^+ Th^+ Sh^+		
204 – 213	Clay silt Ag^3 As^1	D	
213 – 224	Clay:silt with some organics As^2 Ag^2 Sh^+		
224 – 232	Clay silt Ag^3 As^1		
232 – 257	Clay:silt with some organics As^2 Ag^2 Sh^+		
257 – 263	Brown grey clay:silt As^2 Ag^2		
263 – 283	Silty clay with possible molluscs/ostracoda layers As^3 Ag^1 Ld^+		
283 – 285	Coarse sand Gs^4		
285 – 349.5	Clay with silt As^4 Ag^+		
349.5 – 354.5	Silty clay As^3 Ag^1		
354.5 – 366	Clay with silt As^4 Ag^+		
366 – 404	Dark grey silty clay highly stratified with light grey bands, becoming less stratified with depth As^3 Ag^1	D	
404 – 432	Clay with silt and mottled appearance As^4 Ag^+		
432 – 438	Clay with silt		

	As^4 Ag^+		
438 – 442	Clay As^4		
442 – 462	Dark grey silty clay highly stratified with light grey bands, v.dark bands at 443.5 – 444cm and 460 – 460.5cm As^3 Ag^1		
462 – 490	Clay with silt and occasional v.dark grey horizons (2-3mm) As^4 Ag^+		
490 – 507	Dark grey silty clay highly stratified with light grey bands, becoming less stratified with depth As^3 Ag^1		
507 – 536.5	Clay with silt and occasional dark grey horizons As^4 Ag^+		
536.5 – 551	Dark grey silty clay highly stratified As^3 Ag^1		
551 – 591	Clay with silt As^4 Ag^+		
591 – 600.5	Grey silty clay with sand As^3 Ag^1 Ga^+		
600.5 – 603	Light grey clay with silt As^4 Ag^+		
603 – 610.5	Grey clay with silt and dark grey horizons (2-3mm) As^4 Ag^+		
610.5 – 611	Darker grey silty clay loam with sand As^2 Ag^2 Ga^+		
611 – 613	Grey clay with silt and dark grey horizons (2-3mm) As^4 Ag^+		
613 – 615	Dark grey sand with silt Ga^4 Ag^+		
615 – 632	Grey clay with silt and dark grey horizons (2-3mm) As^4 Ag^+		

Lake Plav, Core 4 (B4) stratigraphy:

Depth (cm)	Description and composition	Boundary	Colour
0 – 23	Dry dark olive brown clay silt with high organic content, including *Equisetum*, rootlets and oxidised mottles Th^2 Ag^1 As^+	VG	
23 – 50	Dark grey clay:silt, organic remains vertically bedded *Equisetum* nodule at 33-34cm As^2 Ag^2		
50 – 70	Dark grey clay:silt with orange mottling As^2 Ag^2		
70 – 76.5	Dark grey clay:silt As^2 Ag^2	VG	

76.5 – 150	Silty clay with horizontally bedded As^3 Ag^1 organic rich layers brown grey layers at 105-109cm, 144-146cm (As^2 Ag^1 Th^1)		
150 – 152	Greater organic content, remains silty clay As^2 Th^1 Ag^+ Sh^+		
152 – 154	Light grey clay:silt As^2 Ag^2		
154 – 174	Grey clay:silt with organics As^2 Ag^2 Sh^+	S	
174 – 176	Black brown detrital peat Sh^4	S	
176 – 177	Grey clay:silt with organics As^2 Ag^2 Sh^+	S	
177 – 178	Black brown detrital peat Sh^4	S	
178 – 180	Dark grey clay with high organic content As^2 Sh^2 Ag^+		
180 – 181	Black brown detrital peat Sh^4	S	
181 – 200	Dark grey organic rich clay, with silt As^2 Sh^2 Ag^+		
200 – 203	V. dark grey silty clay with high organic content As^2 Ag^1 Sh^1	VS	
203 – 210.5	Grey clay silt Ag^3 As^1	VS	
210.5 – 257.5	Olive grey clay:silt with occasional horizontally bedded rootlets As^2 Ag^2 Th^+		
257.5 – 298	Olive brown-grey slightly mottled fine sand silt, becoming more grey with depth Ag^3 Ga^1	S	
298 – 316	Dark grey silty clay with molluscs As^3 Ag^1 Ld^+ (moll.)		
316 – 462	Silty clay with sand and molluscs As^3 Ag^1 Ga^+ Ld^+ (moll.)	D	
462 – 480	Clay silt Ag^3 As^1		
480 - 505	Clay silt with molluscs Ag^3 As^1 $Ld(moll.)^+$	S	
505 – 508	Stratified light grey clay As^4	S	
508 – 539	Clay silt Ag^3 As^1		
539 – 541	Stratified light grey clay As^4		
541 – 568.5	Laminated silty clay with fine sand and molluscs 565 – 566cm As^3 Ag^1 Ga^+ Ld^+(moll.)		
568.5 – 570	Mollusc rich dark grey silty clay As^2 Ag^1 $Ld(moll.)^1$		

570 – 571	Stratified light grey clay As^4	S	
571 – 632.5	Grey silty clay with fine sand and molluscs As^2 Ag^1 Ga^1 Ld^+(moll.) with small <1cm layers as follows (grey highlighted)		
575 – 576	Light grey clay As^4		
578 – 578.5	Dark grey clay with fine sand As^3 Ga^1		
582 – 583	Layer as 578 – 578.5cm		
585 – 586	Layer as 578 – 578.5cm		
591.5 – 591.8	Layer as 578 – 578.5cm		
592.4 – 592.7	Layer as 578 – 578.5cm		
596 – 597	Layer as 578 – 578.5cm		
632.5 – 633.5	Grey clay silt As^1 Ag^3		
633.5 – 634	Light grey clay silt with sand Ag^3 As^1 Ga^+	S	
634 – 639	Olivey grey stratified clay silt Ag^3 As^1	VG	
639 – 648.5	Mottled clay silt, mottles light grey to olivey grey Ag^3As^1		
648.5 – 658.5	Dark grey clay silt with sand, coarser texture Ag^2 As^1 Ga^1		
658.5 – 716	Clay silt with some molluscs. Unit becomes more stratified with depth, strata 4 – 8mm Ag^3As^1 Ld(moll.)$^+$	D	

Lake Plav, Core 5 (B5) stratigraphy:

Depth (cm)	Description and composition	Boundary	Colour
0 – 3	Clay silt with high organic content Th^2 As^1 Ag^1		
3 – 26	Silty clay with grey/brown mottling (2.5Y 4/1) As^2 Ag^1 Th^1		
26 – 38.5	HIATUS		
38.5 – 73	Stratified silty clay with horizontally bedded organics As^3 Ag^1 Th^+	VG	
73 – 92	Grey/brown mottled silty clay As^2 Ag^1 Th^1	VS	
92 – 101.5	Silty clay with organics horizontally bedded/stratified As^3 Ag^1 Th^+	G	
101.5 – 133	Some roots found within the silty clay As^2 Ag^1 Th^1		
133 – 139	Slightly stratified silty peat, with some sand Sh^3 Ag^1 Ga^+		2.5Y 2.5/1

139 – 144.5	Silty clay with less organics, but those there are horizontally bedded $As^3 Ag^1 Th^+$		2.5Y 4/1
144.5 – 149	Slightly stratified silty peat, with some sand $Sh^3 Ag^1 Ga^+$		2.5Y 4/1
149 – 163	HIATUS		2.5Y 4/1
168 – 171	Clay silt some sand and matted organics $As^2 Ag^1 Th^1$		2.5Y 4/1
171 – 177	Clay rich layer with some organics $As^3 Ag^1 Sh^+$		2.5Y 4/1
177 – 179	Peaty silt $Ag^2 Sh^2$		2.5Y 4/1
179 – 195	Silty clay some horizontally bedded roots $As^3 Ag^1 Th^+$		2.5Y 4/1
195 – 199	Clay rich layer with no stratification $As^4 Ag^+$	VG	2.5Y 4/1
199 – 217	Darker clay rich layer $As^4 Ag^+$		2.5Y 3/1
217 – 225	Silty clay with some sand and organic fragments $As^3 Ag^1 Th^+ Ga^+$		2.5Y 3/1 to 0Y 3/1
225 – 239	Silty clay with sand and organics $As^3 Ag^1 Th^+ Ga^+$		2.4Y 4/1
239 – 339	Silty clay with sand and organics $As^3 Ag^1 Th^+ Ga^+$ Laminations within this section follow in highlighted in grey		
253 – 253.3	Very small charcoal fragments		
293 – 294.5	Some organic developments $As^3 Ag^1 Th^+ Ga^+ Sh^+$		
309.5 – 311.5	Some organic developments $As^3 Ag^1 Th^+ Ga^+ Sh^+$		
311.5 – 339	Matted roots in silty clay sediments $As^2 Ag^1 Th^1$		
339 – 439	Silty clay with fine sand and reducing organics with depth $As^2 Ag^1 Ga^+ Th^+$ Molluscs visible 345 – 365cm $As^2 Ag^1 Ld^+$ (moll.)		2.5Y 4/1
439 – 639	Silty clay with fine sand, occasional organics and molluscs $As^3 Ag^1 Ga^+ Th^+ Ld^+$ (moll.)		2.5Y 5/1
639 – 674	Dark greenish gray silty clay and continued fine sand $As^3 Ag^1 Ga^+$		Gley1 4/10Y

APPENDIX IX

Lake Plav, Core 6 (B6) stratigraphy:

Depth (cm)	Description and composition	Boundary	Colour
0 – 17.5	Very dark greyish brown with many rootlets, silty peat $Sh^2 Ag^1 Th^1$	D	2.5Y 3/2
17.5 – 33	Some rootlets in silty clay $As^3 Ag^1$	D	2.5Y 4/1
33 – 95	Matted fibrous detrital peat with silt, sand and large fragments of Phragmites $Sh^2 Ag^1 Th^1 Ga^+$		2.5Y 3/1
95 - 169	Nothing retrieved, but base at 169cm		